Preliminary Appraisal of a Correlation Between Glaciations and Large Igneous Provinces Over the Past 720 Million Years

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ABSTRACT

Earth has gone through periods of cooling including global, near global, or regional glaciations, which are observed in the Archean, Paleoproterozoic, Neoproterozoic, Ordovician, Permo-Carboniferous, and Cenozoic times. We review the mechanisms by which large igneous provinces (LIPs) and silicic LIPs (SLIPs) can cause global cooling. Then we investigate the correlation of LIPs with important glaciation events, focusing on those of Neoproterozoic and Phanerzoic age. The 720 Ma Franklin-Irkutsk LIP, a large part of which was emplaced into an evaporite basin and all of which was emplaced in the tropics, is linked with the start of the Sturtian glaciation, one of the longest and most severe glaciations in Earth history. The ca. 579 Ma pulse of the Central Iapetus Magmatic Province (CIMP) is associated with the start and end of the Gaskiers glaciation. The Hirnantian glaciation (ca. 440 Ma) may be associated with poorly dated ca. 450 Ma intraplate magmatism in several regions, including eastern Siberia, South Korea, Argentina, Iran, and elsewhere. It is also coincident with a huge volume of silicic volcanic provinces generated by supereruptions. Permo-Carboniferous glaciations (P1–P4, 300–260 Ma) can be correlated with widespread intraplate magmatism of the European northwest African magmatic province (and its initiation as the 300 Ma Skagerrak LIP), and also the 259 Ma Emeishan LIP of China. A recently recognized ca. 34 Ma initial pulse of the Afro-Arabian LIP matches the Eocene-Oligocene cooling (Oi-1 glaciation). More precise dating of both the LIPs and cooling events is required to confirm the correlations and to assess the role of LIPs relative to other causes proposed for global and regional glaciations.

8.1. INTRODUCTION

Note the default cause for the end of glaciations in rebuilding of CO2 levels in the atmosphere due to steady state volcanism until a critical point is reached. We note cases where a specific LIP event may have had a cause in the end of the glaciation (Table 8.1).
| Glaciation | Start (LIP/ SLIP) | Finish (LIP) | Correlation quality (Robust/Possible/ No known link) |
|------------|------------------|--------------|-----------------------------------------------------|
| ca. > 30 Ma | Afro-Arabian precursor (starting at 34 Ma) | | Possible |
| Eocene-Oligocene (34 Ma) | | | |
| Cretaceous | Caribbean-Colombian (90 Ma), Madagascar (89 Ma), & second pulse of HALIP (95 Ma) | Deccan (66 Ma) | Possible (start) Robust (finish) |
| Late Turonian through Maastrichtian (90–66 Ma) | | | |
| Late Valanginian– earliest Hauterivian cooling (133.9 Ma) | Parana-Etneka (134 Ma), “Trap” event of SW Greenland (ca. 130 Ma) | | Possible |
| Jurassic | Pre-Karoo LIP | Karoo-Ferrar (183 Ma) | Possible |
| Late Pliensbachian cooling Aalenian (174.1 Ma to about 170.3 Ma) through Bajocian (170.3 Ma to around 168.3) | Late pulse of Karoo (based on Ar-Ar dating) | | Possible |
| Late Callovian (ending at 166.1 ± 4.0 Ma) does not have any known LIP link. | No LIP identified | No known link | |
| Permo-Carboniferous 273–260 Ma P3-P4 | Multipulsed EUNWA (327–250 Ma), Emeishan (260 Ma), Tarim (280 Ma), Panjal-Qiangtang (285 Ma), & Choiyoi (265 Ma) | Multipulsed EUNWA (327–250 Ma) & Emeishan (259 Ma) | Possible (start & finish) |
| 287–280 Ma P2 | Multipulsed EUNWA (327–250 Ma), Tarim (280), & Panjal-Qiangtang (285 Ma) | Multipulsed EUNWA (327–250 Ma) & Tarim (280 Ma)? | Possible (start & finish) |
| 300–290 Ma P1 | Multipulsed EUNWA (327–250 Ma; e.g., Skagerak, 300 Ma) | Multipulsed EUNWA (327–250 Ma) & Tarim (280 Ma)? | Possible (start & finish) |
| ca. 330–290 Ma LPIA | No LIP identified | Tarim (280 Ma)? | No known link (start) Possible (finish) |
| End-Ordovician 440 Ma Hirnantian | Suordakh (ca. 450 Ma) | | Robust (e.g., direct Hg evidence)/Possible (e.g., no LIP robustly identified) |
| Cambrian | Kkarindji (ca. 511 Ma) | | Possible |
| Possibly cooling during the Middle to Late Cambrian, between ca. 513 and ca. 488 Ma | | | |
| Lower Cambrian (ca. 532–530 Ma) | Wichita (540–530 Ma) with the main pulse at 532–530 | | Possible |
| Neoproterozoic ca. 579 Ma Gaskiers | Ouazarzate felsic superruptions (ca. 580 Ma) and other CIMP LIPs | Ouazarzate mafic continental flood basalts (ca. 580 Ma) and other CIMP LIPs | Robust (start & finish) |
| 640–635 Ma Marinoan | ? Wudang dyke swarm (ca. 650–630 Ma) | ? Wudang dyke swarm (ca. 650–630 Ma) | Possible (start? finish?) |
| 720–660 Ma Sturtian | Franklin-Irkutsk & Mutare-Fingeren (723–715 Ma) | | Robust |
8.1.1. LIPs and Their Climatic Effect

Large igneous provinces (LIPs) represent large volume (>0.1 Mkm³; frequently above ~1 Mkm³), mainly mafic (ultramafic) magmatic events of intraplate affinity, which occur in both continental and oceanic settings, and are typically of short duration (<5 Myr) or consist of multiple short pulses over a maximum of a few tens of Myr (Coffin & Eldholm, 1994, 2005; Bryan & Ernst, 2008; Bryan & Ferrari, 2013; Ernst, 2014; Ernst & Youbi, 2017, Ernst et al., Chapter 1 this volume; and references therein). They comprise volcanic packages (flood basalts), and a plumbing system of mafic dyke swarms, sill complexes, mafic-ultramafic layered intrusions, and a lower crustal magmatic underplate. LIPs can also be associated with silicic magmatism (including dominantly silicic events termed silicic LIPs, or SLIPs, sometimes including the so-called supereruptions/supervolcanoes), and also carbonatites and kimberlites.

LIPs and SLIPs can have significant global climatic effects including causing mass extinction events, via a complex web of changes (characteristically rapid) in atmospheric/oceanic acidification, oceanic anoxia, sea level, toxic metal input (e.g., Hg), and, most relevant to this appraisal, both warming and cooling (e.g., Rampino et al., 1988; Stothers, 1993; Courtillot et al., 1996; Wignall, 2001, 2005; Courtillot & Renne, 2003; Kelley, 2007; Bond & Wignall, 2014; Rampino & Self, 2015; Burgess & Bowring, 2015; Ernst & Youbi, 2017; Bond & Grasby, 2017; Rampino & Caldeira, 2017). In the broadest sense, LIPs can affect (or even induce) shifts between icehouse, greenhouse and supergreenhouse climatic states (e.g., Kidder & Worsley, 2010, 2012). An important concept is that while LIP events are large (often millions of square kilometers in horizontal extent), they are not global in extent. But the important insight is that they can have a global effect on the environment both atmospheric and oceanic (e.g., Ernst & Youbi, 2017).

We first review the range of mechanisms by which LIPs and SLIPs can cause global cooling, and then consider the evidence for specific global cooling events to be linked with LIPs. We also consider which LIP events could be responsible for the termination of cooling events.

Table 8.1 provides a summary of Neoproterozoic and Phanerozoic glaciations and summarizes inferred origins and speculates on potential links with LIPs in each case. In most cases, the links we propose are plausible but not proven. High precision geochronology (down to uncertain levels to <50,000 yr) is required for fully testing the relationships as has been achieved for LIPs such as the 252 Ma Siberian Traps, the 201 Ma Central Atlantic Magmatic Province (CAMP), and the 66 Ma Deccan Traps, which now have compelling links with end Permian, end Triassic and end Cretaceous mass extinctions, respectively (e.g., Burgess et al., 2015b; Davies et al., 2017; Schoene et al., 2015; 2019; Kasbohm et al., Chapter 2 this volume).

In this contribution, we test the hypothesis that glaciations correlate with LIPs; or stated conversely, we test the null hypothesis that there is no correlation between climatic cooling and LIPs. Since we find that a majority of climatic cooling events correlate either robustly or possibly with coeval LIPs, we cannot reject the hypothesis that LIP emplacement promotes planetary cooling; or stated conversely, we can at least reject the hypothesis that there is no correlation between glaciations and LIPs.

We readily admit that there are other viable approaches to testing this hypothesis. For example, Park et al. (Chapter 7 this volume) test the same hypothesis, but choose to incorporate additional theoretical considerations. Surely many theoretical constraints, such as the paleolatitude of LIPs, are likely important given what we know about the climatic importance of LIPs emplaced during most recent Cretaceous and Cenozoic times (Kent & Muttoni, 2008; 2013). However, since even further complications such as true polar wander, e.g., the ~30° Mesozoic “monster shift” (Kent et al., 2015; Fu & Kent, 2018; Fu et al., 2020), are not accounted for in the paleolatitude estimates of Park et al. (Chapter 7 this volume), we argue that additional considerations such as paleolatitude are potentially ambiguous.

### Table 8.1 (Continued)

| Glaciation       | Start (LIP/SLIP)                  | Finish (LIP)                  | Correlation quality |
|------------------|-----------------------------------|--------------------------------|---------------------|
| Paleoproterozoic | ? Hekpoort (ca. 2.25 Ma)          | ? Ungava-Nipissing (2.22–2.21 Ga) | Possible            |
| ca. 2.33 Ga Gowganda-Rooihoogte | ? Kuito-Taivalkovski (2.33–2.31 Ga) | ? Kuito-Taivalkovski (2.33–2.31 Ga) | Possible            |
| 2.37 Ga Bruce-Duitschland | ? Bangalore (2.37 Ga) | ? Bangalore (2.37 Ga) | Possible            |
| ca. 2.43 Ga Ramsey | ? Ongeluk (2.43 Ga)                  | ? Ongeluk (2.43 Ga)                  | Robust              |

Note: See referencing in the text, apart from Paleoproterozoic portion, which is after Gumsley et al. (2017) and Ernst and Youbi (2017).
at best. There are also concerns about the concept of a “LIP half-life” that are complicated to execute, particularly the choice to exclude LIPs that might have been initially buried, for example, even buried LIPs can later be weathered if erosion is deep enough (Mitchell et al., 2019). Again, incorporating additional theoretical constraints such as paleolatitude and half-life is surely ideal, but quite complicated in practice. Our own approach is therefore more focused on the empirical testing of the putative relationship, and we do so by offering side-by-side timelines of climate events and LIPs over the past 720 Myr.

### 8.1.2. Mechanisms for Global Cooling via LIPs

Earth goes through periods of global cooling (Table 8.1) that can include global, near global, or regional glaciations, which are observed in the Archean, Paleoproterozoic, Neoproterozoic, Cambrian, Ordovician, Permo-Carboniferous, Eocene–Oligocene, Eocene to middle Miocene, and Quaternary times (e.g., Evans et al., 1997; Augustin et al., 2004; Eyles, 2008; Hoffmann, 2009; Cather et al., 2009; Bradley, 2011; Prave et al., 2016b; Gunsley et al., 2017; Ernst & Youbi, 2017). There is an extensive literature on global and regional glaciations and a variety of causes have been considered (e.g., Raymo, 1991; Berner, 2004, and references therein). For instance, glaciations have been linked to silicate weathering (and CO2 drawdown) during major orogenic episodes such as the formation of the Himalayas (Cenozoic glaciation) and the assembly of Pangea (Permo-Carboniferous glaciation). Major land plant innovations (and their ability to extract CO2 and release oxygen) have also been thought to be a significant factor in causing or at least favoring glaciations, for example, for the Ordovician and Permo-Carboniferous glaciations (Lenton et al., 2012; Kidder & Worsley, 2010; Algeo et al., 2016; Boyce & Lee, 2017). In addition, it is also now recognized that LIPs can contribute to global cooling via at least three different mechanisms: (1) LIP input of SO2 into the atmosphere (and conversion to sulfate aerosols), (2) weathering of LIP units and CO2 drawdown, and (3) increased oceanic biologic productivity and resulting increased CO2 drawdown (SLIPs are particularly important in the latter).

**Volcanic Winter**

SO2 is a greenhouse gas and causes warming for days to weeks. But on a longer term, it causes cooling because it forms sunlight blocking sulfate aerosols (e.g., Bond & Wignall, 2014). The sulfur content of a LIP can be relevant to the amount of SO2 released by a LIP. Callegaro et al. (2014) showed that the CAMP and Deccan LIPs, both strongly linked to extinction events, have basalts with high sulfur content (up to 1.900 ppm) while the less damaging (not associated with mass extinction) Paraná-Etendeka LIP has basalts with lower sulfur content (less than 800 ppm). The effects are more dramatic if the gases are injected into the stratosphere via Plinian eruptions. So called supereruptions (Self & Blake, 2008), can cause cooler climate (Rampino et al., 1988; Robock, 2004; Self, 2006; Self & Blake, 2008; Stern et al., 2008). The 720 Ma Franklin LIP (and its extension into formerly attached southern Siberia after Ernst et al., 2016), a large part of which was emplaced into an evaporite basin, correlates closely with the start of the Sturtian glaciation (Macdonald et al., 2010), one of the longest and most severe glaciations in Earth history. Injection of sulfate aerosols into the stratosphere has recently been proposed as the mechanistic link between Franklin and the Sturtian snowball glaciation (Macdonald & Wordsworth, 2017). The Franklin LIP event was furthermore emplaced in the tropics (Denyszyn et al., 2009), rendering it additionally climatically important (next section).

**Weathering and CO2 Drawdown**

Another mechanism for driving cooling on the Earth’s surface is related to silicate weathering. Weathering of continental silicates is enhanced under a warmer and (assumed) wetter climate (at lower latitudes), and basaltic volcanic rocks weather about ~5 to 10 times faster than felsic compositions (Goddéris et al., 2003, 2014; Dessert et al., 2003; White & Brantley, 1995). Therefore, CO2 drawdown due to weathering of flood basalts can lead to global cool down and even glaciations (Cox et al., 2016). LIPs emplaced in the tropical weathering belt may have particularly significant climatic effects, as has been demonstrated for recent Cenozoic and Cretaceous times (Kent & Muttoni, 2008, 2013).

The sedimentary record should reflect certain trace element and isotopic compositions if LIPs are being massively weathered in contrast to average continental crust. For example, erosion of LIPs (and particularly their flood basalts) should produce geochemical/isotopic characteristics (radiogenic Nd and unradiogenic Sr and Os), which are typical of a dominantly mafic provenance (e.g., Mills et al., 2014; Cox et al., 2016).

**CO2 Drawdown Due to Increased Oceanic Productivity**

A related mechanism is increased CO2 drawdown due to increased ocean productivity (e.g., Cox et al., 2016). Oceanic productivity can be enhanced by increasing fertilization of the oceans. For instance, due to the apatite content in basalt, its weathering can effectively fertilize the oceans (Horton, 2015; Cox et al., 2018), with P as a key nutrient on geological timescales (Tyrrell, 1999). Iron fertilization is an effective climatic forcing mechanism and can be caused by great volumes of silicic volcanic ash (Cather et al., 2009). Cather et al. (2009) further conclude that most Phanerozoic cool-climate episodes were coeval with major explosive volcanism in silicic LIPs, suggesting a common link between these phenomena, which they term the icehouse-SLIP
Figure 8.1 Comparison of timing and volume of major silicic large igneous provinces (SLIPs) and large igneous provinces (LIPs) with paleoclimate data for the Phanerozoic (After Cather et al., 2009, modified). SLIPs (red) are modified from Bryan (2007) and represent silicic volcanic provinces with documented volumes >10^5 km^3. Numbers in white boxes are minimum eruptive volumes in millions of cubic kilometers (Bryan, 2007); major silicic volcanic episodes with uncertain eruptive volumes are queried. LIPs (green, range of Ar-Ar and U-Pb ages and yellow, main pulses) are modified from Ernst (2014) and Ernst and Youbi (2017). The age range of Ordovician-Silurian episode of major explosive volcanism is from Huff et al. (1998) and Huff et al. (2000), and that of Late Devonian-early Carboniferous silicic volcanism in Australia is from Bryan et al. (2004). Purple bands are peak pulses of volcanism showing age ranges (Ma) from Bryan (2007), Pankhurst et al. (2000), Huff et al. (1992), and Min et al. (2001). Note that temporally overlapping episodes of major silicic volcanism in northern Europe (ca. 300–280 Ma ago; Neumann et al., 2004), in the Parana-Etendeka province (ca. 133–128 Ma ago; Peate, 1997), and in Afro-Arabia (ca. 30–28 Ma ago; Ukshtins Peate et al., 2003) are omitted for clarity. Permian ignimbrite volcanism in South America (López-Gamundi et al., 1994; Breitkreuz & Van Schmus, 1996) is not plotted for lack of adequate age and volume information. See Figure 2 of Cather et al. (2009) for details of Cenozoic ignimbrite flare-up (IFU) volcanism. Light pink and light purple columns allow visual comparison between volcanic episodes and cold paleoclimate intervals. Timing and paleolatitudinal distribution of glaciogenic detritus and other features (blue) and peak glacial intervals (dark blue) are from Frakes and Francis (1988), Frakes et al. (1992), Crowell (1999), Crowley (2000), Isbell et al. (2003a,b), Brenchley et al. (1994), Saltzman and Young (2005), Cherns and Wheelely (2007), Grahn and Caputo (1992), Dromart et al. (2003), Pirrie et al. (1995), Alley and Frakes (2003), Gröcke et al. (2005), and Zachos et al. (2001). Possible short-lived Late Cretaceous-Eocene glacial events in Antarctica (e.g., Miller et al., 2005; Bornemann et al., 2008) are not depicted. Mean tropical sea-surface temperature (black line) has been detrended and smoothed using a 50 Ma window stepping at 10 Ma increments (Veizer et al., 2000) but has not been corrected for pH of seawater (see Royer et al., 2004). Timescale is from Gradstein et al. (2004).
hypothesis (Fig. 8.1). There is also a potential link with LIPs and black-shale-forming oceanic anoxia events (OAEs) (Turgeon & Creaser, 2008; Zhang et al., 2018).

8.1.3. How LIPs Can Cause the End of a Glaciation

Once an ice age is established, planetary albedo increases because snow and ice cover reflects more energy back into space, which would tend to preserve the colder climate (Willeit & Ganopolski, 2018). Therefore, \( \text{CO}_2 \) emissions from volcanism would be key to taking the planet out of a global ice age. An alternative mechanism is that the low background (i.e., non-LIP-related) volcanic flux on Earth could continue to cause an accumulation of \( \text{CO}_2 \) to reach a tipping point, and so a LIP event is not required to end an ice age. Nonetheless, a LIP event that quickly releases large amounts of \( \text{CO}_2 \) into the atmosphere could cause the sudden end of the glaciation. Multi-eruption modeling of the \( \text{CO}_2 \) emissions associated with the 66 Ma Deccan Traps, for example, suggests plausibly large enough values to have contributed to the warming observed across the Cretaceous/Paleogene mass extinction (Tobin et al., 2017).

8.2. POSSIBLE LINKS BETWEEN GLACIATIONS AND LIPS

Here we step through the main glaciations since the Neoproterozoic and consider LIPs that are temporally correlated (or not). See summary in Table 8.1. Some correlations have been previously noted and others are proposed here for the first time. We grade each glaciation-LIP correlation as either “robust,” “possible,” or “no known correlation.” More precise dating of both LIPs and glaciations is required to properly assess the proposed age correlations.

8.2.1. Sturtian Glaciation (ca. 717–660 Ma)

The ca. 717–660 Ma Sturtian glaciation (Macdonald et al., 2010; Young 2013; Rooney et al., 2015) represents the first major glaciation since the Paleoproterozoic (Table 8.1; Fig. 8.2). It is recognized on many blocks (See Fig. 4B of Hoffman et al., 2017) and was preceded by a deep negative \( \delta^{13} \text{C} \) anomaly, the Islay excursion (Strauss et al., 2014).

The start of the Sturtian glaciation has been matched to the timing of the 720 Ma Franklin LIP of northern Canada (Macdonald et al., 2010) and coeval Irkutsk LIP in formerly attached southern Siberia (Ernst & Bleeker, 2010; Ernst et al., 2016). Both the low paleolatitude and the emplacement into an evaporite basin might have rendered the Franklin-Irkutsk LIP climatically important for abetting the onset of the Sturtian glaciation due to enhanced weathering and sulfate aerosol injection into the stratosphere, respectively (Macdonald & Wordsworth, 2017). Additional magmatism of this age is also present in the Kalahari craton (Mutare swarm) and formerly attached Dronning Maud Land region of Antarctica (Fingeren) (Gumsley et al., 2019).

There is, however, no known LIP associated with the end of the Sturtian glaciation ca. 660 Ma, and it therefore is possible that the termination of the glaciation was due to ambient buildup of \( \text{CO}_2 \) from a continuous background level of global volcanism.

8.2.2. Marinoan Glaciation (ca. 640–635 Ma)

The other important Neoproterozoic glaciation is the ca. 640–635 Ma Marinoan glaciation (Table 8.1, Fig. 8.2; Hoffman et al., 1998; Schrag et al., 2002; Prave et al., 2016b). Its start is not well defined, and the size of the gap with the end of the prior 720–660 Ma Sturtian glaciation is also poorly known. Like the Sturtian glaciation, diamictics of the Marinoan glaciation are terminated by transgressive sequences and postglacial cap carbonate units, which provide evidence for a strong \( \text{CO}_2 \) hysteresis leading to extreme weathering during the postglacial supergreenhouse (Bao et al., 2008; Hoffman et al., 1998; Hoffman et al., 2017).

We speculate on a potential climatic link with the newly recognized ca. 650–630 Ma Wudang dyke swarm of South China (Zhao & Asimow, 2018). Numerous ca. 650–630 Ma mafic and ultramafic dykes intruded the Neoproterozoic Wudang Group in the northern Yangtze Block, South China (Wudang mafic dykes). U-Pb zircon ages from the Wudang mafic dykes span from ca. 651–627 Ma and their petrogenesis argues that the dykes are most likely interpreted as a LIP and not related back-arc extension (Zhao & Asimow, 2018). More precise U-Pb geochronology is needed to determine a link with either the start or end of the Marinoan glaciation.

8.2.3. The Gaskiers Glaciation (579 Ma)

Following the two presumed snowball glaciations (Sturtian and Marinoan), Ediacaran glacial deposits have been found on nine different continents (Hoffman & Li, 2009; see also McGee et al., 2015). Assuming synchronicity, this 579 Ma Ediacaran event referred to as the Gaskiers glaciation was likely too short (<340,000 years) to have been a multimillion-year-long snowball glaciation (Pu et al., 2016). Except for those in low-latitude Australia, Ediacaran glacial deposits are found on mid-latitude or high-latitude continents, similarly inconsistent with a snowball Earth origin (Hoffman & Li, 2009).

The age of the Ediacaran Gaskiers glaciation is remarkably similar to the middle, most extensive
Figure 8.2 Selected major trends in Cryogenian and Ediacaran geologic history (modified from Ogg et al., 2016). The carbon-isotope curve is a smoothed version modified by Ogg et al. (2016) from the synthesis for the late Proterozoic by Cohen and Macdonald (2015) calibrated by them to the Cryogenian timescale of Rooney et al. (2015). Ranges and images of organic-walled microfossils, Ediacaran metazoans, and bioturbation styles are from Narbonne et al. (2012). Timing of events associated with continental collision and convergent and divergent plate margin activities related to assembly and dispersal of Rodinia and subsequent assembly of Gondwana is from Cawood et al. (2016). Timing of large igneous provinces (LIPs) is from Ernst (2014), Ernst and Youbi (2017), Macdonald and Wordsworth (2017). Additional geochemical trends, biostratigraphic ranges, regional stages, and details on calibrations are compiled in Shields-Zhou et al. (2012) and Narbonne et al. (2012). SE = Shuram; Tr = Trezona; Tai = Taishir; ICIE = Islay carbon-isotope excursions; Gond = Gondwana.
whether warming (through CO₂ release) or cooling cise timing and nature of their climatic contributions, matism), also need to be evaluated to determine the pre­
the Gaskiers glaciation (consisting of mainly mafic mag­
and Baltica during this ca. 580 Ma time interval spanning 2018, 2019, 2020). Other portions of CIMP in Laurentia
f
f
f
f

The Ouarzazate group of Morocco includes both a SLIP injection of a shared CIMP LIP, until the opening of the Iapetus Ocean (Robert et al., 2018; Mitchell et al., 2011).

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The ca. 511 Ma Kalkarindji LIP is widespread in

The 540–530 Ma Wichita LIP is exposed in the Ouchita aulacogen of the southern United States, but is covered by younger rocks outside of it. Geology and geophysics reveal suites such as the Roosevelt gabbros, the Glen Mountains mafic-ultramafic layered intrusion, and the Navajo Mountain basalt–spilite group (variably altered basaltic to intermediate volcanic rocks), as well as a significant volume of associated A-type silicic magmatism consisting of Wichita granites and Carleton rhyolites (e.g., Hanson et al., 2013). Given its 540–530 Ma age, the Wichita LIP can be tentatively linked to the cause of the Lower Cambrian glaciation (ca. 535 Ma) in Avalonia and Baltica.

The ca. 511 Ma Kalkarindji LIP is widespread in Australia as basaltic remnants of a much more extensive flood basalt province that also includes intrusive components (Jourdan et al., 2014; Marshall et al., 2018, and Chapter 19 this volume, and references therein). Magmatism likely began at around 512 Ma, with final eruption occurring between 509 and 498 Ma. The Kalkarindji LIP (Australia) shows synchrony with the Early and Middle Cambrian (Stage 4–5) extinction and could have also been linked to the possible cooling period and/or glaciation that occurred during the Middle to Late Cambrian, between ca. 513 and ca. 488 Ma.

8.2.5. The Hirnantian Glaciation (ca. 440 Ma)

The end-Ordovician mass extinction consists of two pulses, a base Hirnantian (end-Katian) pulse and a late Hirnantian pulse, with ages of ca. 445.2 Ma and ca. 443.8 Ma, respectively (Ogg et al., 2016). The former is associated with the onset of global cooling and associated sea level drop (nearly 100 m) and the latter is associated with a global reworking, sea level rise and widespread anoxia (Harper et al., 2014). Furthermore, the glaciation is preceded by a period of warming in the late Katian known as the Boda event (Fortey & Cocks, 2005; Lefebvre et al., 2010).

There are indications of magmatism being involved in the end-Ordovician extinction. Buggisch et al. (2010) concluded the presence of major volcanism, on the basis of the prominent Deicke, Millbrig K-bentonite beds of North America and the Kinnekulle K-bentonite bed of Europe, both related to SLIP (i.e., supereruptions, Fig. 8.1; Cather et al., 2009), the latter precisely dated at 454.52 +/- 0.50 Ma (U-Pb zircon; Svensen et al., 2015). Recently, five of the K-bentonites have been dated by high-precision chemical abrasion-thermal ionization mass spectrometry (CA-TIMS) U-Pb zircon geochronology, where the Kinnekulle K-bentonite gives an age of 454.06 ± 0.43 Ma. The entire volcanic activity recorded in the Ordovician at Sinsen section in Oslo has a duration of 4.05 ± 0.91 Myr (Ballo et al., 2019).

Hg values in marine strata exposed at the Wangjiawan Riverside section in South China show a significant peak that links the Hirnantian glacial maximum and the Late Ordovician (ca. 444 Ma) mass extinction directly to mafic volcanism (Jones et al., 2017). LIP candidates of approximately the correct age have been suggested (e.g., Kravchinsky, 2012; Khudoley et al., 2013; Retallack, 2015). These include the ca. 450 Ma (457 ± 34 Ma – 379 ± 27 Ma) Suordakh dolerite event in eastern Siberia (Khudoley et al., 2013; Chamberlain et al., 2018), the Ongnyeobong Formation volcanics in South Korea (Cho et al., 2014), flood basalts of Sierra del Tigre in Argentina (González-Menéndez et al., 2013; Retallack, 2015), and the intraplate basaltic lava flows of the Soltan Maidan Basaltic Complex with thickness up to about 1,300 m located in the eastern Alborz zone, north of Iran (Derakhshi et al., 2017). More precise and better age determinations are required to test the link to the Hirnantian glacial maximum and Late Ordovician mass
extinction. Nonetheless, Hg concentrations in marine strata are known to be associated with well-studied LIPs, such as the Siberian Traps, the CAMP, and the Deccan Traps, suggesting that their occurrence immediately before the Hirnantian glacial maximum provides some of the first direct evidence linking LIP emplacement with climatic cooling (Jones et al., 2017, and references therein).

Other causes have been proposed for the end-Ordovician extinction. Widespread glaciations have been linked to silicate weathering during major orogenic episodes (e.g., Raymo, 1991; Berner, 2004). Major land plant innovations have also been thought to be a significant factor in causing or at least favoring glaciations. Lenten et al. (2012) applied this observation specifically to the end-Ordovician glaciation, as did Kidder and Worsley (2010).

### 8.2.6. Devonian Glaciations

Widespread glaciation occurred in Gondwana in the Late Famennian, at the end of the Late Devonian (e.g., McGhee, 2014a,b). Evidence in Gondwana include biostratigraphically dated glacial tillites, glacial striated pavements and clasts, and ice-rafted dropstones where continental ice sheets covered at least $16 \times 10^6$ km$^2$ of land (Lopez-Gumundi & Buatois, 2010). In Laurussia, there were also Late Famennian glacial tillites produced by lowland glaciers as close as 30$^\circ$S to the equator. This ca. 360 Ma end-Devonian glaciation (Lopez-Gumundi & Buatois, 2010) can be correlated with the ca. 370–360 Ma Yakutsk-Vilyui LIP of the Siberian craton and the ca. 370 Ma Kola-Dnieper LIP of the Baltica craton.

The Yakutsk-Vilyui LIP is widespread in the Siberian craton and dominantly consists of a giant radiating mafic dyke swarm and radiating rift system converging to a mantle plume center on the eastern side of the craton (Kiselev et al., 2012). The largest volumes of flood basalts are preserved in the Vilyui-Marka rift system. U-Pb geochronology indicates emplacement in two pulses, at ca. 374 Ma and ca. 363.4 Ma, that have been linked with mass extinctions (Ricci et al., 2013; Polyansky et al., 2017, 2018; Ernst et al., 2019). There are also associated carbonatites and the diamondiferous kimberlites. The possible environmental impact of carbonatites and associated alkaline magmatism has been discussed by Ray and Pande (1999).

The ca. 370 Ma Kola-Dnieper LIP of Baltica (also known as the East European craton) includes coeval mafic magmatism of the Dnieper-Donets rift, a dyke swarm extending for 2,000 km along the Urals, and magmatism in the Kola peninsula and elsewhere in the Baltic craton (Nikishin et al., 1996; Kravchinsky, 2012; Puchkov et al., 2016). Two plume centers are recognized (Puchkov et al., 2016). Kimberlites (Archangelsk) and carbonatites (Kola Alkaline province) are also associated with this event.

### 8.2.7. Permo-Carboniferous Glaciations (300–260 Ma)

In Permo-Carboniferous time, the so-called P1, P2, P3, and P4 glaciations occurred (Fig 8.3; Table 8.1; Fig. 10.1 in Ogg et al., 2016). The maximum paleolatitudinal range for the glaciations occurred between the middle Stephanian (Gzhelian) stage (ca. 305 Ma ago) and near the end of the Sakmarian stage (ca. 284 Ma ago) (Ibsell et al., 2003a,b, 2012; Fielding et al., 2008; Ogg et al., 2016).

These repeated glaciations can be potentially correlated with widespread intraplate magmatic events (i.e., Shellnut, 2016). These includes the multipulsed (327–250 Ma) European North West African magmatic province (EUNWA or EUNWAMP) (e.g., Doblas et al., 1998; Wilson et al., 2004; Torsvik et al., 2008).

To date, a range of chronometers has been applied to determine eruption ages from across the region of the EUNWA group including whole-rock Rb-Sr and K-Ar dating; $^{40}$Ar/$^{39}$Ar dating of mineral separates; and U-Pb dating of zircon, titanite, and perovskite (e.g., Timmerman et al., 2009). The duration of activity is currently estimated to span a period of ca. 100 million years, from Early Carboniferous to Upper Permian–Early Triassic (350–250 Ma), with several hiatuses (Upton et al., 2004). Three main pulses can be distinguished at ca. 300 Ma, 290–275 Ma, and 250 Ma, and each of these pulses can be considered a separate LIP within the overall EUNWA group.

The distinct 300 Ma pulse of this multipulsed LIP, Skagerrak-Centered Large Igneous Province (the “SCLIP”; Torsvik et al., 2008; see also Ernst & Buchan, 1997) can be considered plume related. The start of this glacial time period (associated with P1) coincides with a particularly well-defined portion of this EUNWA event, the 300 Ma SCLIP pulse that includes magmatism (volcanics, sills, and dykes) in the UK, Norway, Sweden, and in parts of the intervening North Sea, and including a giant radiating dyke swarm focused at the southern end of the Oslo rift (Ernst & Buchan 1997; Torsvik et al., 2008).

The EUNWA event(s) is well represented in Morocco in northwestern Africa and also in southern Scandinavia and northern Germany. The huge volume of extruded and intruded magmatic products of the EUNWA event(s) (example in the Oslo Graben, the estimated volume is at ca. 35,000 km$^3$ while in the North German Basin, the total volume of felsic volcanic rocks, mainly rhyolites and rhyodacites, was of the order of 48,000 km$^3$) has led to suggestions of a thermally anomalous mantle plume to explain this pulse of the magmatism (Ernst & Buchan, 1997; Torsvik et al., 2008). A significant detractor from a plume hypothesis is the duration of activity and the helium isotope signature of lithospheric mantle xenoliths from the Scottish Permo-Carboniferous dykes, sills, and vents (Kirstein et al., 2004). While the EUNWA event(s) qualifies as a significantly large volume of magmatism,
there is some uncertainty whether it technically qualifies as a LIP pending better geochronology.

The EUNWA event(s) may have contributed to the great Gondwanan glaciation that occurred from the Late Devonian to the Late Permian (Veevers & Powell, 1987; Crowell, 1999; Isbell et al., 2003a,b). Glaciers achieved their maximum paleolatitudinal range between the middle Stephanian (ca. 305 Ma ago) and near the end of the Sakmarian (ca. 284 Ma ago) (Isbell et al., 2003a,b). This hypothesis is termed the “icehouse-silicic large igneous province (SLIP)” hypothesis (Cather et al., 2009).

The P2 and P3 glaciations (ca. 290 and ca. 270 Ma, respectively) can be linked with the second and the third pulses of the EUNWA event and the Tarim LIP, which is widespread in Central Asia (Xu et al., 2014) and the poorly dated but also ca. 290 Ma Panjal-Qiangtang LIP (Zhai et al., 2013; Shellnut, 2016), and potentially with the 300–280 Ma Tianshan LIP (Kravchinsky 2010). The P4 glaciation (ca. 260 Ma) approximately matches with the 260 Ma Emeishan LIP of South China (e.g., Shellnut, 2014). The Choiyoi silicic province (possible SLIP) with a peak age of 265 Ma may also have had a role in P3 (Kimbrough, 2016a,b).

For Early Carboniferous time, the Jebilets, Rehamna, and Fourhal basins of western Meseta (Central Hercynian Massif of Morocco) show compelling similarities in tectonostratigraphic development. Their deposits record large instabilities and disorganization with huge thickness and lithological variations related to synsedimentary tectonics. At the same time, tectonic tilting affects the basement of these basins, with blocks controlled by bordering transfer faults. Abundant traces of magmatic activity during the Carboniferous period are recognized in Morocco, particularly in the Jebilets, Rehamna, and Fourhal areas. These rocks constitute a magmatic province (called here the Moroccan Meseta LIP) consisting of basaltic lavas, mafic sills and dykes, and gabbroic intrusions, together with subordinate layered ultramafic intrusions and silicic volcanic/intrusive rocks exposed in the Meseta Domain as part of the Moroccan Variscan belt.
indications of volcanism younger than 181.8 Ma) even however, Svensen et al. (2012) concludes “there are no Ma) on the basis of Ar‐Ar dating (Jourdan et al., 2008). Pulse of the Karoo LIP is thought to be younger (178 Ma) associated with any known LIP activity, except that a Bajocian (170.3 Ma to around 168.3 Ma) cooling is not Late Pliensbachian cooling. LIP event earlier than ca. 183 Ma that would explain the Late Pliensbachian cooling event, but there is no known LIP link.

8.2.9. Cretaceous Glaciations

The Cretaceous (~145–66 Ma) was long considered a warm greenhouse period. However, there is an abundance of evidence for short intervals of cold climatic conditions during the Early Cretaceous (late Valanginian–earliest Hauterivian, late early Aptian, latest Aptian–earliest Albian; Bodin et al., 2015). Each of these intervals is associated with rapid and high amplitude sea level fluctuations, supporting the hypothesis of transient growth of polar ice caps. As evidenced by positive carbon isotope excursions, each cold episode is associated with enhanced burial of organic matter on a global scale.

From a LIPs perspective, there is a relatively good match between the timing and size of LIP eruptions and the amplitude of Early Cretaceous warming episodes (Bodin et al., 2015). In addition, with a slight lag of a few million years in some cases, there is the onset of a cooling. For instance, the late Valanginian–earliest Hauterivian cooling (136–134 Ma, Fig. 3 in Bodin et al., 2015) would correspond to a number of events of age ca. 135 Ma (cf., Bryan & Ferriera, 2013), precisely dated 135–133 Ma Parana–Etendeka (Pinto et al., 2011; Almeida et al., 2018; Hartmann et al., 2019), initial 134 Ma Comei–Bunbury pulse of Kerguelen LIP (Zhu et al., 2009), and ca. 130 Trap LIP event of southwest Greenland. The late early Aptian cooling (124–123 Ma; Fig. 3 in Bodin et al., 2015) could be associated with the precise U‐Pb dating of the initial 126–121 Ma pulse of HALIP (ages mainly from Corfu et al., 2013; see additional ages and summaries in Kingsbury et al., 2018; Dockman et al., 2018) and the latest Aptian earliest Albian (117–112 Ma; Fig. 3 in Bodin et al., 2015) could be associated with stages of the Kerguelen oceanic plateau and also the Rajmahal Traps of eastern India (Fig. 8.4).

The interval following the late Turonian through Maastrichtian (ca. 90–66 Ma) is considered to have been a period of significant global cooling, possibly driven by a combination of declining pCO2 levels and opening ocean gateways (Cramer et al., 2011; Linnert et al., 2014). From a LIPs perspective, this timing starts with three impressive ca. 90–95 Ma LIP events: Caribbean-Colombian, Madagascar, and the second 95 Ma pulse of the High Arctic LIP (HALIP) (Kingsbury et al., 2018), and therefore the possibility of any or all three contributing to global cooling remains plausible. Furthermore,
Large Igneous Provinces (LIPs) record, and their main pulses. Silicic Large Igneous Provinces (SLIPs) record, and their main pulses. Periods of enhanced organic carbon burial on a global scale (as indicated by positive $\delta^{13}C$ excursions).

Figure 8.4 Evolution of major events in Early Cretaceous as recorded in the Vocontian basin (after Bodin et al., 2015, modified). LIPs and SLIPs records are modified from Ernst (2014) and Ernst and Youbi (2017). Note the links between LIPs and the climatic state (H = Hothouse, G = Greenhouse, and C = Coldhouse) as already noted by Bodin et al. (2015).
this cooling trend post-LIP emplacement may have continued as is observed due to the long-term effect of carbon burial associated with OAE2 reducing CO₂ (Berner, 2006), which itself may have been triggered by these aforementioned LIPs (Turgeon & Creaser, 2008). The end of this interval would correlate in timing with the 66 Ma Deccan LIP, which has been modeled to have driven significant climatic warming (Tobin et al., 2017).

8.2.10. Cenozoic Glaciations

The Paleocene and Eocene epochs of the early Cenozoic (66–33 Ma) are regarded as greenhouse climates on account of the highly elevated concentrations of CO₂ (>750 ppm; Zachos et al., 2001). Following this peak warmth of the Cenozoic, however, marine oxygen isotope records indicate a global cooling trend that appears to have culminated rapidly at the Eocene/Oligocene boundary 33.9 Ma (Zachos et al., 2001; Mudelsee et al., 2014; Fig. 8.5a). The first continental-scale glaciation of Antarctica occurred in the earliest Oligocene epoch (33.9 Ma), followed by the onset of northern-hemispheric glacial cycles in the late Pliocene epoch about 31 Myr later (Zachos et al., 2001).

The ca. 30 Ma Afro Arabian LIP is associated with opening of the Red Sea and Gulf of Aden and extension in the East African rift system. Precise dates indicate a main pulse of LIP magmatism at 31–29 Ma (e.g., Ukstins et al., 2002), which would be a few million years too young to be the cause of the 33.9 Ma onset of the continental
scale glaciation in Antarctica (the Oi2 global cooling event). However, Prave et al. (2016a) obtained geological and Ar-Ar and U-Pb geochronological data that define four pulses of volcanism for the Lake Tana region in the northern Ethiopian portion of the Afro-Arabian LIP, the oldest of which is a ∼1-km-thick flood basalt likely as old as ∼34 Ma (40Ar/39Ar age of 34.05 ± 0.54/0.56 Ma), but of unknown duration. This 34 Ma age is older than the 31–29 Ma ages typically attributed to Ethiopian flood basalts and could be responsible for the 33.9 Ma global cooling event. The felsic volcanism was the product of super eruptions that created a 60–80 km diameter caldera marked by km-scale caldera-collapse fault blocks and a steep-sided basin filled with a minimum of 180 m of sediment and the present-day Lake Tana (Prave et al., 2016a) (Fig. 8.5). CO₂ decline in the late Eocene has been implicated to drive glaciation on Antarctica (DeConto & Pollard, 2003), and we argue that the Afro-Arabian LIP provided the driver for this CO₂ drawdown.

8.3. DISCUSSION AND CONCLUSIONS

1. LIPs and SLIPs magmatism can have significant global climatic effects including causing mass extinction events via a complex web of changes (characteristically rapid) in atmospheric/oceanic acidification, oceanic anoxia, sea level, toxic metal input (e.g., mercury: Hg), and most significantly in temperature, both warming and cooling. In the broadest sense, LIPs and SLIPs can affect (or even induce) shifts between icehouse, greenhouse, and hot-house climatic states (e.g., Kidder & Worsley 2010, 2012).

2. Earth has gone through periods of cooling including global, near global, or regional glaciations, which are observed in the Archean, Paleoproterozoic, Neoproterozoic, Ordovician, Permo-Carboniferous, and Cenozoic times. We have reviewed the ways in which LIPs and SLIPs can cause global cooling: We focus on glaciations of Neoproterozoic and Phanerozoic age to identify potentially age-correlated LIP events. The 720 Ma Franklin-Irkutsk LIP is linked with the start of the Sturtian glaciation and the Wudang dyke swarm (ca. 650–630 Ma) is a LIP that spans the Marinoan glaciation. The ca. 580 Ma pulse of the Central Iapetus Magmatic Province (CIMP) is associated with the start and end of the 579 Ma Gaskiers glaciation. The Hirnantian glaciation (ca. 440 Ma) may be associated with poorly dated ca. 440 Ma intraplate magmatism in several regions, including eastern Siberia, South Korea, Argentina, and elsewhere. Permo-Carboniferous glaciations (P1–P4, 300–260 Ma) can be correlated with widespread intraplate magmatism of the European northwest African magmatic province (and its initiation as the 300 Ma Skagerrak LIP), and also the 260 Ma Emeishan LIP of China. The ca. 30 Ma initial pulse of the Afro-Arabian LIP approximately matches the Eocene-Oligocene cooling. More precise dating of both the LIP and cooling events is required to confirm the correlations and to assess the role of LIPs relative to other causes of global cooling.

3. A variety of causes have been considered for global and regional glaciations. For instance, glaciations have been linked to silicate weathering (and CO₂ drawdown) during major orogenic episodes such as the formation of the Himalayas (Cenozoic glaciation) and the assembly of Pangea (Permo-Carboniferous glaciation). Major land plant innovations (and their ability to extract CO₂ and release oxygen) have also been thought to be a significant factor in causing or at least favoring glaciations (e.g., for the Ordovician and Permo-Carboniferous glaciations). In addition, it is also now recognized that LIPs can possibly contribute to global cooling via at least three different mechanisms: (1) LIP input of SO₂ into the atmosphere (and conversion to sulfate aerosols), (2) weathering of LIP units and CO₂ drawdown, and (3) increased oceanic biologic productivity and resulting increased CO₂ drawdown (SLIPs are particularly important in the latter).

4. There are many LIP events that are not linked to shifts to hotter or colder conditions, and so it could be inferred that environmental conditions at the time of LIP emplacement must already be close to a tipping point such that the LIP provides the final critical threshold that causes a rapid climatic shift. In the present context, we could view the non-LIP factors such as supercontinent breakup and its palaeogeography during rifting and continental arc (McKenzie et al., 2016; Macdonald et al., 2019) and submarine volcanism as external forces to have the effect of preparing the climate for a superimposed shock from LIP magmatism (silicic initially, and then mafic). These other factors are generally of longer duration and so the changes they would introduce would also be of longer duration. In contrast, a short-duration LIP pulse could cause a sudden change in climate.

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REFERENCES

Algeo, T., Marenco, P. J., & Saltzman, M. R. (2016). Co-evolution of oceans, climate, and the biosphere during the “Ordovician Revolution”: A review. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 458, 1–11. http://dx.doi.org/10.1016/j.palaeo.2016.05.015

Alley, N. F., & Frakes, L. A. (2003). First known Cretaceous glaciation: Livingston Tillite member of the Cadna-owie Formation, South Australia. *Australian Journal of Earth Sciences*, 50, 139–144.

Almeida, V. V., Janasi, V. A., Heaman, L. M., Shaulis, B. J., Hollanda, M. H. B. M., & Renne, P. (2018). Contemporaneous alkaline and tholeiitic magmatism in the Ponta Grossa Arch, Paraná-Étendeka Magmatic Province: Constraints from precise U-Pb zircon/baddeleyite and 40Ar-39 Ar phlogopite dating of the José Fernandes Gabbro and mafic dykes. *Journal of Volcanology and Geothermal Research*, 355, 55–65.

Augustin, L., Barbante, C., Barnes, P., et al. (2004). Eight glacial cycles from an Antarctic ice core. *Nature*, 429, 623–628. http://dx.doi.org/10.1038/nature02599

Ballo, E. G., Augland, L. E., Hammer, O., & Svensen, H. H. (2019). A new age model for the Ordovician (Sandbian) K-bentonites in Oslo, Norway. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 520, 203–213. doi.org/10.1016/j.palaeo.2019.01.016

Bao, H., Lyons, T. W., & Zhou, C. (2008). Triple oxygen isotope evidence for elevated CO₂ levels after a Neoproterozoic glaciation. *Nature*, 453, 504–506. doi:10.1038/nature06959

Berner, R. A. (2004). *The Phanerozoic carbon cycle*. Oxford: Oxford University Press.

Berner, R. A. (2006). GEOCARBSULF: A combined model for Phanerozoic atmospheric CO₂ and O₂. *Geochimica et Cosmochimica Acta*, 70, 5653–5664.

Best, M. G., & Christiansen, E. H. (1991). Limited extension and mass extinctions: An update. In G. Keller & A. C. Kerr (Eds.), *Volcanism, impacts and mass extinctions: Causes and effects* (pp. 29–55). Geological Society of America Special Paper 505.

Bornemann, A., Norris, R. D., Friedrich, O., Beckmann, B., Schouten, S., Damsté, J. S. S., Vogel, J., et al. (2008). Isotopic evidence for glaciation during the Cretaceous supergreenhouse. *Science*, 319, 189–192.

Boyce, C. K., & Lee, J. E. (2017). Plant evolution and climate over geological timescales. *Annual Review of Earth and Planetary Sciences*, 45, 61–87

Bradley, D. C. (2011). Secular trends in the geologic record and the supercontinent cycle. *Earth Science Review*, 108, 16–33. http://dx.doi.org/10.1016/j.earscirev.2011.05.003

Breitkreuz, C., & Van Schmus, W. R. (1996). U-Pb geochronology and significance of Late Permian ignimbrites in northern Chile. *Journal of South American Earth Sciences*, 9, 281–293.

Brenchley, P. J., Marshall, J. D., Carden, G. A. F., Robertson, D. B. R., Long, D. G. F., Meidla, T., Hints, L., et al. (1994). Bathymetric and isotopic evidence for a short-lived Late Ordovician glaciation in a greenhouse period. *Geology*, 22, 295–298.

Bryan, S., & Ernst, R. E. (2008). Revised definition of Large Igneous Provinces (LIPs). *Earth Science Reviews*, 86, 175–202.

Bryan, S. E. (2007). Silicic large igneous provinces. *Episodes*, 30(1), 20–31.

Bryan, S. E., & Ferrari, L. (2013). Large Igneous Provinces and Silicic Large Igneous Provinces: Progress in our understanding over the last 25 years. *Geological Society of America Bulletin*, 125, 1053–1078.

Bryan, S. E., Allen, C. M., Holcombe, R. J., & Fielding, C. R. (2004). U-Pb zircon geochronology of Late Devonian to Early Carboniferous extension-related silicic volcanism in the northern New England fold belt. *Australian Journal of Earth Sciences*, 51, 645–664. doi:10.1111/j.1400-0952.2004.01079.x

Buggisch, W., Joachimski, M. M., Lehner, O., Bergström, S. M., Repetski, J. E., & Webers, G. F. (2010). Did intense volcanism trigger the first Late Ordovician icehouse? *Geology*, 38(4), 327–330.

Burgess, S. D., & Bowring, S. A. (2015). High-precision geochronology confirms voluminous magmatism before, during and after Earth’s most severe extinction. *Science Advances*, 1(7), e1500470. http://dx.doi.org/10.1126/sciadv.1500470

Burgess, S. D., Bowring, S. A., Fleming, T. H., & Elliot, D. H. (2015a). High-precision geochronology links the Ferrar large igneous province with early-Jurassic ocean anoxia and biotic crisis. *Earth and Planetary Science Letters*, 415, 90–99.

Burgess, S. D., Muirhead, J. D., & Bowring, S. A. (2015b). Initial pulse of Siberian Traps sills as the trigger of the end-Permain mass extinction. *Nature Communications*, 8(1), 164. doi:10.1038/s41467-017-00083-9

Callegaro, S., Baker, D. R., De Min, A., Marzoli, A., Geraki, K., Bertrand, H., Viti, C., & Nestola, F. (2014). Microanalyses link sulfur from large igneous provinces and Mesozoic mass extinctions. *Geology*, 42(10), 895–898. http://dx.doi.org/10.1130/G35983.1

Caputo, M. V., & Crowell, J. C. (1985). Migration of glacial centers across Gondwana during Paleozoic Era. *Geological Society of America Bulletin*, 96, 1020–1036.

Cather, S. M., Dunbar, N. W., McDowell, F. W., McIntosh, W. C., & Scholle, P. A. (2009). Climate forcing by iron fertilization from repeated ignimbrite eruptions: the icehouse-silicic large igneous province (SLIP) hypothesis. *Geosphere*, 5, 315–324.
Cawood, P. A., Strachan, R. A., Pisarevsky, S. A., Gladkochub, D. P., & Murphy, J. B. (2016). Linking collisional and accretionary orogens during Rodinia assembly and breakup: Implications for models of supercontinent cycles. Earth and Planetary Science Letters, 449, 118–126.

Chamberlain, K. R., Khudoley, A. K., & Ernst, R. E. (2018). Improved U-Pb dating of the ca. 450 Ma Suordakh mafic event in eastern Siberia will test whether this is the missing LIP related to end-Ordovician mass extinction. Petrology of Magmatic and Metamorphic Complexes X All-Russian Science Conference With International Participation, 27–30 November 2018, Tomsk Russia.

Chapin, C. E., Wilks, M., McIntosh, W. C., & Cather, S. M. (2018). Late Cretaceous–Neogene trends in deep ocean temperature and continental ice volume: Reconciling records of benthic foraminiferal geochemistry 818O and Mg/Ca with sea level history. Journal of Geophysical Research, 116, C12023.

Crowell, J. C. (1999). Pre-Mesozoic ice ages: Their bearing on understanding the climate system. Geological Society of America Memoir 192. Boulder.

Crowley, T. J. (2000). Causes of climate change over the past 1000 years. Science, 289, 270–277.

Davies, J., Marzoli, A., Bertrand, H., Youbi, N., Ernesto, M., & Schaltegger, U. (2017). End-Triassic mass extinction started by intrusive CAMP activity. Nature Communications. doi: 10.1038/NCOMM15596

DeConto, R. M., & Pollard, D. (2003). Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO2. Nature, 421, 245–249.

Denyszyn, S. W., Halls, H. C., Davis, D. W., & Evans, D. A. D. (2009). Paleomagnetism and U-Pb geochronology of Franklin dykes in high Arctic Canada and Greenland: A revised age and paleomagnetic pole constraining block rotations in the Nares Strait region. Canadian Journal of Earth Sciences, 46, 689–705.

Derakhshi, M., Ghasemi, H., & Miao, L. (2017). Geochemistry and petrogenesis of Soltan Maidan basalts (E Alborz, Iran): Implications for asthenosphere-lithosphere interaction and rifting along the N margin of Gondwana. Chemie der Erde, 77, 131–145. doi: 10.1016/j.chemer.2017.01.002

Dessert, C., Dupre, B., Gaillardet, J., Francois, L. M., & Allegre, C. J. (2003). Basalt weathering laws and the impact of basalt weathering on the global carbon cycle. Chemical Geology, 202, 257–273.

Doblas, M., Oyarzun, R., Lopez-Ruiz, J., Cebria, J. M., Youbi, N., Mahecha, V., Lago, M., et al. (1998). Permo-Carboniferous volcanism in Europe and northwest Africa: A superplume exhaust valve in the center of Pangaea. Journal of African Earth Science, 26, 89–99.

Dockman, D. M., Pearson D. G., Heaman, L. M., Gibson, S. A., & Sarkar, C. (2018). Timing and origin of magmatism in the Sverdrup basin, Northern Canada: Implications for lithospheric evolution in the high Arctic Large Igneous Province (HALIP). Tectonophysics, 742–743, 50–65. https://doi.org/10.1016/j.tecto.2018.05.010

Dromart, G., Garcia, J.-P., Picard, S., Atrops, F., Lécuyer, C., & Sheppard, S. M. F. (2003). Ice age at the Middle-Late Jurassic transition? Earth and Planetary Science Letters, 213 (3–4), 205–220.

Ernst, R. E. (2014). Large igneous provinces. Cambridge: Cambridge University Press.

Ernst, R. E., & Bell, K. (2010). Large igneous provinces (LIPs) and carbonatites. Mineralogy and Petrology, 98, 55–76. doi: 10.1007/s00710-009-0074-1

Ernst, R. E., & Bleeker, W. (2010). Large igneous provinces (LIPs), giant dyke swarms, and mantle plumes: Significance for breakup events within Canada and adjacent regions from 2.5 Ga to the present. Canadian Journal of Earth Science, 47, 695–739.

Ernst, R. E., & Buchan, K. L. (1997). Giant radiating dyke swarms: Their use in identifying Pre-Mesozoic large igneous provinces and mantle plumes. In J. Mahoney & M. Coffin (Eds.), Large igneous provinces: Continental, oceanic, and planetary volcanism (pp. 297–333). American Geophysical
Union, Geophysical Monograph Series, 100. doi.org/10.1029/GM100p0297
Ernst, R. E., & Youbi, N. (2017). How large igneous provinces affect global climate, sometimes cause mass extinctions, and represent natural markers in the geological record. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 478, 30–52, 726. doi: 10.1016/j.palaeo.2017.03.014
Ernst, R. E., Hamilton, M. A., Söderlund, U., Hanes, J. A., Gladkochub, J. A., et al. (2016). Long-lived connection between southern Siberia and northern Laurentia in the Proterozoic. *Nature Geoscience*, 9, 464–469.
Ernst, R. E., Rodgyn, S. A., & Grinev, O. M. (2019). Large igneous provinces as probable trigger of Devonian biotic crises. *Global Planetary Change* (in press).
Evans, D. A., Beukes, N. J., & Krischvink, J. L. (1997). Low-latitude glaciation in the Paleoproterozoic era. *Nature*, 386, 262–266. http://dx.doi.org/10.1038/386262a0
Eyles, N. (2008). Glacio-epochs and the supercontinent cycle after ~3.0 Ga: Tectonic boundary conditions for glaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 258, 89–129.
Ferrari, L., López-Martínez, M., & Rosas-Elguera, J. (2002). Igminbrite flare-up and deformation in the southern Sierra Madre Occidental, western Mexico: Implications for the late subduction history of the Farallón plate. *Tectonics*, 21, 17.
Fielding, C. R., Frank, T. D., Birgenheier, L. P., Rygel, M. C., Jones, A. T., & Roberts, J. (2008). Stratigraphic record and facies associations of the late Paleozoic ice age in eastern Australia (New South Wales and Queensland). In C.R. Fielding et al. (Eds.), *Resolving the Late Paleozoic ice age in time and space* (pp. 41–57). Geological Society of America Special Paper 441. doi: 10.1130/2008.2441(03)
Fortey, R. A., & Cocks, L. R. M. (2005). Late Ordovician global warming: The Boda event. *Geology*, 33(5): 405–408. https://doi.org/10.1130/G21180.1
F rak e s, L. A., & Francis, J. E. (1988). A guide to Phanerozoic cold polar climates from high-latitude ice-rafting in the Cretaceous. *Nature*, 333, 547.
F rak e s, L. A., Francis, J. E., & Syktus, J. I. (1992). *Climate modes of the Phanerozoic: The history of the Earth’s climate over the past 600 million years*. Cambridge: Cambridge University Press.
Fu, R. R., & Kent, D. V. (2018). Anomalous Late Jurassic motion of the Pacific plate with implications for true polar wander. *Earth and Planetary Science Letters*, 490, 20–30.
Fu, R. R., Kent, D. V., Hemming, S. R., Gutiérrez, P., & Creveling, J. R. (2020). Testing the occurrence of Late Jurassic true polar wander using the La Negra volcanics of northern Chile. *Earth and Planetary Science Letters*, 529, 115835. https://doi.org/10.1016/j.epsl.2019.115835
Goddéris, Y., Donnadieu, Y., Le Hir, G., Lefèvre, V., & Nardin, E. (2014). The role of palaeogeography in the Phanerozoic history of atmospheric CO2 and climate. *Earth-Science Reviews*, 128, 122–138.
Goddéris, Y., Donnadieu, Y., Nédélec, A., et al. (2003). The Sturtian “snowball” glaciation: Fire and ice. *Earth and Planetary Science Letters*, 211, 1–12.
González-Menéndez, L., Gallastegui, G., Cuesta, A., Heredia N., & Rubio-Ordóñez, A. (2013). Petrogenesis of Early Paleozoic basalts and gabbros in the western Cuyania terrane: Constraints on the tectonic setting of the southwestern Gondwanamargin (Sierra del Tigre, Andean Argentine Precordillera). *Gondwana Research*, 24, 359–376.
Gradstein, F. M., Ogg, J. G., Smith, A. G., Agterberg, F. P., Bleeker, W., Cooper, R. A., et al. (2004). *A geologic time scale 2004*. Cambridge: Cambridge University Press, 500.
Grahn, Y., & Caputo, M. V. (1992). Early Silurian glaciations in Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 99, 9–15.
Gröcke, D. R., Price, G. D., Robinson, S. A., Baraboshkin, E. Y., Mutterlose, J., & Ruffell, A. H. (2005). The Upper Valanginian (Early Cretaceous) positive carbon-isotope event recorded in terrestrial plants. *Earth and Planetary Science Letters*, 240, 495–509.
Gumsley, A., Salacińska, A., Knoper, M., Mamuse, A., Chew, D., Söderlund, U., Kamo, S., et al. (2019). *The Kalahari crater in the fiery heart of Rodinia*. Goldschmidt 2019, Barcelona Spain, 18–23 August.
Gumsley, A. P., Chamberlain, K. R., Bleeker, W., Söderlund, U., de Kock, M. O., Larsson, E. R., & Bekker, A. (2017). *Timing and tempo of the Great Oxidation Event*. PNAS, 114 (8), 1811–1816. http://dx.doi.org/10.1073/pnas.1608824114.
Hanson, R. E., Puckett, R. E., Jr, Keller, G. R., Bruecke, M. E., Bulen, C. L., Mertzman, S. A., Finegan, S. A., et al. (2013). Intraplate magmatism related to opening of the southern Iapetus Ocean: Cambrian Wichtia igneous province in the southern Oklahoma rift zone. *Lithos*, 174, 57–70.
Harper, D. A. T., Hammarlund, E. U., & Rasmussen, C. M. Ø. (2014). End Ordovician extinctions: A coincidence of causes. *Gondwana Research*, 23, 1294–1307.
Hartmann, L. A., Baggio, S. B., Brückmann, M. P., Knijnik, D. B., Lana, C., Massonne, H. J. Opitz, J., et al. (2019). U-Pb geochronology of Paraná volcanics combined with trace element geochemistry of the zircon crystals and zircon Hf isotopic data. *Journal of South American Earth Sciences*, 89, 219–226. doi.org/10.1016/j.jsames.2018.11.026
Hastie, W. W., Watkeys, M. K., & Aubourg, C. (2014). Magma flow in dyke swarms of the Karoo LIP: Implications for the mantle-plume hypothesis. *Gondwana Research*, 25, 736–755.
Hoffman, P. F. (2009). Pan-glacial: A third state in the climate system. *Geology Today*, 25, 100–107.
Hoffman, P. F., & Li, Z. X. (2009). A palaeogeographic context for Neoproterozoic glaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 277, 158–172. doi.org/10.1016/j.palaeo.2009.03.013
Hoffman, P. F., Abbot, D. S., Ashkenazy, Y., Benn, D. I., Brooks, J. J., Cohen, P. A., et al. (2017). Snowball Earth climate dynamics and Cryogenian geology-geobiology. *Science Advances*, 3(11), 1–43. doi: 10.1126/sciadv.1600983
Hoffman, P. F., Kaufman, A. J., Halverson, G. P., & Schrag, D. P. (1998). A Neoproterozoic Snowball Earth. *Science*, 281, 1342–1346. doi:10.1126/science.281.5381.1342
Horton, F. (2015). Did phosphorus derived from the weathering of large igneous provinces fertilize the Neoproterozoic ocean? *Geochemistry. Geophysics and Geosystems*, 16 (6), 1723–1738. https://doi.org/10.1002/2015GC005792
Huff, W. D., Bergström, S. M., & Kolata, D. R. (1992). Gigantic Ordovician volcanic ash fall in North America and Europe:
Biological, tectonomagmatic, and event stratigraphic significance. *Geology*, 20, 875–878. doi: 10.1130/0091-7613 (1992)020<0875:GVAFI>2.3.CO;2

Huff, W. D., Bergström, S. M., & Kolata, D. R. (2000). Silurian K-bentonites of the Dnestr Basin, Podolia, Ukraine. *Journal of the Geological Society*, 157, 493–504.

Huff, W. D., Bergström, S. M., Kolata, D. R., & Sun, H. (1998). The Lower Silurian Osmundsberg K-bentonite. Part II: Mineralogy, geochemistry, chemosтратigraphy and tectonomagmatic significance. *Geological Magazine*, 135, 15–26.

Isbell, J. L., Henry, L. C., Gulbranson, E. L., Limarino, C. O., Fraiser, M. L., Koch, Z. J., Ciccioli, P. L., et al. (2012). Glacial paradoxes during the late Paleozoic ice age: Evaluating the equilibrium line altitude as a control on glaciation. *Gondwana Research*, 22, 1–19.

Isbell, J. L., Lenaker, P. A., Askin, R. A., Miller, M. F., & Babcock, L. E. (2003a). Reevaluation of the timing and extent of late Paleozoic glaciation in Gondwana: Role of the Transantarctic Mountains. *Geology*, 31, 977–980.

Isbell, J. L., Miller, M. F., Wolfe, K. L., & Lenaker, P. A. (2003b). Timing of Late Paleozoic glaciation in Gondwana: was glaciation responsible for the development of northern hemisphere cyclothems? In M. A. Chan & A. W. Archer (Eds.), *Extreme depositional environments: Mega end members in geologic time* (pp. 5–24). Geological Society of America Special Paper, 370.

Jones, D. S., Martini, A. M., Fike, D. A., & Kunio Kaiho, K. (2017). A volcanic trigger for the Late Ordovician mass extinction? Mercury data from south China and Laurentia. *Geology*, 45(7), 631–634. https://doi.org/10.1130/G38940.1

Jourdan, F., Féraud, G., Bertrand, H., Watkeys, M. K., & Renne, P. R. (2008). The 40Ar/39Ar ages of the sill complex of the Lower Silurian Osmundsberg K-bentonite. Part II: Mineralogy, geochemistry, chemosтратigraphy and tectonomagmatic significance. *Geological Magazine*, 135, 15–26.

Kimbrough, D. L., Mahoney, J. B., Mescua, J., Giambiagi, L. B., & Buelow, E. (2016a). *The Choiyoi silicic large igneous province of Argentina and Chile and its possible linkage to middle Permian climate change and mass extinction*. Abstract 5112. In 35th Abstract International Geological Congress Abstracts, Cape Town South Africa (available from American Geosciences Institute).

Kimbrough, D. L., Mahoney, J. B., Mescua, J. F., Giambiagi, L. B., & Grove, M. (2016b). *The Choiyoi silicic large igneous province of Argentina and Chile and its possible influence on Permain environmental degradation and mass extinction*. Geological Society of America Annual Meeting, Denver, Colorado, USA.

Kirstein, L. A., Dunai, T. J., Davies, G. R., Upton, B. G. J., & Nikogosian, I. K. (2004). Helium isotope signature of lithospheric mantle xenoliths from the Permo-Carboniferous magmatic province in Scotland: No evidence for a lower-mantle plume. In M. Wilson et al. (Eds.), *Perm-Carboniferous magmatism and rifting in Europe* (pp. 243–258). Geological Society, London, Special Publications, 223. doi: 10.1144/GSL.SP.2004.223.01.11

Kiselev, A. I., Ernst, R. E., Yarmolyuk, V. V., & Egorov, K. N. (2012). Radiating rifts and dyke swarms of the middle Paleozoic Yakutsk plume, of eastern Siberian craton. *Journal of Asian Earth Sciences*, 45, 1–16.

Kjoll, H. J., Andersen, T. B., Corfu, F., Labrousse, L., Tegner, C., Abdelmalak, M. M., & Planke, S. (2019). Timing of break-up and thermal evolution of a pre-Caledonian Neoproterozoic exhumed magma-rich rifted margin. *Tectonics*, 38, 1843–1862. doi: 10.1029/2018TC005375

Korte, C., & Hesselbo, S. P. (2011). Shallow marine carbon and oxygen isotope and elemental records indicate icehouse-greenhouse cycles during the Early Jurassic. *Paleoceanography*, 26, PA4219

Korte, C., Hesselbo, S. P., Ullmann, C. V., Ruhl, M., Schweigert, G., & Thibault, N. (2015). Jurassic climate mode governed by greenhouse cycles. *Climate of the Past*, 11, 923–936. doi: 10.5194/cp‐9‐923‐2015

Korte, C., & Hesselbo, S. P. (2011). Shallow marine carbon and oxygen isotope and elemental records indicate icehouse-greenhouse cycles during the Early Jurassic. *Paleoceanography*, 26, PA4219

Korte, C., Hesselbo, S. P., Ullmann, C. V., Ruhl, M., Schweigert, G., & Thibault, N. (2015). Jurassic climate mode governed by ocean gateway. *Nature Communications*, 6, 10015. doi: 10.1038/ncomms10015

Kračevski, V. A. (2012). Paleozoic large igneous provinces of Northern Eurasia: Correlation with mass extinction events. *Global and Planetary Change*, 86–87, 31–36. doi: 10.1016/j.gloplacha.2012.01.007

Landing, E. (2011). No Late Cambrian shoreline ice in Laurentia. *GSA Today*, 21, doi: 10.1130/G113C.1, e19
Landing, E., & MacGabhann, B. A. (2010). First evidence for Cambrian glaciation provided by sections in Avalonian New Brunswick and Ireland-Additional data for Avalon-Gondwana separation by the earliest Palaeozoic. Palaeogeography, Palaeoclimatology, Palaeoecology, 295, 174–185.

Larsen, R. B., Grant, T., Sørensen, B. E., Tegner, C., McEnroe, S., Pastore, Z., Fichler, C., et al. (2018). Portrait of a giant deep-seated magmatic conduit system: The Seiland Igneous Province. Lithos, 296–299, 600–622. doi:10.1016/j.lithos.2017.11.013

Lefèvre, V., Servais, T., François, L. M., & Averbuch, O. (2010). Did a Katian large igneous province trigger the Late Ordovician glaciation? Palaeogeography, Palaeoclimatology, Palaeoecology, 296, 310–319. doi:10.1016/j.palaeco.2010.04.010

Lenton, T. M., Crouch, M., Johnson, M., Pires, N., & Dolan, L. (2012). First plants cooled the Ordovician. Nature Geoscience, 5, 86–89.

Lindström, M. (1972). Cold age sediment in Lower Cambrian of south Sweden. Geologica et Palaeontologica, 6, 9–23.

Linnert, C., Robinson, S. A., Lees, J. A., Bown, P. R., Pérez-Rodríguez, I., Petrizzo, M. R., Falzoni, F., et al. (2014). Evidence for global cooling in the Late Cretaceous. Nature Communications, 5, 4194. doi:10.1038/ncomms5194

Lipman, P. W. (2000). Central San Juan caldera cluster: Regional volcanic framework. In P. M. Bethke & R. L. Hay (Eds.), Ancient Lake Creede: Its volcano-tectonic setting, history of sedimentation, and relation to mineralization in the Creede mining district (pp. 9–69). Geological Society of America Special Paper 468.

López-Gamundi, O. R., & Buatois, L. A. (2010). Introduction: Late Palaeozoic glacial events and post-glacial transgressions in Gondwana (pp. v–viii). Geological Society of America Special Paper 468.

López-Gamundi, O. R., Espejo, I. S., Conaghan, P. J., Powell, C. M., & Veever, J. J. (1994). Southern South America. In J. J. Veever & C. M. Powell (Eds.), Permian-Triassic Pangean basins and foldbelts along the Panhimalian margin of Gondwanaland (pp. 281–329). doi:10.1130/ME184-p281

Macdonald, F. A., & Wordsworth, R. (2017). Initiation of Snowball Earth with volcanic sulfur aerosol emissions. Geophysical Research Letters, 44(4), 1938–1946. doi:10.1002/2016GL072335

Macdonald, F. A., Schmitz, M. D., Crowley, J. L., Roots, C. F., Jones, D. S., Maloof, A. C., Strauss, J. V., et al. (2010). Calibrating the Cryogenian. Science, 327, 1241–1243. doi:10.1126/science.1183325

Macdonald, F. M., Swanson-Hysell, N. L., Park, Y., Lisiecki, L., & Jagoutz, O. (2019). Arc-continental collisions in the tropics set Earth’s climate state. Science, 364, 181–184.

Marshall, P. E., Halton, A. M., Kelley, S. P., Widdowson, M., & Sherlock, S. C. (2018). New 40Ar/39Ar dating of the Antrim Plateau Volcanics, Australia: Clarifying an age for the eruptive phase of the Kalkarindji continental flood basalt province. Journal of the Geological Society of London, 175, 974–985. https://dx.doi.org/10.1144/jgs2018-035

McGee, B., Collins, A. S., Trindade, R. I., & Jourdan, F. (2015). Investigating mid-Ediacaran glaciation and final Gondwana amalgamation using coupled sedimentology and 40Ar/39Ar detrital muscovite provenance from the Paraguay Belt, Brazil. Sedimentology, 62, 130–154. doi:10.1111/sed.12143

McGhee, G. R., Jr. (2014a). The Late Devonian (Frasnian/Famennian) mass extinction: A proposed test of the glaciation hypothesis. Geological Quarterly, 58(2), 263–268.

McGhee, G. R., Jr. (2014b). The search for sedimentary evidence of glaciation during the Frasnian/Famennian (Late Devonian) biodiversity crisis. Sedimentary Record, 12(2), 4–8.

McDowell, F. W. (2007). Geologic transect across the Sierra Madre Occidental volcanic field, Chihuahua and Sonora, Mexico. Geological Society of America Digital Map and Chart Series DMCH006.

McDowell, F. W., & McIntosh, W. C. (2007). October Timing of intense magmatic episodes in the northern Sierra Madre Occidental, Mexico. Geological Society of America, Abstracts with Programs.

McIntosh, W. C., & Bryan, C. (2000). Chronology and geochemistry of the Boot Heel volcanic field. In T. F. Lawton et al. (Eds.), Southwest passage: A trip through the Phanerozoic (pp. 157–174). New Mexico Geological Society Field Conference Guidebook 51.

McIntosh, W. C., Geissman, J. W., Chapin, C. E., Kunk, M. J., & Henry, C. D. (1992). Calibration of the latest Eocene-Oligocene geomagnetic polarity time scale using 40Ar/39Ar dated ignimbrites. Geology, 20, 459–463.

McKenzie, N. R., Horton, B. K., Loomis, S. E., Stockli, D. F., Planavsky, N. J., & Lee, C. T. A. (2016). Continental arc volcanism as the principal drivers of icehouse-greenhouse variability. Science, 352, 444–447.

Miller, K. G., Wright, J. D., & Browning, J. V. (2005). Visions of ice sheets in a greenhouse world. Marine Geology, 217, 215–231.

Mills, B., Daines, S. J., & Lenton, T. M. (2014). Changing tectonic controls on the long-term climate cycle from Mesozoic to present. Geochemistry, Geophysics, Geosystems, 15, 4866–4884.

Min, K., Renne, P. R., & Huff, W. D. (2001). 40Ar/39Ar dating of Ordovician K-bentonites in Laurentia and Baltoscandia. Earth and Planetary Science Letters, 185, 121–134. doi: 10.1016/S0012-821X(00)00365-4

Mitchell, R. N., Gerton, T. M., Nordsvan, A., Cox, G. M., Li, Z. X., & Hoffman, P. F. (2019). Hit or miss: Glacial incisions of snowball Earth. Terra Nova, 31, 381–389. https://doi.org/10.1111/ter.12400

Mitchell, R. N., Kilian, T. M., Rauh, T. D., Evans, D. A. D., Bleeker, W., & Maloof, A. C. (2011). Sutton hotspot: Resolving Ediacaran-Cambrian tectonics and true polar wander for Laurentia. American Journal of Science, 311, 651–663.

Moye, F. J. (1988). Extensional control of Eocene volcanism and plutonism, Idaho and Washington. Geological Society of America, Abstracts with Programs, 20, 434.

Mudelsee, M., Bickert, T., Leary, C. H., & Lohmann, G. (2014). Cenozoic climate changes: A review based on time series analysis of marine benthic δ18O records. Reviews of Geophysics, 52, 333–374. doi:10.1002/2013RG000440

Narbonne, G. M., Xiao, S., Shields, G. A., & Gehling, J. G. (2012). The Ediacaran Period. The Geologic Time Scale, 1, 413–435.
Neumann, E. R., Wilson, M., Heeremans, M., Spencer, E. A., Obst, K., Timmerman, M. J., & Kirstein, L. (2004). Carboniferous-Permian rifting and magmatism in southern Scandinavia, the North Sea, and northern Germany: A review. *Geological Society, London, Special Publications*, 223, 11–40.

Nikishin, A. M., Ziegler, P. A., Stephenson, R. A., et al. (1996). Late Precambrian to Triassic history of the East European Craton: Dynamics of sedimentary basin evolution. *Tectonophysics*, 268, 23–63.

Ogg, J. G., Ogg, G. M., & Gradstein, F. M. (2016). A concise geologic time scale 2016. Elsevier. http://dx.doi.org/10.1016/B978-0-444-59467-9.00008-X

Pankhurst, R. J., Riley, T. R., Fanning, M., & Kelley, S. P. (2011). Zircon U-Pb geochronology from the Paraná bimodal volcanic province support a brief eruptive cycle at ~135 Ma. *Journal of Petrology*, 52, 30–43.

Pinto, V. M., Hartmann, L. A., Santos, J. O. S., McNaughton, N. J., & Wildner, W. (2011). Zircon U-Pb geochronology from the Paraná bimodal volcanic province support a brief eruptive cycle at ~135 Ma. *Chemical Geology*, 287, 93–102.

Pirrie, D., Doyle, P., Marshall, J. D., & Ellis, G. (1995). Cool Cretaceous climates: New data from the Albian of Western Australia. *Journal of the Geological Society*, 152, 739–742.

Polyansky, O. P., Prokopiev, A. V., Koroleva, O. V., Tomsin, M. D., Reverdatto, V. V., Selyatitsky, A. Y., Travin, A. V., et al. (2017). Temporal correlation between dyke swarms and crustal extension in the middle Proterozoic Viluy rift basin, Siberian platform. *Lithos*, 282–283, 45–64.

Polyansky, O. P., Prokopiev, A. V., Koroleva, O. V., Tomsin, M. D., Reverdatto, V. V., Babichev, A. V., Sverdlova, V. G., et al. (2018). The nature of the heat source of mafic magmatism during the formation of the Viluy rift based on the ages of dike swarms and results of numerical modeling. *Russian Geology and Geophysics*, 59(10), 1217–1236. doi: 10.1016/j.rgg.2018.09.003

Prave, A. R., Bates, C. R., Donaldson, C. H., Tolland, H., Condon, D. J., Mark, D., & Raub, T. D. (2016a). Geology and geochronology of the Tana Basin, Ethiopia: LIP volcanism, super eruptions and Eocene-Oligocene environmental change. *Earth and Planetary Science Letters*, 443, 1–8. http://dx.doi.org/10.1016/j.epsl.2016.03.009

Prave, A. R., Condon, D., Hoffmann, K.-H., Tapster, S., & Fallick, A. E. (2016b). Duration and nature of the end-Cryogenian (Marinoan) glaciation. *Geology*, 44, 631–634. doi:10.1130/G38089.1

Pu, J. P., Bowring, S. A., Ramezani, J., Myrow, P., Raub, T. D., Landing, E., Mills, A., et al. (2016). Dodging snowballs: Geochronology of the Gaskiers glaciation and the first appearance of the Ediacaran biota. *Geology*, 44, 955–958.

Puchkov, V., Ernst, R. E., Hamilton, M. A., Söderlund, U., & Sergeeva, N. (2016). A Devonian >2000-km long dolerite swarm-belt and associated basalts along the Urals-Novozemelian fold-belt: Part of an East-European (Baltica) LIP tracing the Tuzo Superswell. *GFF*, 138, 6–16.

Puffer, J. H. (2002). A Late Neoproterozoic eastern Laurentian superplume: Location, size, chemical composition, and environmental impact. *American Journal of Science*, 302, 1–27. http://dx.doi.org/10.2475/ajs.302.1.1

Rampino, M. R., & Caldeira, K. (2017). Correlation of the largest craters, stratigraphic impact signatures, and extinction events over the past 250 Myr. *Geoscience Frontiers*, 8, 1241–1245. doi:10.1016/j.gsf.2017.03.002

Rampino, M. R., & Self, S. (2015). Large Igneous Provinces and biotic extinctions. In H. Sigurdsson (Ed.), *The encyclopedia of volcanoes* (2nd ed.) (pp. 1049–1058). Oxford, Elsevier. doi.org/10.1016/B978-0-12-385938-9.00061-4

Rampino, M. R., Self, S., & Stothers, R. B. (1988). Volcanic winters. *Annual Review of Earth and Planetary Sciences*, 16, 73–99.

Ray, J. S., & Pande, K. (1999). Carbonatite-alkaline magmatism associated with continental flood basalts at stratigraphic boundaries: Cause for mass extinctions. *Geophysical Research Letters*, 26, 1917–1920.

Raymo, M. E. (1991). Geochemical evidence supporting T. C. Chamberlin’s theory of glaciation. *Geology*, 19, 344–347.

Retallack, G. J. (2015). Late Ordovician basalts of Sierra del Tигre, Argentine Precordillera, and the Hirnantian mass extinction. November 2015 LIP of the Month. http://www.largeigneousprovinces.org/15nov

Ricci, J., Quidelleur, X., Pavlov, V., Orlov, S., Shatsillo, A., & Courillot, V. (2013). New 40Ar/39Ar and K-Ar ages of the Viluy traps (Eastern Siberia): Further evidence for a relationship with the Frasnian-Famennian mass extinction. *Palaeeogeoigraphy, Palaeeclimatology, Palaeeoeology*, 386, 531–540.

Robert, B., Greff-Lefftz, M., & Besse, J. (2018). True polar wander: A key indicator for plate configuration and mantle convection during the Late Neoproterozoic. *Geochemistry, Geophysics, Geosystems*, 19, 3478–3495. https://doi.org/10.1029/2018GC007490

Robock, A. (2004). Climatic impact of volcanic emissions. In R. S. J. Sparks & C. J. Hawkeworth (Eds.), *State of the planet* (pp. 125–134). American Geophysical Union, Geophysical Monograph 150, IUGG Volume 19.

Rooney, A. D., Strauss, J. V., Brandon, A. D., & Macdonald, F. A. (2015). A Cryogenian chronology: Two long-lasting, synchronous Neoproterozoic snowball Earth glaciations. *Geology*, 43, 459–462. doi:10.1130/G36511.1

Royer, D. L., Berner, R. A., Montañez, I. P., Tabor, N. J., & Beerling, D. J. (2004). CO2 as a primary driver of Phanerozoic climate. *GSA Today*, 14, 4–10.

Runkel, A. C., Mackey, T. J., Cowan, C. A., & Fox, D. L. (2010). Tropical shoreline ice in the late Cambrian: Implications for Earth’s climate between the Cambrian explosion and the Great Ordovician Biodiversification Event. *GSA Today*, 20(11), 4–10.

Saltzman, M. R., & Young, S. A. (2005). Long-lived glaciation in the Late Ordovician? Isotopic and sequence-stratigraphic evidence from western Laurentia. *Geology*, 33, 109–112.

Schoene, B., Eddy, M. P., Samperton, K. M., Keller, C. B., Keller, G., Adatte, T., & Khadri, S. F. R. (2019). U-Pb con-
strains on pulsed eruption of the Deccan Traps across the end-Cretaceous mass extinction. Science, 363, 862–866.
Schoene, B., Samperton, K. M., Eddy, M. P., Keller, G., Adatte, T., Bowring, S. A., Khadiri, S. F. R., et al. (2015). U-Pb geochronology of the Deccan Traps and relation to the end-Cretaceous mass extinction. Science, 347 (6218), 182–184.
Schrag, D. P., Berner, R. A., Hoffman, F., & Halverson, G. P. (2002). On the initiation of snowball Earth. Geochemistry, Geophysics, Geosystems, 3(6), 1–21. doi:10.1029/2001GC000219
Self, S. (2006). The effects and consequences of very large explosive volcanic eruptions. Philosophical Transactions of the Royal Society A, 364, 2073–2097.
Self, S., & Blake, S. (2008). Consequences of explosive supereruptions. Elements, 4(1), 41–66. doi.org/10.2113/GSELEMENTS.4.1.41
Shellnut, J. G. (2014). The Emeishan large igneous province: A synthesis. Geoscience Frontiers, 5, 369–394.
Shellnut, J. G. (2016). The Early Permian Panjal Traps of the Western Himalaya. Geosciences Canada, 43, 251–264.
Shields-Zhou, G. A., Hill, A. C., & Macgabhanbh, B. A. (2012). The Cryogenian Period. Elsevier Science Limited.
Shumlyanskyy, L., Nosova, A., Billström, K., Söderlund, U., Andréasson, P.-G., & Kuzmenkova, O. (2016). The U-Pb zircon and baddeleyite ages of the Neoproterozoic Volyn Large Igneous Province: Implication for the age of the magmatism and the nature of a crustal contaminant. GFF, 138(1), 17–30.
Stern, R. J., Avigad, D., Miller, N., & Beyth, M. (2008). From volcanic winter to snowball Earth: Alternative explanation for Neoproterozoic biosphere stress. In Y. Dilek, H. Furnes, & K. Muehlenbachs (Eds.), Links between geological processes, microbial activities, and evolution of life (pp. 313–337). Springer Solid Earth Series.
Stoerhers, R. B. (1993). Flood basalts and extinction events. Geophysical Research Letters, 20, 1399–1402.
Strauss, J. V., Rooney, A. D., Macdonald, F. A., Brandon, A. D., & Knoll, A. H. (2014). 740 Ma vase-shaped microfossils from Yukon, Canada: Implications for Neoproterozoic chronology and biostratigraphy. Geology, 42(8), 659–662. doi:10.1130/G35736.1
Svensen, H. H., Corfu, F., Polteau, S., Hammer, Ø., & Planke, S. (2012). Rapid magma emplacement in the Karoo Large Igneous Province. Earth and Planetary Science Letters, 325–326, 1–9.
Svensen, H. H., Hammer, Ø., & Corfu, F. (2015). Astronomically forced cyclicity in the Upper Ordovician and U-Pb ages of interlayered tephra, Oslo Region, Norway. Palaeogeography, Palaeoclimatology, Palaeoecology, 418, 150–159. https://doi.org/10.1016/j.palaeo.2014.11.001
Tegner, C., Andersen, T. B., Kjell, H. J., Brown, E. L., Hagen-Peter, G., Corfu, F., Planke, S., et al. (2019). A mantle plume origin for the Scandinavian dyke complex: A “piercing point” for 615 Ma plate reconstruction of Baltica? Geochimie, Geophysie, Geosystemes, 43(11), 1011–1020. doi:10.1029/2018GC007941
Timmerman, M. J., Heeremans, M., Kirstein, L. A., Larsen, B. T., Spencer-Dunworth, E-A., & Sundvoll, B. (2009). Linking changes in tectonic style with magmatism in northern Europe during the late Carboniferous to latest Permian. Tectonophysics, 473, 375–390.
Tobin, T. S., Bitz, C. M., & Archer, D. (2017). Modeling climatic effects of carbon dioxide emissions from Deccan Traps volcanic eruptions around the Cretaceous-Paleogene boundary. Palaeogeography, Palaeoclimatology, Palaeoecology, 478, 139–148.
Torsvik, T. H., Smethurst, M. A., Burke, K., & Steinberger, B. (2008). Long term stability in deep mantle structure: Evidence from the >300 Ma Skagerrak-centered Large Igneous Province (the SCLIP). Earth and Planetary Science Letters, 267, 444–452.
Turgeon, S. C., & Creaser, R. A. (2008). Cretaceous oceanic anoxic event 2 triggered by a massive magmatic episode. Nature, 454, 323–326.
Tyrrell, T. (1999). The relative influences of nitrogen and phosphorus on oceanic primary production. Nature, 400, 525–531.
Ukstins, I. A., Renne, P. R., Wolfenden, E., Baker, J., Ayalew, D., & Menzies, M. (2002). Matching conjugate volcanic rifted margins: 40Ar/39Ar chrono-stratigraphy of pre- and syn-rift bimodal flood volcanism in Ethiopia and Yemen. Earth and Planetary Science Letters, 198, 289–306.
Ukstins Peate, I. U., Baker, J. A., Kent, A. J., Al-Kadasi, M., Al-Subbary, A., Ayalew, D., & Menzies, M. (2003). Correlation of Indian Ocean tephra to individual Oligocene silicic eruptions from Afro-Arabian flood volcanism. Earth and Planetary Science Letters, 21, 311–327.
Upton, B. G. J., Stephenson, D., Smedley, P. M., Wallis, S. M., & Fitton, J. G. (2004). Carboniferous and Permian magmatism in Scotland. In M. Wilson, E.-R. Neumann, G. R. Davies, M. J. Timmerman, M. Heeremans, & T. G. Larsen (Eds.), Permo-Carboniferous magmatism and rifting in Europe (pp. 195–218). Geological Society, London, Special Publications 223.
Veivers, J. J., & Powell, M. (1987). Late Paleozoic glacial episodes in Gondwanaland reflected in transgressive-regressive depositional sequences in Euramerica. Geological Society of America Bulletin, 98, 475–487.
Veizer, J., Godderis, Y., & François, L. M. (2000). Evidence for decoupling of atmospheric CO₂ and global climate during the Phanerozoic eon. Nature, 408, 698.
White, A. F., & Brantley, S. L. (1995). Chemical weathering rates of silicate minerals: An overview. Reviews in Mineralogy and Geochemistry, 31, 1–22.
Wierzbowski, H. (2015). Seawater temperatures and carbon isotope variations in central European basins at the Middle-Late Jurassic transition (Late Callovian–Early Kimmeridgian). Palaeogeography, Palaeoclimatology, Palaeoecology, 440, 506–523.
Wignall, P. B. (2001). Large igneous provinces and mass extinctions. Earth-Science Reviews, 53, 1–33.
Wignall, P. B. (2005). The link between large igneous province eruptions and mass extinctions. Elements, 1, 293–297.
Willeit, M., & Ganopolski, A. (2018). The importance of snow albedo for ice sheet evolution over the last glacial cycle. Climate of the Past, 14, 697–707. https://doi.org/10.5194/ cp-14-697-2018
Wilson, M., Neumann, E.-R., Davies, G. R., Timmerman, M. J., Heeremans, M., & Larsen, N. T. (2004). Permo-Carboniferous magmatism and rifting in Europe:
Introduction. In M. Wilson, E.-R. Neumann, G. R. Davies, M. J. Timmerman, M. Heeremans, & B. T. Larsen (Eds.), Permo-Carboniferous magmatism and rifting in Europe (pp. 1–10). Geological Society of London, Special Publications, 223.

Xu, Y.-G., Wei, X., Luo, Z.-Y., Liu, H.-Q., & Cao, J. (2014). The Early Permian Tarim Large Igneous Province: Main characteristics and a plume incubation model. *Lithos, 204*, 20–35.

Youbi, N., Aarab, E. M., Ait Lahna, A., Tassinari, C. C. G., Ernst, R., Söderlund, U., Gerdes, A., et al. (2017). The 348–340 Ma Jebilet-Rehamna-Fourhal Large Igneous Province of the Meseta domain (Variscan Belt, Morocco): U-Pb geochronology, geochemistry. 2nd Colloquium of the International Geoscience Program IGCP 638, Casablanca, Morocco, 7–12 November 2017.

Youbi, N., Ernst, R. E., Söderlund, U., Boumehdi, M. A., & Ait Lahna, A. (2018). Did the Central Iapetus Magmatic Province (CIMP) both trigger and end the c. 580 Ma Gaskiers glaciation? Goldschmidt/Goldschmidt 2018, Abstract, Boston.

Youbi, N., Ernst, R. E., Söderlund, U., Boumehdi, M. A., Ait Lahna, A., Tassinari, C. C. G., El Moume, W., et al. (2020). The Central Iapetus Magmatic Province (CIMP): An updated review and link with the c. 580 Ma Gaskiers glaciation. In T. Adatte, G. Keller, & D. Bond (Eds.), *New developments on volcanism, impacts and mass extinctions*. Geological Society of America Special Paper 544 (in press).

Youbi, N., Ernst, R. E., Söderlund, U., Boumehdi, M. A., Bensalah, M. K., Mata, J., Madeira, J., et al. (2019). *Multiple LIP pulses of the Central Iapetus Magmatic Province (CIMP) and link with glaciations during the Ediacaran-Cambrian*. Goldschmidt 2019, Abstract, Barcelona.

Youbi, N., Ernst, R. E., Söderlund, U., Soulaimani, A., Doblas, M., Bertrand, H., Marzoli, A., et al. (2011). The Central Iapetus Magmatic Province (CIMP) Large Igneous Province: Distribution, nature, origin, and environmental impact. The American Association of Petroleum Geologists (AAPG) Search and Discovery. Article #90137.

Young, G. M. (2013). Evolution of Earth’s climatic system: Evidence from ice ages, isotopes, and impacts. *GSA Today, 23* (10), 4–10. http://dx.doi.org/10.1130/GSATG183A.1

Zachos, J., Pagani, M., Sloan, L., Thomas, E., & Billups, K. (2001). Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science, 292* (5517), 686–693.

Zhai, Q.-G., Jahn, B.-M., Su, L., Ernst, R. E., Wang, K.-L., Zhang, R.-Y., Wang, J., et al. (2013). SHRIMP zircon U-Pb geochronology, geochemistry and Sr-Nd-Hf isotopic compositions of a mafic dyke swarm in the Qiangtang terrane, northern Tibet and geodynamic implications. *Lithos, 174*, 28–43.

Zhang, S. H., Ernst R. E., Pej, J. L., Zhao, Y., Zhou, M. F., & Hu, G. H. (2018). A temporal and causal link between ~1380 Ma large igneous province and black shales: Implications for the Mesoproterozoic time scale and paleoenvironment. *Geology, 46*, 963–966.

Zhao, J.-H., & Asimow, P. D. (2018). Formation and evolution of a magmatic system in a rifting continental margin: Neoproterozoic arc- and MORB-like dike swarms in South China. *Journal of Petrology, 59*, 1811–1844.

Zhu, D.-C., Chung, S.-L., Mo, X.-X., Zhao, Z.-D., Niu, Y.-L., Song, B., & Yang, Y.-H. (2009). The 132 Ma Comei-Bunbury large igneous province: Remnants identified in present-day southeastern Tibet and southwestern Australia. *Geology, 37*, 583–586.