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Letter

Did mantle plume magmatism help trigger the Great Oxidation Event?

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1. Introduction

The Great Oxidation Event (GOE; Holland, 2002) represents the first sustained appearance of free oxygen in the atmosphere. This event which was fundamental to the evolution of complex, multicellular life on Earth (Caitling et al., 2005; Lane and Martin, 2010) occurred at ~2450 Ma with a shift from prevailing atmospheric anoxia to more oxic conditions (e.g., Lyons et al., 2014). This change is evidenced in the geological record by the absence of detrital minerals that are unstable in the presence of oxygen, in sediments younger than the Archean. These minerals are however common in older rocks (Frimmel, 2005; Rasmussen and Buick, 1999). Moreover, mass-independent fractionation (MIF) of sulphur isotopes [proposed to indicate a reducing atmosphere (Farquhar and Wing, 2003)] has not been found in rocks deposited since the earliest Proterozoic (Johnston, 2011) (Fig. 1). Indeed, it has been shown that in order to inhibit the MIF of sulphur isotopes, atmospheric oxygen must have risen to at least 1×10^{-5} present atmospheric levels at the start of the Proterozoic (e.g., Kump, 2008; Pavlov and Kasting, 2002).

Although it has long been argued that atmospheric oxygenation at 2450 Ma was driven by cyanobacterial photosynthesis (e.g, Canfield, 2005; Kopp et al., 2005), the Archean stratigraphic record contains even older evidence of Archean oxygenic photosynthesis in the form of distinct isotopic signatures (e.g., δ^{98}Mo values that are consistent with interaction with Mn oxides, and strongly negative δ^{13}C) and biomarker molecules (indicative of cyanobacterial metabolic processes) are preserved within sediments that were deposited hundreds of millions of years prior to the GOE (Planavsky et al., 2014; Rosing and Frei, 2004). Aside from these isotopic data, paleontological evidence in the form of fossilised tufted microbial mats – which in the modern era are dominated by cyanobacteria – are preserved in rocks as old as 2.72 Ga (Flannery and Walter, 2011). The time lag between the onset of photosynthesis and the GOE suggests that other mechanisms are likely to have operated ~2450 Ma to either increase the rate of oxygen production, or alternatively, inhibit the ability of the Archean sinks to remove O_2 as it was photosynthesised.

Various mechanisms have been suggested to explain the lag between the onset of photosynthesis and the GOE (see review in Kasting, 2013). For example, partial cessation of ultramafic volcanism towards the end of the Archean aeon may have caused a decrease in the flux of nickel into the oceans (Konhauser et al., 2009). Limited nickel supply could have arrested the activity of Archean methanogens leading to a decrease in the amount of atmospheric methane which would have otherwise consumed photosynthetic O_2. Alternatively, periods of continental collision and orogenesis may have been the trigger for the GOE by increasing primary productivity in the oceans through increased nutrient supply, as well as increasing organic carbon burial rates (Campbell and Allen, 2008). Other models propose that increasing the size of the early Proterozoic continental shelf seas during episodes of continental rifting promoted organic carbon burial (Lenton et al., 2004). A change in the nature of volcanic gases from more reduced compositions during the Archean to more oxidised compositions during the Proterozoic has also been suggested as a cause of the GOE (Kump and Barley, 2007). However, all these mechanisms have difficulty in explaining the apparent abruptness (see Lyons et al., 2014 for an alternative interpretation) of the change at the time of the GOE.

Alternatively, it has recently been proposed that O_2 released by the reduction of volcanogenic SO_2 (as sulphate ions in seawater) derived from Proterozoic subaerial volcanism may have driven the GOE (Gaillard et al., 2011). In this paper, we propose that the Matachewan LIP represents the main volcanic event responsible for the initial oxidation of the Earth's atmosphere. During volcanic eruptions, volatile species dissolved in the magma are released to a degree that is dictated by the confining pressure at which the eruption occurs (Gaillard et al., 2011). In modern systems, significant degassing of sulphur does not occur during subaqueous eruptions under confining pressures >100 bars, but can be almost total during subaerial eruptions (Rhodes and Vollinger, 2005). Empirically estimating the amount of sulphur (and hence, SO_2) released by ancient volcanic eruptions can be achieved by comparing the compositions of degassed lavas with those of related (but undegassed) subsurface magmas (Thordarson and Self, 2003). Here we focus on the Kaminak dyke swarm, a key component of the Matachewan LIP, preserved in the Central Hearne Supracrustal Belt, Nunavut, Canada (Sandeman and Ryan, 2008).

The Matachewan LIP is a reconstructed magmatic system of dyke swarms, layered intrusions, and the eroded remains of one of the largest continental flood basalt provinces in the geological record, fragments of which are now preserved in North America and Scandinavia (Fig. 2) (Ciborowski et al., 2015). The radiating geometry of the dyke swarms and high volumes of mafic rock preserved in this LIP imply a mantle plume origin for the magmatism (Ernst and Buchan, 2002). U—Pb ages for the different igneous suites show that magmatism began with the intrusion of the dyke swarms and layered intrusions ~2495 Ma (Vogel et al., 1998). Following the establishment of this crustal magmatic system, continental flood basalts were erupted on the Karelia, Kola, Hearne, and Superior cratons. Crucially, the preserved flood basalts record eruptive ages of between ~2432 and ~2453 Ma (Ketchum et al., 2013; Melezhik, 2006) with an average age of 2442 Ma (Table 1; Fig. 1) – indistinguishable from age estimates of the GOE (Johnston, 2011). Given the remarkably coeval nature of the GOE and Matachewan LIP volcanism, especially in light of previous studies that have

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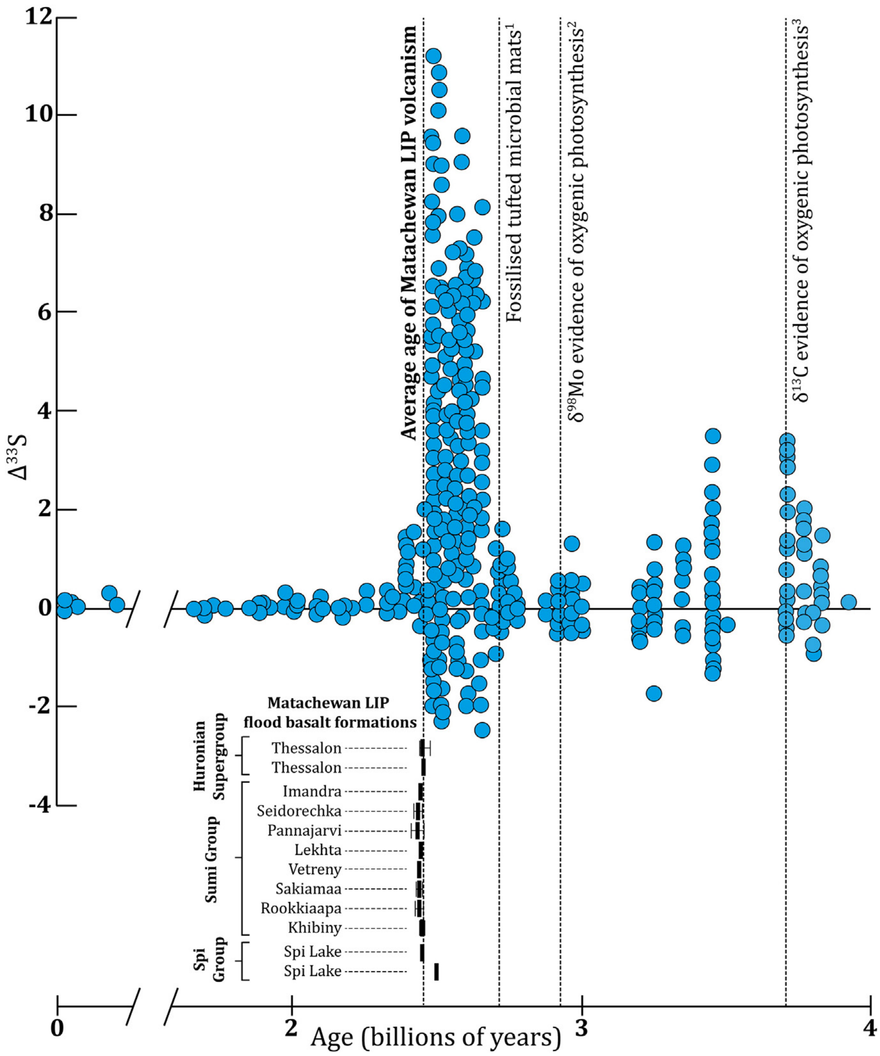

Fig. 1. Δ^{33}S versus age. Note the prevalence of MIF of sulphur isotopes in rocks older than 2450 Ma and absence in rocks deposited after. Also plotted are age estimates for each of the individual Matachewan LIP flood basalt formations (Table 1). Δ^{33}S data from (Johnston, 2011). Superscripts: 1 – Flannery and Walter (2011), 2 – Planavsky et al. (2014), 3 – Rosing and Frei (2004).

demonstrated the ability of LIPs to help drive fundamental change in the biosphere (e.g., Courtillot et al., 1986; Renne et al., 1995), the potential of a causal link between the Matachewan LIP and the GOE must be explored.

The Kaminak dyke swarm is made up of hundreds of NNE trending gabbroic dykes, which range in thickness from 1 to 40 m. Stratigraphically above, and orientated parallel to the strike of, the Kaminak dykes is the 8 km^2 Spi Basin which contains a 75–100 m thick sequence of basaltic lava flows and intercalated sediments. These basalts have identical trace element (Fig. 3) and radiogenic isotopic compositions to the Kaminak dykes and are interpreted to be their eruptive equivalents, emplaced into the basin during a period of crustal extension (Sandeman and Ryan, 2008). The Kaminak dykes and the cogenetic Spi Basin lavas are therefore an ideal system for investigating the amount of sulphur released by the Matachewan LIP.

2. Results

Using a thickness of 75 m for the Spi Basin lavas across the 8 km^2 basin and an average basaltic density of 2900 kg m^{-3}, the mass of lava can be estimated as 1.74×10^{12} kg. Whole rock data (15 samples) for the Kaminak dykes and Spi Basin lavas, on average, show that they contain 1036 and 710 ppm sulphur, respectively (Sandeman and Ryan, 2008), which implies that the 326 ppm deficit was lost to the atmosphere during degassing of the erupted lava. Using established methods used for calculating volatile release in modern eruptions (Thordarson and Self, 2003), the sulphur difference between the degassed lavas and undegassed dykes can be calculated to represent an absolute flux of 5.67×10^{10} kg of sulphur released into the Archean–early Proterozoic atmosphere during the eruption of the Spi Basin lavas (1).

$$(0.1036\text{–}0.0710) \times \left(1.74 \times 10^{12}\text{kg}\right) = 5.67 \times 10^{10}\text{kg} \tag{1}$$

The remaining Matachewan LIP flood basalts are preserved in the Huronian Supergroup (southern Ontario) and Sumi Group (Kola-Kaleria), the volcanic portions of which have average thicknesses of 1200 and 2500 m, respectively (Ketchum et al., 2013; Melezhik, 2006). Based on the presently exposed areas of these two groups, a minimum quantity of lava erupted as part of the Matachewan LIP may be estimated to be ~34,000 km^3 or ~ 9.91×10^{16} kg of basalt. If we assume that the Huronian Supergroup and Sumi Group basalts emitted sulphur in the same way as the coeval (and potentially cogenetic) Spi Basin lavas, the Matachewan LIP flood basalts would have released ~3.23×10^{15} kg of S into the early Proterozoic atmosphere. It is critical to note that these estimates of lava volumes are based on current eroded remnants of the province which, in comparison to modern, less-eroded LIPs, are significantly smaller. If instead, we were to assume that the

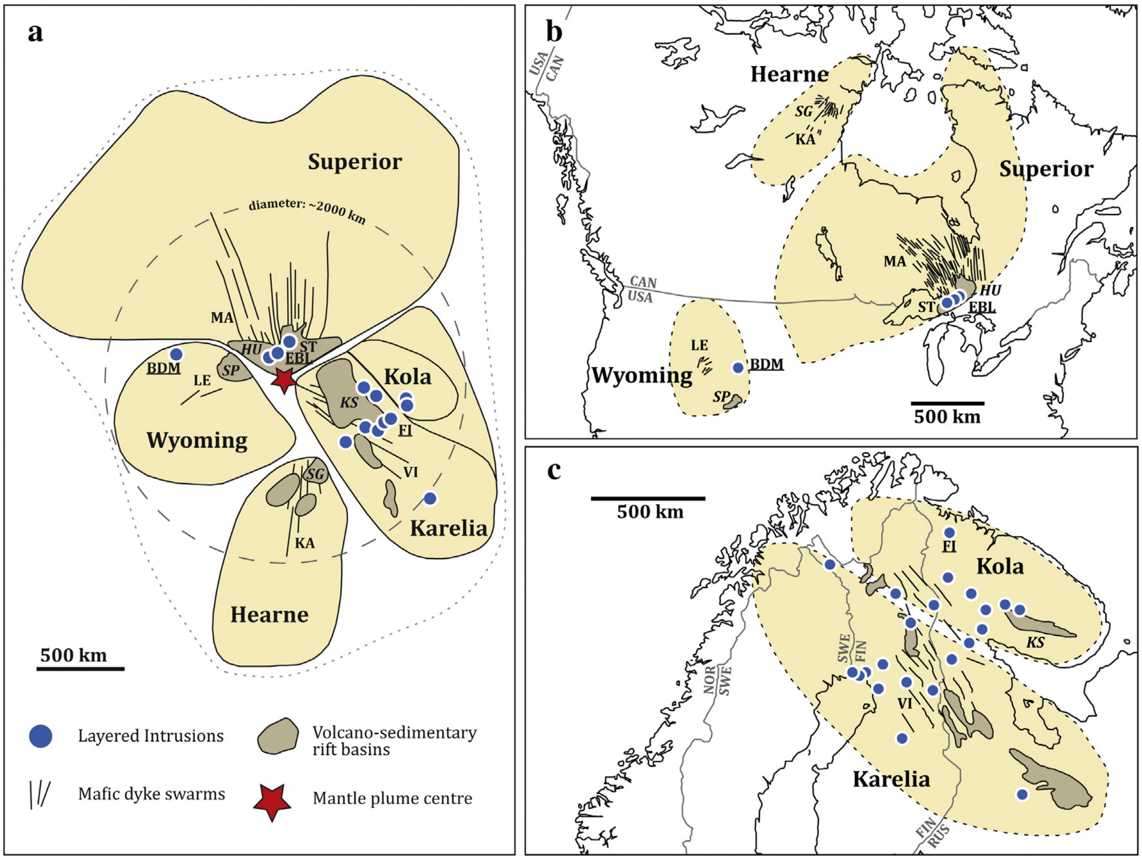

Fig. 2. (a) Early Proterozoic Matachewan LIP reconstruction (Modified after Bleeker and Ernst, 2006) and the Matachewan LIP's constituent suites' present-day distribution (b) and (c). When reconstructed to their inferred primary distribution, the composite radiating dyke swarm defines a mantle plume locus, melting at which triggered the emplacement of the LIP. Mafic dyke swarms: MA – Matachewan, VI – Viianki, LE – Leopard, KA – Kaminak, ST – Streich; Volcano-sedimentary rift basins: HU – Huronian Supergoup, KS – Karelia Supergroup, SP Snowy Pass Supergroup, SG – Spi Group; Layered Intrusions: EBL – East Bull Lake Suite, FI – Fennoscandian Intrusions, BDM – Blue Draw Metagabbro.

original Matachewan LIP was comparable in size to Phanerozoic analogues (as is suggested by the spatial distribution of the radiating dyke swarms of the province), then the estimates presented here should be increased by at least an order of magnitude.

During volcanic eruptions, sulphur is not emitted from lava in its native state and is instead lost as a mixture of S_2, H_2S, and SO_2. The relative proportions of these species vary as a function of pressure (Gaillard et al., 2011). The lack of pillow lavas within, and the preservation of regolith below, the Spi Basin, Huronian Supergroup, and Sumi Group lavas suggest that they were erupted subaerially under normal atmospheric pressures (Ketchum et al., 2013; Melezhik, 2006; Sandeman and Ryan, 2008). Under such conditions, tholeiitic-basalt eruptions emit SO_2, S_2, and H_2S in a molar ratio of approximately 2.5:1:1 (Fig. 4) (Gaillard et al., 2011). Thus, for every mole of sulphur emitted by volcanic eruptions, 0.544 moles of SO_2 and 0.222 and 0.234 moles each of H_2S and S_2 are released. By converting the 3.23×10^{15} kg of sulphur released by

Table 1
Summary of cited U—Pb ages for constituent Matachewan LIP Flood Basalt Provinces. Abbreviations: bad – baddeleyite, zir – zircon.

Flood Basalt Province	Age (Ma)	Analysis type	Reference
Huronian Supergroup			
Thessalon Formation	2450^{+25}_{-10}	zir	Krogh et al. (1984)
	2453 ± 3	zir	Ketchum et al. (2013)
Karelia Supergroup			
Seidorechka Formation			
Seidorechka	2434 ± 15	bad + zir	Bayanova and Balashov (1995)
Imandra	2442 ± 2	bad	Amelin et al. (1995)
Paanajärvi	2432 ± 22	zir	Buiko et al. (1995)
Lekhta	2443 ± 5	zir	Levchenkov et al. (1994)
Vetreny Belt	2437 ± 3	zir	Puchtel et al. (1997)
Sakiamaa	2438 ± 11	zir	Räsänen and Huhma (2001)
Rookkiaapa	2438 ± 14	zir	Manninen et al. (2001)
Khibiny	2448 ± 8	zir	Chashchin et al. (2008)
Spi Group			
Spi Lake Formation	2450 ± 2	bad	Heaman (1994)
	2498 ± 1	bad	Sandeman et al. (2013)

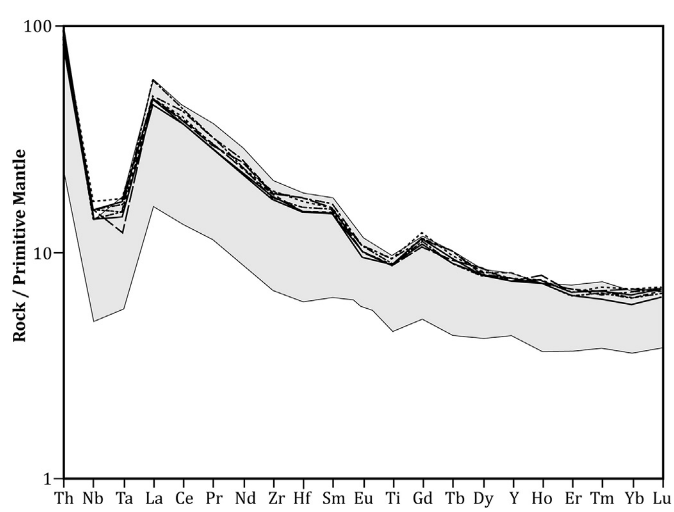

Fig. 3. Primitive Mantle-normalised trace element diagram showing the compositions of the Spi Group basalts (black lines) and the cogenetic Kaminak dykes (grey field). The Spi Group data (n = 7) are from Sandeman and Ryan (2008) while the Kaminak dyke data (n = 57) are from Ciborowski et al. (2015). Primitive Mantle normalising values from McDonough and Sun (1995).

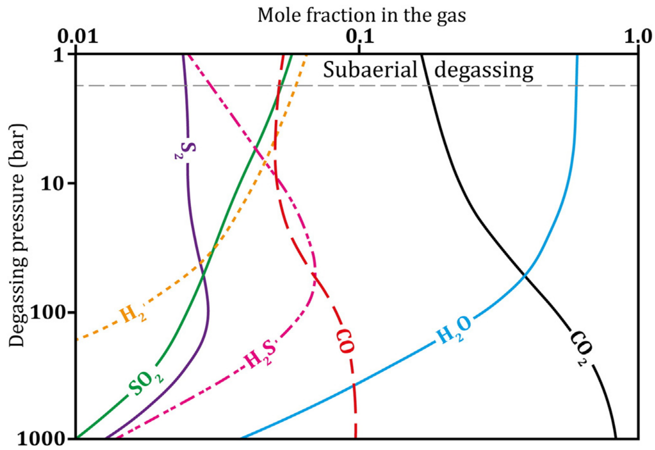

Fig. 4. Calculated compositions of volcanic gases as a function of pressure (Gaillard et al., 2011). During subaerial eruptions at very low degassing pressures, tholeiitic lavas lose sulphur as a mixture of SO_2, H_2S, and S_2.

the Matachewan lavas into moles of S (1.01×10^{17}) and using the eruptive molar ratios noted above, the number of moles of SO_2, S_2, and H_2S emitted to the atmosphere during the eruptions of the Matachewan LIP lavas can be estimated to be at least 5.49×10^{16}, 2.24×10^{16}, and 2.36×10^{16}, respectively.

3. Discussion

Recent work by Gaillard et al. (2011) suggests that decomposition of the SO_2 released by Paleoproterozoic subaerial volcanism (like that which characterised the Matachewan LIP) would have resulted in significant amounts of sulphuric acid (H_2SO_4) forming (2) in the atmosphere (Symonds et al., 2001):

$$4SO_2 + 4H_2O \rightarrow H_2S + 3H_2SO_4 \tag{2}$$

Dissolution of this H_2SO_4 in the Paleoproterozoic oceans produced sulphate (SO_4^{2-}) ions which would have been metabolised by sulphate-reducing bacteria to produce sedimentary pyrite (FeS_2) (Berner and Canfield, 1989; Kasting, 2013). The net result of sulphate reduction and subsequent pyrite formation (and burial) is an increase in atmospheric O_2 that can be summarised (3) by the following reaction (Lyons and Gill, 2010):

$$2Fe_2O_3 + 16Ca^{2+} + 16HCO_3^- + 8SO_4^{2-} \rightarrow 4FeS_2 + 16CaCO_3 + 8H_2O + 15O_2 \tag{3}$$

Balancing the two reactions presented above in terms of SO_4^{2-} shows that for every 32 moles of SO_2 released into the atmosphere by subaerial volcanism, 45 moles of O_2 may be produced via sulphate reduction in the oceans (4).

$$32SO_2 + 32H_2O \rightarrow 8H_2S + 24H_2SO_4$$

$$6Fe_2O_3 + 48Ca^{2+} + 48HCO_3^- + 24SO_4^{2-} \rightarrow 12FeS_2 + 48CaCO_3 + 24H_2O + 45O_2 \tag{4}$$

Using the 5.49×10^{16} moles of SO_2 estimated to have been released by the Matachewan LIP, we can calculate that the number of moles of O_2 produced by these reactions would have 7.72×10^{16}. This equates to a mass of O_2 of 2.47×10^{15} kg. The mass of the modern atmosphere is approximately 5.15×10^{18} kg (Trenberth and Smith, 2005) of which 23% (1.18×10^{18} kg) is O_2. Thus, the eruption of the Matachewan LIP flood basalts may have released an amount of O_2 equivalent to ~0.2% of that in the present-day atmosphere. This value is significant as it represents a potential oxygen input, greater in magnitude than that required to arrest the MIF of sulphur observed in the Archean geological record before the GOE (Kump, 2008; Pavlov and Kasting, 2002).

In comparison to the largest lava flow in recorded history – the 1783–1784 Laki eruption (15.3 km^3) – where the SO_2 released was similarly calculated to have been 1.22×10^{11} kg (Thordarson and Self, 2003), the Matachewan LIP is a truly enormous eruption. More significantly, the estimated eruptive volumes of the Matachewan LIP flood basalts are calculated from the size of current exposures and thus represent an absolute minimum estimate. Given the ages and post-intrusion tectonic histories of the constituent volcanic provinces, the modern day exposures represent only a small fraction of the original erupted Matachewan LIP flood basalt volume (Ernst and Buchan, 2002). Thus, the calculations above may significantly underestimate the actual amount of SO_2 released into the early Proterozoic atmosphere via the eruption of the Matachewan LIP. Indeed, if we were to assume that the original volume of the Matachewan LIP prior to its erosion was similar to that of the Siberian Traps (~4×10^6 km^3; i.e., ~120 times larger) see Maysatis, 1983), then the mass of oxygen released via the mechanisms explained above could have been up to 2.89×10^{17} kg – equivalent to ~20% of that in the present-day atmosphere.

It is important to note that the calculations above are underpinned by several assumptions. Firstly, estimating eruptive rates and associated fluxes of gas species to the atmosphere is difficult. This is largely the result of (often poor) temporal constraints on regions of the Matachewan LIP, with many of the constituent flood basalt provinces (currently) being constrained by one – very occasionally two – U—Pb radiometric ages. That said, the age of the Matachewan LIP volcanic rocks hosted within the Huronian Supergroup is bracketed by the ~2491 Ma Agnew lake intrusion, upon which the Huronian Supergroup sits unconformably (Vogel et al., 1998), and the ~2450 Ma Copper Cliff Rhyolite (Ketchum et al., 2013) preserved at the top of the volcanic sequence. Thus, the absolute maximum lifespan of the Matachewan LIP flood basalt province can be estimated to be ~40 myr, though is likely to be much less than this (Fig. 1). Without further geochronological constraints, the calculation of atmospheric fluxes is necessarily speculative.

The second assumption comes from the fact that we use sulphur analyses from the relatively minor Spi Basin lavas to constrain the SO_2 release of the LIP as a whole, despite the potential for variation across the LIP. Further, we assume that all of the SO_2 released by the volcanism is reduced in the oceans during sedimentary pyrite formation, and that the O_2 produced during sulphate reduction is free to accumulate in the atmosphere. This latter assumption is questionable as, aside from the SO_2 released by the Matachewan LIP magmatism, sizeable amounts of other reductants (e.g., CO, H_2 H_2S, and S_2) that could have consumed a portion of the O_2 produced through sulphate reduction (5, 6, 7, 8) would also have been released (Fig. 4; Gaillard et al., 2011).

$$2CO + O_2 \rightarrow 2CO_2 \tag{5}$$

$$2H_2 + O_2 \rightarrow 2H_2O \tag{6}$$

$$2H_2S + 3O_2 \rightarrow 2H_2O + 2SO_2 \tag{7}$$

$$S_2 + O_2 \rightarrow 2SO_2 \tag{8}$$

However, even when we factor in the O_2 consumption resulting from the release of the reductants listed above (the number of moles of which can be calculated using the relative proportions supplied in Fig. 2), as well the effects of reseeding of the atmosphere with the SO_2 produced via the oxidation of volcanogenic H_2S and S_2 (Eqs. (7) and (8)), the overall effect is still a significantly positive increase (6.00×10^{16} moles) in the amount of O_2 delivered to the Paleoproterozoic atmosphere. If we convert this number of moles of O_2 into kg, we can calculate a mass of O_2 produced by the Matachewan LIP eruptions of ~1.92×10^{15} kg.

Again, using Trenberth and Smith's (2005) estimate of the mass of the modern atmosphere (5.15×10^{18} kg) of which 23% (1.18×10^{18} kg) is O_2 and assuming that the Proterozoic atmosphere was not significantly different in terms of mass, we can see that the bulk O_2 addition caused by the eruption of the Matachewan LIP equates to an O_2 concentration equal to 1.62×10^{-3} that of present atmospheric levels – i.e., sufficient to inhibit the MIF of sulphur isotopes observed in sediments deposited during the Archean aeon (Kump, 2008; Pavlov and Kasting, 2002). Lastly, the effects of oxidative weathering of the continents (Campbell and Allen, 2008), as well as potential changes in methanogen production (Konhauser et al., 2009) following the erosion of the Matachewan LIP basalts, are not accounted for in our model.

Despite the potential buffering effects of these last two processes and the very conservative estimate of magmatic volumes, we must also remember that the Archean–Paleoproterozoic transition heralds the first emergence of the continents from the Archean oceans (Flament et al., 2008). One of the myriad effects of this fundamental change in the Earth system is that the composition of volcanic gases changed from more reducing to more oxidising as eruptions became more subaerial in nature (Gaillard et al., 2011; Kump and Barley, 2007).

4. Conclusion

Against this backdrop of global change, the Matachewan LIP represents one of the first, truly massive igneous events that occurs after the emergence of the continents and this shift in volcanic gas chemistry. Given the contemporaneous nature of the Matachewan LIP with the GOE, and the calculations detailed above, there is a strong case for the Matachewan LIP to be a principal driver of the GOE. The O_2 released during reduction of volcanogenic SO_2 in the oceans represents an input of sufficient magnitude to inhibit the MIF of sulphur and ultimately drive the oxygenation of Earth's atmosphere.

If valid, this mechanism for the initiation of the GOE will require a re-evaluation of these enormous magmatic systems, not just as drivers of biotic stress (Sobolev et al., 2011), but also to acknowledge their apparent ability to enable, and to drive, the evolution of life into the complex, and multicellular forms we see today.

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Appendix A. Supplementary data

Attached is a .doc file which details the calculations described in the discussion. Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.lithos.2015.12.017.

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