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Bryan, Scott (2021) Environmental Impact of Silicic Magmatism in Large Igneous Province Events. In Ernst, Richard R., Dickson, Alex, & Bekker, Andrey (Eds.) Large Igneous Provinces: A Driver of Global Environmental and Biotic Changes. American Geophysical Union, Hoboken, pp. 133-151.

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https://doi.org/10.1002/9781119507444.ch6
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Environmental Impact of Silicic Magmatism in Large Igneous Province Events

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ABSTRACT

Silicic magmatism is a feature of all continental LIP events, and where volumetrically significant, occurs as high-frequency (~1,000–10,000 yr recurrence intervals), large-magnitude (>M8) explosive supereruptions producing vast ignimbrite sheets. Silicic supereruptions inherently have the eruptive mechanism to deliver aerosols and ash to the stratosphere for global dispersal, and thus overcome eruptive barriers that exist for flood basalts built up by long-lived, low effusion and low vigor fountains that lack height and persistent stratospheric penetration. The historical record demonstrates the climate forcing capabilities of silicic supereruptions, which during LIP events, were likely associated with large CO₂, SO₂, halogen, and Hg emissions, and through tephra deposition, could cause iron fertilization in the world’s oceans, thereby kick-starting phytoplanktonic biological pumps to significantly draw down atmospheric CO₂. What may be important, therefore, for LIP events to cause the most environmental impact and trigger a mass extinction, is the combined effect of closely spaced basaltic and silicic, or effusive and explosive, eruptions that work in tandem to overload the troposphere and stratosphere with volcanic aerosols producing rapid decadal-scale, extreme fluctuations in pH driven by acid rain, S-, or iron fertilization-driven temperature chills, and toxic UV radiation bursts. These effects could be repeated within as little as a few hundred years of each other particularly during hyperactive LIP pulses.

6.1. INTRODUCTION

Large igneous province (LIP) events and large asteroid impacts are the two most plausible drivers for global environmental impact to cause mass extinctions. The nature of past mass extinctions requires that their drivers must occur relatively episodically, be rapid events with global reach leading to extinction of a significant part of all life across the planet where rapid and major global environmental change forced the apparently indiscriminate loss of life. LIPs and asteroid impacts are consistent with this in that they are both exceptional, episodic, and geologically rapid events that at their extremes can cause global environmental damage.

A number of factors make LIPs a particularly attractive trigger to cause mass extinctions. First, LIPs have a temporal coincidence with all known mass extinction events (e.g., Courtillot & Renne, 2003; Bond & Wignall, 2014; Ernst, 2014). Second, LIPs are unheralded episodic igneous events recording exceptional volumes and rates of magma production and coupled aerosol budgets (Bryan & Ernst, 2008; Bryan & Ferrari, 2013). Continental LIPs (Fig. 6.1) are the only known loci for both basaltic and silicic supereruptions on Earth (Bryan et al., 2010). The total volume of magma intruded and released during the main igneous pulses is 10⁶–10⁷ km³ and where the volume of magma emitted during individual eruptions is typically 10²–10⁴ km³. Consequently, the huge cumulative volume of LIPs most likely requires hundreds of supereruptions occurring at high frequency where each can erupt up to 10,000 km³ of magma (Bryan et al., 2010). Third, because of the relative longevity of the main
Figure 6.1 Global distribution of Phanerozoic continental large igneous provinces (LIPs). LIPs are differentiated into those that are dominated by flood basalts produced by effusive eruptions; mafic LIPs with volumetrically significant volumes of silicic igneous rocks (>10^4 km^3; Bryan et al., 2002); LIPs with volumetrically significant volumes of mafic volcaniclastic deposits (>10^4 km^3) reflective of explosive phreatomagmatic eruptive styles (e.g., Ross et al., 2005); and the silicic LIPs, which are volumetrically dominated by silicic ignimbrite (>2.5 x 10^5 km^3; Bryan et al., 2002; Bryan, 2007; Bryan & Ferrari, 2013). Annotated ages denote the onset of the main phase or first pulse of magmatism to the LIP event. The inferred extent of some of the Paleozoic LIP events is shown by a dashed line, due to poor exposure and limited study. CAMP = Central Atlantic magmatic province; HALIP = High Arctic large igneous province; NAIP = North Atlantic igneous province; R-S = Rajmahal-Sylhet Traps; SRP = Snake River Plain; KCA = Kennedy-Connors-Auburn. Figure is updated and modified from Bryan and Ernst (2008) and Bryan and Ferrari (2013).
erupted volume) and the Grande Ronde basalts only (150,100 km³, 72.3%). Kasbohm and Schoene (2018) noted that eruption rates, both for the main phase of the Columbia River LIP (Steens, Imnaha, and Grande Ronde basalts [GRB], 92% of total volume), are at least two orders of magnitude greater than the Siberian Traps (~300–800,000 km³, 72.3%). Kasbohm and Schoene (2018) noted that eruption rates were likely briefly higher during the main phase of Columbia River with pulses >1,000 km³/Kyr. The Deccan and CAMP were temporarily related with the Cretaceous-Paleogene and end-Triassic mass extinction events, respectively. Two different estimates are available for phase 2 of the Deccan, but both yield averaged eruption rates >1,000 km³/Kyr. A total province volume of 4 x 10⁵ km³ (Vasiliev et al., 2000) is assumed for the Siberian Traps, where two-thirds of the eruptive volume is estimated to have been erupted in ~300 Kyr prior to the mass extinction event; overall duration of the Siberian Traps main phase is estimated to be ~800 Kyr.

**Table 6.1 Summary of LIPs Where Detailed Volume-Duration Constraints Are Available**

| LIP            | Duration of main pulse (Kyr) | Volume of main pulse (km³) | Averaged eruption rate (km³/Kyr) | References                                      |
|----------------|------------------------------|----------------------------|----------------------------------|------------------------------------------------|
| Columbia River | ~750                         | 192,900                    | 334                              | Kasbohm & Schoene (2018)                        |
| GRB only:      | ~420                         | 150,100                    | 375                              | Barry et al. (2010, 2013)                       |
| Deccan (phase 2)| ~500                         | ~800,000                   | 1,600                            | Chenet et al. (2009)                            |
|                | ~100                         | ~253,000                   | 2,530                            | Courtillot & Fluteau (2014)                     |
| CAMP           | ~600                         | >800,000                   | 1,333                            | Blackburn et al. (2013)                         |
| Siberian Traps | ~300                         | 2,600,000                  | 8,888                            | Burgess & Bowring (2015)                        |
|                | ~800                         | 4,000,000                  | 5,000                            |                                                 |

Note: Note the nearly order of magnitude higher averaged eruption rate for the Deccan and Central Atlantic Magmatic Province (CAMP) and particularly the Siberian Traps compared with the youngest LIP, the Columbia River Flood Basalt Province. Two different approaches to constrain eruption ages for Columbia River (U-Pb zircon dating of interbedded silicic tuffs, Kasbohm & Schoene, 2018; ⁴⁰Ar/³⁹Ar dating of whole-rock fresh matrix, Barry et al., 2010) yield similar averaged eruption rates, both for the main phase of the Columbia River LIP (Steens, Imnaha, and Grande Ronde basalts [GRB], 92% of total erupted volume) and the Grande Ronde basalts only (150,100 km³, 72.3%). Kasbohm and Schoene (2018) noted that eruption rates were likely briefly higher during the main phase of Columbia River with pulses >1,000 km³/Kyr. The Deccan and CAMP were temporarily related with the Cretaceous-Paleogene and end-Triassic mass extinction events, respectively. Two different estimates are available for phase 2 of the Deccan, but both yield averaged eruption rates >1,000 km³/Kyr. A total province volume of 4 x 10⁵ km³ (Vasiliev et al., 2000) is assumed for the Siberian Traps, where two-thirds of the eruptive volume is estimated to have been erupted in ~300 Kyr prior to the mass extinction event; overall duration of the Siberian Traps main phase is estimated to be ~800 Kyr.

Igneous pulses (10⁵–10⁹ yr), LIPs, through their aerosol emissions, represent a sustained assault on the global atmosphere, hydrosphere, and biosphere. It is not just one eruption that does the damage, but likely the sustained effect of multiple, closely spaced supereruptions happening every few hundred years or less that maintain environmental stress, prevent full recovery, and culminate in biological turnover. A number of recent studies are now indicating that some LIPs may have particularly hyperactive pulses that last as little as 100–500 Kyr during which a substantial igneous volume is emplaced (Table 6.1). This suggests averaged eruption rates of 10⁵–10⁹ km³ of magma every 1,000 yr (Table 6.1). In contrast, our understanding of large asteroid impacts associated with mass extinctions is that they are single impact events requiring the asteroid to be of sufficient size (>10 km diameter) to be able to cause rapid global devastation in one hit (e.g., Schulte et al., 2010; Kring, 2007; Rumpf et al., 2017). Even for large asteroids, the impact effects are predicted to last only years to perhaps a couple of decades, before the Earth returns to preimpact conditions (Kring, 2000). This remains an important point of difference for how LIPs and asteroid impacts could cause mass extinctions.

LIPs are understood to drive rapid environmental change in three main ways: (1) by changing global surface temperatures through aerosol addition to the atmosphere; (2) by volcanic aerosols attacking atmospheric chemistry, in particular, removing ozone; and (3) by causing environmental toxicity through the introduction of heavy metals (e.g., Hg) and/or causing acid rain. The relevance and impact of each is subtly dependent on the eruptive style, magma compositions, and eruptive setting where additional aerosol additions can come from magma interactions with surrounding country rock (e.g., Svensen et al., 2004, 2007, 2009; Black et al., 2014; Reynolds et al., 2017) and degassing of intruded magma and underplate (Bryan et al., 2012; Armstrong McKay et al., 2014). Because LIPs are basalt-dominated (Collin & Eldholm, 1994; Bryan & Ernst, 2008), much emphasis has been given to basalt-related aerosol emissions and effects. Based on historic and directly observed basaltic eruptions, high CO₂ and SO₂ emissions can occur (Gerlach, 1991; Burton et al., 2010; D’Aleo et al., 2016). Of the cocktail of gases emitted, CO₂ has gained precedence chemically reactive aerosols can quickly be flushed out of the atmosphere by rain if only delivered to the troposphere.
CO₂ is excluded here because, as with present-day anthropogenic CO₂ emissions, volcanic CO₂ emissions can load and be retained in the troposphere to generate greenhouse effects as well as acid rain (Black et al., 2014). One problem then is that the flood basalt lavas (the main building block of LIPs) are dominantly pahoehoe lava fields and analogous to modern-day pahoehoe eruptions such as at Iceland and Hawaii (e.g., Self et al., 1996; 1997; Thordarsson & Self, 1998; White et al., 2009). These flood basaltic lava fields are mainly built up by long-lived low effusion and low vigor fountains that lack height and persistent stratospheric penetration. In an excellent, detailed study by Brown et al. (2014) of the near-vent deposits to the ~1,300 km³ Rosa lava flow of the Columbia River basalts, a series of near-vent pyroclastic cones were documented along the fissure, spaced on average every 1 to 2 km. The pyroclastic cones do provide evidence for repeated and more vigorous fountaining in which convective gas-rich plumes likely reached altitudes of 10 km or more during which SO₂ emissions could have been three times the emissions from the 1991 Pinatubo eruption per day (Glaze et al., 2017). However, the bulk of the Rosa lava (as with all flood basalt lavas) must have been emplaced by low-fountains and essentially lava effusion, and when tallied up, the pyroclastic cones represent <3% total duration of a decadal eruption. For the majority of their life then, flood basalt eruptions do not appear to generate persistent stratospheric penetrating eruption columns. A solution to overcome this eruptive barrier for mafic magmas is for eruptions to become explosive and phreatomagmatic due to magma interaction with surface or ground waters (e.g., Ross et al., 2005; Uktstins Peate & Bryan, 2008). This appears to have been particularly important for the end-Permain Siberian Traps (Black et al., 2012, 2014; but cf. Jerram et al., 2016a), Emeishan (Uktstins Peate & Bryan, 2008; Jerram et al., 2016b), and North Atlantic Igneous Province (Uktstins Peate et al., 2003). Further potential explosive venting from large hydrothermal vent structures seems to be an increasing observation in a number of LIPs (e.g., Svensen et al., 2009; Reynolds et al., 2017), providing significant additional gas fluxes from these vents, however, it remains unclear how high these are able to penetrate the atmosphere.

Silicic igneous rocks are associated with all continental flood basalt provinces and volcanic rifted margins, where the silicic volcanic rocks can form substantial parts of the eruptive stratigraphy and have eruptive volumes >10⁶ km³ (Bryan et al., 2002). It has also been recognized that some silicic igneous provinces meet the criteria of a LIP (Bryan & Ernst, 2008), but have low proportions of basalt expressed at the surface, and these have been referred to as silicic LIPs (Bryan et al., 2002; Bryan, 2007; Bryan & Ferrari, 2013). Silicic volcanism associated with LIPs is generally explosive as ignimbrites are the dominant, large-volume eruptive products preserved within continental LIPs (Bryan et al., 2002, 2010; White et al., 2009). Consequently, it is the silicic eruptions in LIP events that inherently have the eruptive mechanism to deliver ash and aerosols to the stratosphere for global dispersal. This contribution therefore focuses on the role of silicic volcanism within LIP events and their contribution to causing global environmental impact and biological turnover. Silicic volcanism is an underappreciated component in LIPs (Bryan et al., 2002), and the lower frequency and lack of observation of large explosive silicic eruptions in historic times compared with basaltic eruptions hinders a full understanding of their environmental impacts. This chapter begins by reviewing the observed impacts from small-scale historic silicic explosive eruptions to provide insight into the diversity of impacts from past volcanic eruptions, and where the scale of impact such as during LIP events can be confidently scaled up with eruption magnitude. The effects on humans can also be used as a proxy for understanding impacts on past animal life during LIP events. I then examine the likely impacts from large magnitude basaltic and silicic eruptions, which coexist during the main eruptive pulses of continental LIPs. These impacts are considered at two spatial scales where direct impacts to the environment occur at the provincial scale and indirect impacts at the global scale.

6.2. ENVIRONMENTAL IMPACTS OF HISTORICAL SILICIC VOLCANISM

On average, every week globally, approximately 20 subaerial volcanoes show some sign of unrest and potentially posing a hazard to surrounding communities (Global Volcanism Program, Smithsonian Institution: https://volcano.si.edu/). Most present-day activity, however, reflects small-scale eruptions at active plate boundary settings, about 75% of all known eruptions since 10 ka are VEI 2 or less (Volcanic Explosivity Index, <0.01 km³ of erupted magma; Newhall & Self, 1982), and so our direct observation and experience of the large magnitude (>100 km³ or >VEI 7) basaltic and silicic supereruptions that characterize LIP events (Bryan et al., 2010) are nonexistent. Of the present-day volcanic activity, it is the more infrequent silicic explosive eruptions that pose the greatest hazards and environmental impact. At least 70% of volcanic-related fatalities since 1600 AD have been directly related to explosive eruptions and in particular to pyroclastic density currents, whereas lava flows have accounted for only 0.32% of fatalities over the same period (Auker et al., 2013). Further, five explosive eruptions during this period account for 58% of all known fatalities (1792 Unzen, 1815 Tambora, 1883 Krakatau, 1902 Mount Pelée, and 1985 Nevado del Ruiz) reinforcing that the number of fatalities is strongly correlated with eruption magnitude (Auker et al., 2013).
6.2.1. Impacts From Pyroclastic Density Currents

It is the range of hazards that can be generated from a single explosive eruption that make them so deadly, and more impactful than effusive lava-producing eruptions. Pyroclastic density currents are the most destructive volcanic phenomenon. Their hazardous and destructive nature is because pyroclastic density currents (1) can travel hundreds m/s across the landscape, extending to >100 km from the vent covering huge areas (10^4–10^5 km^2), potentially in less than an hour; (2) are hot flows (>100 to ~1000°C) as evident from carbonized tree logs and welding of juvenile clasts and ash matrix; (3) can leave deposits tens to hundreds of meters thick burying the landscape, and the highly mobile nature of the density currents also is reflected by deposits observed on mountains >500 m high over 60 km from source; and (4) can greatly expand the area of ash fallout to areas beyond the limits of pyroclastic density current run-out (Freundt et al., 2000). Tephra fallout and lahars are additional fatal hazards commonly occurring during explosive eruptions. When volcanoes occur in or near oceans, tsunamis can also be generated by pyroclastic density currents entering water (e.g., Cas & Wright, 1991), and caldera collapse or landslides triggered by flank collapse (e.g., flank collapse of Anak Krakatau and associated tsunami, 22 December 2018).

6.2.2. Impacts From Volcanic Ash

Given its ability to be globally dispersed, ash fall has the greatest potential to directly or indirectly affect the largest number of people, and by proxy, life, worldwide (Simkin et al., 2001, Witham, 2005). Although ash falls rarely endanger human life directly, threats to public health and disruption to critical infrastructure services, aviation, and primary production have at times led to substantial societal impacts and costs, even when deposit thicknesses may be only a few millimeters (Jenkins et al., 2015). This leads on to an important observation from the study of volcano-related fatality records since 1600 AD that secondary or indirect health effects such as famine and disease account for significant numbers (24%) of fatalities (Auker et al., 2013). Population health can be weakened by an eruption (even when of relatively small magnitude) and the effects from an eruption can persist for a considerable time affecting recovery (e.g., restocking of food resources), leading to further loss of life.

Volcanic ash presents a number of impacts on vegetation. Wind-blown ash can “ash-blast” vegetation and cause challenging living conditions for decades after the eruption (e.g., Hudson 1991 eruption; Wilson et al., 2011). Ash deposits have been observed to form surface crusts that can inhibit water infiltration and soil-gas exchange (i.e., water, oxygen, and nitrogen cycles) reducing soil fertility as well as increase surface runoff (Blong, 1984). In particular, survival of agricultural crops/pasture is often severely limited when ash thickness is >10–15 cm (Jenkins et al., 2015). Like ash accumulation on buildings that can cause roofs to collapse, ash accumulation on plants can also lead to defoliation and breakage, and is exacerbated when the ash becomes saturated and heavy. Ash coatings to leaves will inhibit photosynthesis. An additional effect is where readily soluble salts and acids adhered to ash surfaces (mainly HCl, HF, and H_2SO_4, initially occurring as volcanic aerosols) are leached off by rain to damage foliage or fruit; specifically, the protective waxy leaf surface coatings can be altered lowering disease resistance and it may inhibit plant germination and reproduction (Wilson et al., 2007).

The most commonly reported public health effects of ash exposure following explosive eruptions are irritation of the eyes and upper airways and exacerbation of preexisting respiratory conditions, such as asthma (Gudmundsson, 2011; Jenkins et al., 2015). Both acute (e.g., asthma, bronchitis) and chronic health effects of volcanic ash can occur, and depend upon particle size (particularly the proportion of respirable-sized material), mineralogical composition (including the crystalline silica content), and the physicochemical properties of the surfaces of the ash particles (Horwell & Baxter, 2006). Where there is prolonged exposure to ash, an increased risk of chronic silicosis can occur, which results in lung impairment and scarring, from inhalation of the abrasive ash particles. Penetration of ash particles into respiratory tracts is dependent on particle size: larger particles (>10 μm diameter) lodge in the upper airways, while those in the 4 to 10 μm size range deposit in the trachea and bronchial tubes, and very fine (< 4 μm diameter) particles may penetrate deeper into the lungs (Jenkins et al., 2015).

Significant effects on animals/livestock have also been observed. Animals have been reported to being put off their food due to tephra contamination, as well as suffering weight-loss from eating ash-covered fodder (Wilson et al., 2007). Grazing animals are particularly exposed to the effects of ash where ash accumulations may lodge in either the respiratory or gastrointestinal tracts through inhalation or digestion. Thicker ash falls can smother feed and restrict access to drinking water, leading to starvation and/or dehydration (Jenkins et al., 2015). However, acute and chronic fluorosis of grazing animals appears more of an issue following a number of explosive eruptions (e.g., Lonquimay, Chile, Araya et al., 1990; 1995 Mount Ruapehu; Cronin et al., 2003), and has been a particular problem in Iceland (Sigurdsson & Paulsson, 1957; Thorarinsson, 1979; Oskarsson, 1980). Of note is that a number of studies have found fluoride levels are
highest for the smallest (and more easily ingested) ash particle sizes, which will be most distal to the volcano (Rubin et al., 1994; Oskarsson, 1980).

6.2.3. Impacts From Pumice Rafts

In the marine realm, pumice rafts can cause problems for shallow marine ecosystems (Bryan et al., 2004). Observations following the 1984 Home Reef eruption in Tonga indicated that a large kill of deep-sea fish followed the arrival of pumice rafts in the Lau Island Group (Smithsonian Institution, 1984), while pumice rafts asphyxiated marine life when stranded in shallow water, and large fish jumping onto the pumice were apparently unable to penetrate through it and back to safety (Ryan, 1986). Pumice rafts appear to also have an impact on seabirds. Seabirds are susceptible to debris ingestion, commonly mistaking floating debris like pumice for food (Roman et al., 2019a, b). In the spring of 2013, a large wreck (caused by a combination of exhaustion and starvation that affects birds during migration) involving mostly short-tailed shearwaters (Puffinus tenuirostris) occurred along the east Australian coastline (Springer et al., 2018; Roman, 2018). It was discovered that a higher-than-usual ingestion of naturally occurring pumice characterized the wreck. The ingested pumice was sourced from pumice rafts produced by the 2012 Havre eruption in the Kermadec arc (Carey et al., 2014, 2018) that had drifted into eastern Australian and Coral Sea waters during the winter-spring of 2013.

6.3. ENVIRONMENTAL IMPACTS OF LIP MAGMATISM

All of the impacts experienced during small-scale historic explosive volcanic eruptions can be expected to have occurred during LIP eruptions but be strongly scaled up. The historic record shows that silicic explosive eruptions have been more impactful on global climate (e.g., 1815 Tambora, 1883 Krakatau, and 1991 Mount Pinatubo) and caused more fatalities (Auker et al., 2013). Our current understanding of LIP magmatism identifies environmental impact in two main ways and at two spatial and temporal scales: at the provincial scale by directly impacting the landscape through burial under ultimately, kilometers of volcanic rock, and indirectly at the global scale through associated aerosol emissions and ash dispersal. Both mechanisms result in impacts at short (immediate to years) and longer term (>1 Kyr) time scales, with long-term impacts accentuated by the cumulative effects from repeated high-frequency eruptions over 10⁵–10⁶ yr. The following discussion focuses on impacts from continental LIPs and this includes impacts from both flood basaltic and rhyolitic eruptions.

6.3.1. Provincial-Scale Impacts on the Landscape

The largest individual basalt and rhyolite eruptions recorded from LIPs cover areas in excess of 10⁵ km², each being typically a few tens of meters thick, and collectively build provinces with areal extents that often exceed 10⁶ km² (Bryan & Erns, 2008; Bryan et al., 2010). It has been estimated that the largest flood basalts may take decades to erupt and emplace, while the largest silicic ignimbrites may be emplaced in only a few weeks (Self et al., 1997; Bryan et al., 2010). Consequently, one LIP supereruption will lead to immediate landscape modification at a regional scale, burying preexisting drainage systems and destroying all life within that depositional area. The land surface becomes a bare rock pavement. Drainage systems need to be reestablished following eruption and potentially new environments are developed such as freshwater lakes as a result of water ponding in topographic depressions developed on the upper surfaces of the flood volcanics (e.g., sedimentary interbeds of the Ellensburg and Latah formations within the Columbia River Basalt Group; Burns et al., 2011). Alternatively, high permeabilities of lava flow fields arising from jointing can lead to large aquifer systems (White et al., 2009). Due to the high frequency of LIP eruptions, these new land surfaces and sedimentary environments are typically short lived and become buried by the next flood basalt or ignimbrite. As another example, the lower stratigraphy of Paraná-Etendeka flood basin province reveals an active aeolian sand sea quickly becoming buried by flood basaltic lavas, and aeolian processes being terminated over a wide area (Jerram et al., 1999).

Wildfires

Wildfires appear to be a major feature of mass extinction events, and provide a thermogenic mechanism to increase CO₂ and CH₄ loading of the atmosphere during LIP volcanism. Increased wildfire activity associated with the CAMP event is indicated by charcoal records from Greenland, Denmark, Sweden, and Poland (Lindström et al., 2015). Widespread distributions of fire-derived products have been identified for the Permian-Triassic extinction event (Shen et al., 2011, and references therein), and an enrichment in soot in Cretaceous-Tertiary rocks has been known for some time (Wolbach et al., 1985, 1988, 1990).

Any vegetation within the emplacement area of a LIP eruption is destroyed. Where forests existed, forest fires will result and these have the potential to extend beyond the limits of the lava or ignimbrite. Excellent examples of carbonized tree logs are preserved within the Taupo ignimbrite, recording the destructive effect of pyroclastic density currents on forests (Walker, G., et al., 1981). At the margins of ignimbrite sheets, the pyroclastic density
currents may still have enough force and heat to blow down and char trees as was observed with the initial blast of the 1980 Mount St. Helens eruption (Moore & Sisson, 1981). Flood basalts lavas will also burn forests as is regularly observed in eruptions at Kilauea. Evidence for methane generation from the burning of vegetation and ignition from the heat of the advancing lava flows comes from the images and footage of blue burning flames during the 2018 lower Puna eruption along the east rift zone of Kilauea (USGS, 2018a, b).

Volcanic lightning strikes during an eruption have the potential to trigger additional wildfires as well as cause fatalities by direct ground strikes (McNutt & Thomas, 2015). Volcanic lightning results from static electrical charges within volcanic ash plumes and has been recently observed in pyroclastic eruption columns (e.g., 2008 Chaiten eruption), phreatomagmatic ash plumes (2010 Eyjafjallajökull and 2011 Grimsvötn eruptions), and ash plumes above strombolian fountains (2015 Mount Etna). These examples indicate that volcanic lightning and any ground strikes to trigger a wildfire would be localized around the vent, and this would particularly be the case for flood basalts eruptions. However, volcanic lightning can also be developed within co-ignimbrite ash plumes and these can occur at substantial distances (tens to hundreds km) from the vent, providing more scope for numerous wildfires to be triggered by the silicic explosive eruptions in LIP events.

**Weathering of LIP Volcanic Successions**

Silicate (chemical) weathering is thought to modulate climate over geologic timescales through the sequestering of atmospheric CO₂ (e.g., Walker, J., et al., 1981; Brady, 1991). This is evident, for example, from the long-term global cooling trend since the Early Eocene (e.g., Zachos et al., 2008) where Andean and Himalayan orogenesis has accelerated silicate weathering and by interpretation, drawdown of atmospheric CO₂ (Ruddiman et al., 1997). Consideration has been given to the weathering of flood basalt provinces as another mechanism to accelerate atmospheric CO₂ drawdown and thereby drive atmospheric cooling given that basaltic rocks tend to weather at higher rates than other silicate rocks (Goddéris et al., 2003; Dessert et al., 2003; Ernst & Youbi, 2017; see also Park et al., Chapter 7 this volume). Weathering is optimized under warmer and wetter conditions at low latitudes.

There is generally little evidence for large-scale and intense weathering occurring during the emplacement of flood basalt provinces. Well-developed, thick, and extensive paleosols recording weathering are uncommon to rare in LIPs, and, often, lava flow fields and silicic ignimbrites occur in direct contact with each other, consistent with little time break between LIP eruptions. Eruption frequency is a key factor in modulating surface weathering (e.g., Renne et al., 2015). The high frequency of eruption, with >M8 eruptions potentially occurring every 4,000 yr or less during the main LIP pulse (Bryan et al., 2010), minimizes the time available for exposure and weathering of the upper surfaces of flood volcanics. Instead, intercalated sedimentary deposits tend to be a more common feature within flood volcanic piles (e.g., see Fig. 1 of Bryan et al., 2010). The CAMP, Karoo, and Paraná-Etendeka flood basalt provinces (some of the first to occur during Pangea breakup) were located in the interior of the Pangea supercontinent where arid conditions prevailed over large tracts of the supercontinent and landscape relief and elevation were flat and low, respectively. In particular, the Paraná-Etendeka flood basalt province was initially emplaced in a very arid environment, as evident by intercalated aeolian sandstones in the lower part of the stratigraphy (Jerram et al., 1999). Other LIPs like the Ferrar or High Arctic were emplaced in circum-polar regions where weathering rates would also have been much reduced. Weathered basaltic sediment, or “boles” (Wilkins, 1994; Widdowson et al., 1997; Srivastava et al., 2015), are a distinctive feature of the Deccan LIP that was located ~15–30°S at the time of emplacement around the K-T boundary, and have commonly been interpreted as paleosols (e.g., Widdowson et al., 1997; Ghosh et al., 2006; Sayyed, 2014). However, intense weathering and pedogenesis are not recorded in all boles (Srivastava et al., 2015) and a significant number are altered basaltic ash fall deposits (Widdowson & Sumner, 2008), or contain contributions from distally sourced silicic tephras (Schoene et al., 2015). In summary, geologic evidence is lacking for substantial chemical weathering of flood volcanic piles during LIP events as a viable mechanism to cause rapid and global environmental impact through atmospheric cooling.

### 6.3.2. Global Environmental Impacts

While there are significant environmental impacts at the site of the LIP resulting from the eruption and emplacement of catastrophic volumes of magma, it is not the igneous volume that is necessarily critical, but the volcanic aerosols that are emitted during eruptions or via degassing of shallow intrusions and underplated magma that are critical in the debate about what causes mass extinctions. Although the data are still in their infancy, they reveal the substantial total aerosol budgets associated with these provinces, which dwarf element budgets in our present-day atmosphere (Table 6.2). LIPs appear to differ in terms of their halogen budgets where there may be additional contributions from thermal and metamorphic driven degassing of crustal rocks as well as degassing of igneous intrusions (Table 6.2). The following sections
examine the way silicic eruptions may have contributed to a number of widely identified kill mechanisms connected to LIP events (e.g., Bond & Wignall, 2014; Ernst & Youbi, 2017).

**Sulfur and Halogen Emissions**

The main volatile species emitted during LIP eruptions are water, CO$_2$ (both of which are greenhouse gases), SO$_2$ that can lead to atmospheric cooling, and then a number of halogen species (F, Cl, and Br) that are particularly harsh to stratospheric ozone (see also Mather & Schmidt, Chapter 4 this volume). Environmental toxicity is being increasingly identified as an important kill mechanism that may have global expression (e.g., Sanei et al., 2012; Sial et al., 2013, 2014; Grasby et al., 2015, 2016; Percival et al., 2015; Font et al., 2016; Thibodeau et al., 2016). The high volcanic CO$_2$ and SO$_2$ emissions may combine in the atmosphere to produce acid rain in the form of carbonic acid and sulphuric acid that can then alter soils and waterways and lead to oceanic acidification that negatively impacts calcified marine biota (e.g., Kerr, 2005; Kiessling & Simpson, 2011; Hönick et al., 2012; Jones et al., 2016). Acid rain can also be produced by hydrogen halides (HF, HCl, and HBr) emitted during eruptions, which rapidly dissolve in water droplets within volcanic plumes or the atmosphere. In an ash-producing eruption, ash particles are also often coated with hydrogen halides (Rose, 1977).

Our ability to estimate volatile contents typically comes from petrologic constraints where we measure volatile species either in preserved volcanic glasses or in tiny droplets of melt trapped by crystals as they grew in the magma chamber prior to eruption. For eruptions occurring over the last 30 to 40 years, independent direct measurements of SO$_2$ fluxes have been possible using remote sensing techniques such as ground- and air-based UV correlation spectrometers (COSPEC) and satellite-based total ozone mapping spectrometers (TOMS; Wallace et al., 2003). When the SO$_2$ emission data are compared with petrologic estimates, this has led to the identification of an “excess S” problem where concentrations of S dissolved in the magma are typically 1 to 2 orders of magnitude too low to account for the total mass of SO$_2$ emissions measured in an eruption (Andres et al., 1991; Wallace, 2001). This was most evident for the 1991 Mount Pinatubo eruption where 18–19 +/- 4 Mt of SO$_2$ was ejected into the stratosphere (Guo et al., 2004), much more than petrologic estimates indicated (Wallace & Gerlach, 1994). This excess S is a feature of intermediate to silicic explosive eruptions, whereas basaltic eruptions do not show evidence of excess S (Wallace et al., 2003; Sharma et al., 2004; Self et al., 2005). The excess S is explained by the presence of an exsolved S-bearing vapor phase in the magma chamber prior to eruption, and that the ultimate source of the additional S is from underplating basaltic magmas that both heat the crust resulting in crustal partial melting to generate the silicic magmas, and release volatiles that can invade and pool in magma chambers residing at shallower depths (Wallace et al., 2003).

Petrologic estimates for S emissions from LIP silicic eruptions should be considered as minima. Observations from historic eruptions indicate the potential for much more S to be released during silicic explosive eruptions, which also have the eruptive mechanism to efficiently inject large tonnages of S directly into the stratosphere.

**CO$_2$ Regional Degassing**

Active subaerial volcanoes are the main source for the efficient expulsion of magmatic CO$_2$ to the atmosphere (Kerrick, 2001) as headlined by Mount Etna where approximately 25 Mt a$^{-1}$ of CO$_2$ is emitted, representing approximately 15% of present-day global volcanic CO$_2$ emissions (Gerlach, 1991). Given that CO$_2$ can sometimes exceed H$_2$O in concentration in gases exsolved from basaltic magmas and CO$_2$ solubility is higher in alkali basaltic magmas (Lowenstern, 2001), the large tonnages of magmatic CO$_2$ estimated for LIP events (Table 6.2) are generally thought to have been emitted directly from vents during flood basaltic eruptions. However, elevated CO$_2$ emissions are common from dormant volcanoes and, for example, minimum CO$_2$ emissions have been estimated at 45,000 ± 16,000 t d$^{-1}$ from the Yellowstone silicic caldera (Werner & Brantley, 2003). This CO$_2$ flux is among the highest known for individual volcanic centers and is approximately 5% of the estimated global volcanogenic CO$_2$ flux (Werner & Brantley, 2003; Lowenstern & Hurwitz, 2008). Consequently, silicic magmatic systems can also supply significant tonnages of magmatic CO$_2$. This is because much of the CO$_2$ (and heat) are derived from large volumes of degassing basaltic magma that underplate the crust at depth and underpin upper- to mid-crustal rhyolitic magma reservoirs such as at Yellowstone (White et al., 1988; Christiansen, 2001). At a more regional scale, large-scale and sustained mantle thermal and material inputs into the crust are required to generate widespread crustal partial melting and generation of the large volumes of silicic magmas found in continental LIPs (Bryan et al., 2002; Bryan, 2007) as well as significant crustal melt contributions to flood basalt magmas (e.g., Carlson & Hart, 1987; Coble & Mahood, 2012). This is the basis for envisaging silicic LIPs as “hidden” mafic LIPs (Bryan & Ferrari, 2013; Ernst, 2014). Consequently, it can be predicted that large magmatic CO$_2$ fluxes would also occur during silicic LIP events, but might not be directly associated with eruptions. Much greater greenhouse gas emissions may occur over large areas during LIP and rifted margin events, where upwelled asthenosphere residing at shallow depths can lead to...
elevated CO₂ fluxes from the mantle (e.g., D’Alessandro et al., 1997).

Evidence to support significant regional mantle CO₂ degassing during a silicic LIP event is recorded in eastern Australia in association with the Whitsunday Silicic LIP (Bryan et al., 2012). Regional mantle CO₂ degassing in the Early Cretaceous promoted widespread authigenic carbonate and clay mineral precipitation hosted within the late Permian Bowen-Gunnedah-Sydney basin system as well as in central Australian sedimentary basin successions, well inboard of the Whitsunday Silicic LIP along the eastern Australian coast (Baker et al., 1995; Uysal et al., 2011; Golab et al., 2006; Middleton et al., 2014).

**Iron Fertilization**

An additional way volcanism can perturb the global carbon cycle is by enhancing biological productivity in the oceans induced by iron fertilization that increases atmospheric CO₂ drawdown (Iron hypothesis of Martin, 1990). Iron availability is known to limit oceanic plant nutrition and photosynthetic productivity such that approximately 30% of present-day oceans are high-nutrient, low-chlorophyll (HNLC) regions where low concentrations of bioavailable iron limit phytoplanktonic growth. Phytoplankton blooms have been observed following iron additions to ocean waters (e.g., Boyd et al., 2007; Blain et al., 2007). The significance of this “biological pump” mechanism is that Fe-induced productivity could have contributed up to 30% of the 80 ppm drawdown in atmospheric CO₂ observed during past glacial maxima when Fe supply to the oceans was much higher (Martin, 1990; Sigman & Boyle, 2000).

Natural sources of iron addition to surface waters of deep oceans are aeolian dust, volcanic ash, and hydrothermal venting from submarine volcanoes (Jickells et al., 2005; Cather et al., 2009; Achterberg et al., 2013; Hawkes et al., 2014; Zeng et al., 2018). Iron is delivered to surface ocean waters from subaerial explosive eruptions in two main ways: (1) as iron stored in glass occurring as ash particles and pumice clasts and (2) as highly soluble Fe-bearing salt layers coating tephra comprising metal sulphates and halides that form within volcanic plumes through reactions with acidic magmatic gases (i.e., HF, HCl, SO₂, H₂SO₄; Duggen et al., 2010; Achterberg et al., 2013). Iron release into ocean waters can be rapid (minute-scale) ensuring iron is delivered to the sunlit euphotic zone where phytoplankton thrives (Frogner et al., 2001; Duggen et al., 2007). Slower and more sustained release of Fe from glass can then occur as much of the iron in ash is in the ferrous form and thus in a more soluble oxidation state. Buoyant pumice rafts, which can remain afloat for years in the world’s oceans (Bryan et al., 2004, 2012), therefore, have the potential to supply soluble Fe to ocean surface waters.

Despite significant CO₂ fluxes to the atmosphere from volcanoes and predictions for long-term CO₂ forcing by LIPs (e.g., Wignall, 2001; Kidder & Worsley, 2010; Saunders, 2016), short-term but pronounced decreases in atmospheric CO₂ have followed the larger explosive eruptions of the last century (Sarmiento, 1993; Cather et al., 2009). For example, atmospheric CO₂ was reduced for up to 2 years following the 1991 Mount Pinatubo eruption and was accompanied by an increase in atmospheric O₂ (Keeling et al., 1996). Both the negative correlations in CO₂ and O₂ and the timing of these perturbations are consistent with an oceanic response to iron fertilization in the Southern Ocean from the 1991 Pinatubo eruption (Watson, 1997).

A clear temporal correlation exists between very-large-volume silicic volcanism, cold climates in the Phanerozoic, and low-temperature anomalies in tropical sea-surface paleotemperatures (Veizer et al., 2000; Cather et al., 2009; see also Youbi et al., Chapter 8 this volume). This relationship led to the development of the Icehouse-Silicic LIP hypothesis (Cather et al., 2009). LIP events that produce large volumes of tephra, which can be delivered to

### Table 6.2 Comparison of Present-Day Atmospheric Budgets (Gigatons) of Key Magmatic Gases Implicated in Climate Forcing or Ozone Destruction With Estimated Total Emissions From Selected LIP Events

| Source          | CO₂ (Gt) | S (Gt)          | F (Gt)         | Cl (Gt) | Br (Gt) |
|-----------------|----------|-----------------|----------------|---------|---------|
| Atmosphere      | 3,000    | -4.8 x 10⁻³     | 7.5 x 10⁻³     | 8.3     | 3.6 x 10⁻³ |
| Siberian Traps  | 30,000–100,000* | 6,300–7,800 | 7,100–13,600   | 3,400–8,700 | 10      |
| Deccan          | 14,000   | 3,300           | –              | 1,900   | –       |
| Paraná-Etendeka | -14,000  | 3,100–5,400     | 600–1,200      | 70–470  | –       |

* CO₂ emissions from the Siberian Traps includes a substantial contact metamorphic contribution.
ocean basins, particularly to areas of HNLC, have the potential to be a catalyst for biological drawdown of atmospheric CO₂. Key LIPs that fit these criteria include the Phanerzoic silicic LIPs as well as flood basalts provinces with either substantial volumes of silicic volcanic rock (e.g., Afro-Arabian; Uktstins Peate et al., 2005) or mafic volcaniclastic deposits (e.g., the Emeishan, Siberian Traps) as iron fertilization effects have been observed following basaltic phreatomagmatic eruptions (Achterberg et al., 2013). More generally, volcanic rifted margins are a LIP type having a natural proximity to ocean basins where this climate forcing mechanism can be important.

The best examples of the Icehouse Silicic LIP connection are the Eocene-Oligocene (Oligocene isotope event, Oi-1) and Oligocene-Miocene (Miocene isotope event, Mi-1) transitions, and a strong overlap also exists with the Kennedy-Connors-Auburn Silicic LIP and Skaggerak LIP (Fig. 6.1) with peak glacial intervals in the Permo-Carboniferous (Cather et al., 2009). The Oi-1 and Mi-1 events record both surface and bottom water cooling, a reduction in atmospheric CO₂ concentration, and expansion of continental ice volume (e.g., Florindo et al., 2015; Pound & Salzmann, 2017). These two cooling events coincided with eruptive pulses in the Sierra Madre Occidental (32–28 and 24–20 Ma; Bryan & Ferrari, 2013; Ferrari et al., 2013), and the Afro-Arabian LIP (~31–29 Ma; Uktstins Peate et al., 2003). The Sierra Madre Occidental Silicic LIP in western Mexico is located proximal to both the present-day equatorial and north Pacific HNLC regions, however, the geographic extent and location of ancient HNLC regions remain unknown.

The immense eruptive volume of the Whitsunday Silicic LIP (>3 million km³; Bryan et al., 2012), its relative proximity to the eastern margin of Gondwana and a paleolatitude of ~50–80°S (Bryan et al., 1997), coincident with the location of the present-day Southern Ocean HNLC, would suggest iron fertilization and atmospheric CO₂ drawdown should have been an important feature in the Early Cretaceous. However, only moderate cooling was observed in the middle Aptian to early Albian (ca. 120–110 Ma ago; Cather et al., 2009). The Whitsunday Silicic LIP was emplaced at a time of other large mafic LIPs (Ontong Java Nui, High Arctic LIP, Kerguelen-Broken Ridge-Rajmahal-Sylhet; Bryan & Ferrari, 2013), which collectively reinforced early Aptian greenhouse conditions through volcanic CO₂ emissions, such that early Aptian atmospheric CO₂ concentrations were three to four times preindustrial level concentrations (Heimhofer et al., 2004). Much of the eruptive volume of the Whitsunday Silicic LIP was emplaced into a continental seaway (Great Australian Basin; Bryan et al., 1997, 2012) and other continental rift basins minimizing delivery to HNLC ocean regions.

**Mercury Fluxes**

Toxic metal poisoning from LIP eruptions is a recently identified potential kill mechanism based on the occurrence of notable spikes in Hg concentration (and other elements) in sedimentary records spanning the time interval of several mass extinction events (e.g., Sanei et al., 2012; Sial et al., 2014; Percival et al., 2015, 2018; Grasby et al., 2016; Font et al., 2016; Thibodeau et al., 2016; Jones et al., 2019; see also Percival et al., Chapter 11 this volume). Hg is strongly enriched in volcanic eruptions, and volcanoes are the only natural sources of direct Hg emission to the troposphere and stratosphere (Pyle & Mather, 2003). Sill intrusions can also be a source of Hg to the environment, both from magmatic degassing and the volatilization of organic material during contact metamorphism of sedimentary rocks (Jones et al., 2019). This makes Hg a potentially valuable fingerprint for LIP events and a discriminatory tool to identify LIP- or asteroid impact-timings relative to extinction events. In addition, volcanic emissions are the major natural source of several heavy metals (e.g., As, Cd, Cu, Pb, Zn, Tl, Sb, and Sn) to the atmosphere by contributing 20–40% of volatile elements such as Bi, Pb, As, or Sb and up to 40–50% of Cd annually (e.g., Hinkley et al. 1999; Henley & Berger 2013; Bagnato et al., 2014).

The atmosphere is an important pathway for the introduction of Hg into marine and terrestrial aquatic systems (Vandal et al., 1995). Hg is an unusually volatile and highly incompatible metal lending itself to concentrate in silicic magmas. When released into the vapor phase during magma degassing or volcanic eruptions, it can have a long-residence time in the atmosphere (0.5–2 yr), which facilitates global dispersion (Schroeder & Munthe, 1998; Blum et al., 2014). Despite limited observation and difficulties in measurement, substantial short-term increases in the atmospheric Hg burden have been observed following eruptions, overwhelming global Hg budgets (Pyle & Mather, 2003). Although a lack of direct measurements exists on the volatile metal burden of explosive, ash-rich volcanic plumes, available evidence indicates emissions from large explosive (silicic) eruptions are Hg-rich and greater than effusive (basaltic) eruptions (Pyle & Mather, 2003).

The new Hg data emerging from many sedimentary successions highlight some key points. First, the analyses have been from shallow to deep marine sedimentary successions that were accumulating at remote locations from the then-active LIP and thus require long-range transport of Hg (Sanei et al., 2012). Second, this sedimentary record therefore implies substantial and persistent atmospheric loading of Hg had occurred (where atmospheric Hg is then transferred into the ocean waters). Third, it also implies that Hg dispersion was most likely global,
which is best achieved by having been injected into the stratosphere. Substantial stratospheric injection of aerosols requires explosive eruptive styles. However, new studies have found inconsistent signals of sedimentary Hg enrichment indicating that Hg dispersal range may be limited if the LIP volcanism is submarine (oceanic plateaus or degassing from hydrothermal vent complexes) versus subaerial, and that not all subaerial LIP volcanism is necessarily able to perturb the global Hg cycle (Percival et al., 2018; Jones et al., 2019). As with stratospheric delivery of other key climate-forcing aerosols, we remain challenged in that the current and well-established model for continental flood basalt volcanism invokes decadal effusive eruptions to produce giant pahoehoe lava flow fields (e.g., Self et al., 1997; Thordarson & Self, 1998; White et al., 2009). Only relatively brief periods of eruption vigor are evident (White et al., 2009; Brown et al., 2014) analogous to the peak phases of the 1783 Laki eruption in Iceland that may have been subplinian (Thordarson & Self, 2003). However, geochemical evidence is lacking for the Laki eruption plume to have reached altitudes of the stratospheric ozone layer (Lanciki et al., 2012). Consequently, flood basalt eruptions lack a viable eruptive mechanism to efficiently and substantially inject Hg and other aerosols into the stratosphere for effective global dispersion.

Further insight into the problem of global dispersion of Hg from volcanic eruptions is gained from ice core records, which now extend from Greenland to Antarctica (e.g., Zielinski et al., 1997; Cole-dai et al., 2000; Schuster et al., 2002; Yalcin et al., 2007; Ren et al., 2010; Eyrikh et al., 2017). The global ice core archive demonstrates the global reach of large silicic explosive eruptions (VEI 5–7) during the Holocene, such as fallout from the 1991 Mount Pinatubo eruption in Antarctica (Cole-dai et al., 1999) and fallout from the 1815 Tambora eruption in Greenland (Clausen & Hammer, 1988). In contrast, basaltic eruptions are only recorded in ice cores regionally, and the 1783 Laki basaltic eruption (VEI 6) in Iceland, for example, has no expression in Antarctic ice cores (see Cole-dai et al., 2000; Ren et al., 2010). This suggests that Hg and other aerosol emissions for the Laki eruption were essentially restricted to the troposphere. Consistent with this are findings that Hg deposition was regionally constrained during volcanism of the North Atlantic Igneous Province with the magnitude of Hg anomalies in sedimentary sections correlating with proximity to the LIP (Jones et al., 2019). Another emerging feature is that in some cases, records of volcanic eruptions are evident in sulfur signals but are not always accompanied by high Hg concentrations (Eyrikh et al., 2017). This supports the conclusion of Percival et al. (2018) that not all volcanic eruptions may leave an observable Hg imprint in remote sedimentary archives (Eyrikh et al., 2017). The controls on why some eruptions produce an Hg signal and others do not remain poorly understood. Nevertheless, a key feature of the ice records is that the large silicic explosive (plinian) eruptions are generally recorded globally, and consistently show the strongest evidence for perturbing the global mercury cycle.

### 6.4. CONCLUSIONS

A wide variety of direct and indirect impacts are evident from small-scale volcanism experienced in historic times. Pyroclastic density currents and ash fall are a feature of silicic explosive eruptions and strongly impact the landscape. A range of longer term impacts affecting biota health are also evident including fluorosis, silicosis, acid rain, and global mercury/heavy metal fluxes. These impacts occur at different timescales ranging from seconds/minutes to years after the eruption. Although occurring less frequently, the larger silicic explosive eruptions have demonstrably caused more fatalities, triggered brief periods of climate cooling, and led to global disturbances of the carbon, mercury, and sulphur cycles. Collectively, these observations support the conclusion of moving from single (e.g., warming) to multicausal effects in the way that LIP events trigger mass extinctions (Bond & Grasby, 2017).

While a strong temporal relationship exists between LIP events and mass extinctions, a number of observations confuse the relationship: (1) that no correlation exists between the magnitude of a mass extinction and magnitude of the corresponding LIP event(s) as predicted for asteroid impacts; (2) that not all LIPs have caused mass extinctions nor are mass extinctions coincident with a clustering of LIP events; (3) that ecosystems may have already been under stress in those cases where mass extinction occurred; and (4) more than one type of kill mechanism may be operational during LIP events (Bryan & Ferrari, 2013; Bond & Grasby, 2017). All LIPs share a number of common features (Bryan & Ernst, 2008) but subtleties exist for each LIP event such as (1) the total igneous volume and LIP duration; (2) the number, intensity, and brevity of eruptive pulses (Table 6.1); (3) the crustal setting (continental or oceanic) and relative proximity to ocean basins and then-active plate boundaries; (4) the relative proportion of mafic to silicic igneous rocks, and relative timing of mafic and silicic eruptions (Fig. 6.1); (5) the number and magnitude of basaltic and silicic supereruptions; (6) the relative proportion of explosive (including both magmatic- and hydromagmatic-driven explosivity) versus effusive eruptions; (7) the degree of interaction with sedimentary basin substrates where aerosol budgets can be enhanced by thermogenic mechanisms; (8) the paleolatitudinal location
that has relevance both to the height of the tropopause and ease with which LIP eruptions can inject aerosols into the stratosphere, as well as deposit tephra into HNLC regions of the world’s oceans; and (9) the total aerosol budgets and components (Table 6.2).

Despite these complexities to LIP events, a number of requisite features can be identified that promote a LIP event to being a global killer of life. First, new focused and well-constrained age dating studies employing high-precision methods are tightly constraining the short-lived eruptive pulses in LIP events and demonstrate that much of the total eruptive volume is emplaced within a few hundred thousand years (Table 6.1). These data are suggesting, however, that there may be two classes of LIPs: one that is hyperactive with average eruption rates >10^3 km^3/kyr providing scope for multiple supereruptions every 1,000 years, and another class of more “sluggish” LIPs where the bulk of the eruptive volume is emplaced in pulses of 1–5 Myr duration with an order of magnitude lower average eruption rates (<10^3 km^3/kyr), and therefore having a much lower frequency of supereruptions. The Siberian Traps, CAMP, and Deccan LIPs that coincide with three of the top five extinction events (e.g., Bond & Grasby, 2017) are characterized by hyperactive eruptive pulses (Table 6.1). The important point here is that LIPs with hyperactive pulses reduce recovery time. The lack of LIP magnitude/cluster correlation with mass extinctions can then be understood in terms of some LIPs having sluggish volcanic pulses, and the biggest LIPs, the oceanic plateaus, lacking stratosphere penetrating eruptions.

Second, recent studies on volatile estimates from LIP magmas are highlighting differences in total aerosol budgets, particularly in terms of halogens (Table 6.2; Black et al., 2012; Marks et al., 2014). Those LIPs coinciding with the big mass extinction events are characterized by very large aerosol emissions, which may have been enhanced by thermogenic processes and sedimentary basin interaction (Black et al., 2012, 2014; Svensen et al. 2004, 2007, 2009). These data are implicating halogen aerosols as important kill mechanisms. So, rather than warming or cooling, which likely operate on longer timescales, it may be that atmospheric destruction and rapidly repeated cycles of ozone loss and acid rain may be the real kill mechanism for mass extinctions. Rapid decadal-scale, extreme fluctuations in pH driven by acid rain, S-, or iron fertilization-driven temperature chills, and toxic UV radiation bursts could be repeated within as little as a few hundred years of each other during hyperactive LIP pulses. The challenge here then is that such toxic UV/acid rain bursts are likely to be too short lived to detect via geochronology and climate sensitive recorders and are unlikely to cause large shifts in temperature, and biological markers will prove difficult to find for Paleozoic Mass Extinctions.

Third, mass extinctions require rapid and indiscriminate loss of life across the planet in both the terrestrial and marine realms, and the atmosphere is the obvious linkage between these environments (Bond & Grasby, 2017). Toxic metal signals, such as Hg at mass extinction event horizons, provide strong evidence for global dispersion via the atmosphere and are supported by ice core records of historic eruptions (e.g., Schuster et al., 2002; Eyrikh et al., 2017). Consequently, for the huge tonnages of aerosols emitted (Table 6.2), it is about getting them into the stratosphere to maximize residence time, alter atmospheric chemistry, and disperse globally. This is because tropospheric additions are quickly rain flushed and, thus, aerosols (excluding CO2) have short residence times. This requirement highlights a fundamental problem in that the bulk of LIPs are built up by effusive styles of eruption (to construct flood basalt lava fields) that lack eruption mechanisms to persistently inject aerosols to atmospheric heights of 15 km or more. There are two main ways of overcoming this eruptive barrier: by enhancing explosivity for the flood basaltic magmas through magma-water interaction (i.e., phreatomagmatic styles of eruption) or by magmatic-driven explosivity that characterizes the silicic supereruptions in LIPs (Bryan et al., 2010). Large-volume mafic phreatomagmatic volcanism is a feature of the Siberian Traps (e.g., Svensen et al., 2009; Black et al., 2012). Silicic eruptions during LIP events are of similar eruption magnitude to flood basalt eruptions but record much greater intensity eruptions where mass eruption rates are estimated to have reached 10^11 kgs^-1, two orders of magnitude higher than the largest historic plinian and climate-interrupting eruptions (Bryan et al., 2010). These silicic explosive eruptions provide several opportunities for sustained stratospheric delivery of ash and aerosols, via the eruption column at the vent and by co-ignimbrite eruption plumes developed above areally extensive (10^4–10^5 km^2) pyroclastic density currents. Silicic supereruptions would therefore be expected to be a contributing factor in causing major environmental disturbance during LIP events.

The large eruptive fluxes and upper-atmospheric effects of silicic and mafic explosive supereruptions may combine with effects of tropospheric-impacting flood basalt eruptions to force environmental and climatic changes during LIP emplacement, so it is important to understand their timing relative to, and petrogenetic relationships with, coeval flood basalt lavas (White et al., 2009). A critical yet unresolved factor then is the coupled effect of mafic and silicic eruptions from continental LIPs. Closely spaced basaltic and silicic, or effusive and explosive, eruptions that work in tandem to overload the troposphere and stratosphere with volcanic aerosols, could produce the rapid decadal-scale, extreme fluctuations in pH driven by acid rain, S-, or iron fertilization-driven
temperature chills and toxic UV radiation bursts. These
effects could be repeated within as little as a few hundred
years of each other particularly during hyperactive LIP
pulses, making it extremely difficult for ecosystems to
fully recover after each supereruption and culminate in
biological turnover.

ACKNOWLEDGEMENTS

Dougal Jerram and an anonymous reviewer are
thanked for constructive reviews to help improve the
manuscript. Richard Ernst is thanked for the invitation
to contribute to this book.

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