LARGE IGNEOUS PROVINCES: INTEGRATING LESSONS FROM VENUS AND EARTH. R. E. Ernst[1,2] and K. L. Buchan[3], [1]Department of Earth Sciences, Carleton University, Ottawa, Ontario, Canada; richard.ernst@ernstgeosciences.com; [2]Faculty of Geology and Geography, Tomsk State University, Tomsk, Russia; [3]273 Fifth Ave., Ottawa, Ontario, Canada

Introduction: In the absence of plate tectonics, all magmatism on Venus is by definition intraplate. It is thought to be mostly basaltic, except potentially within the oldest terranes, called tesserae, where spectral information indicates possible felsic compositions [1]. The most voluminous Venusian magmatism is considered analogous to that of Large Igneous Provinces (LIPs) on Earth [2-4]. The high atmospheric temperatures (450°C) and the absence of significant surface erosion for much of Venus's later history (subsequent to a proposed Great Climate Change event after ca. 1 Ga [5-7]) means that Venus preserves the surface expression of LIPs, a record which is largely lost on Earth due to erosion.

In this contribution we compare and integrate the terrestrial and Venusian LIP records.

Terrestrial LIP record: LIPs represent large volume (>0.1 Mkm^3; frequently above >1 Mkm^3), mainly mafic (-ultramafic) magmatic events of intraplate affinity that occur in both continental and oceanic settings [8]. They are generally linked to mantle plumes, although non-plume process can also be relevant. The main event, associated with plume head arrival, is typically of short duration (<1 myr) or consist of multiple short pulses over a maximum of a few 10s of myr. Lower flux plume tail magmatism can continue for up to 200 myr and produce a hotspot track as the tectonic plate moves over the plume. Individual LIPs comprise volcanic packages (flood basalts), and a plumbing system of dyke swarms, sill complexes, layered intrusions, and crustal underplates. They can also have associated silicic (felsic) magmatism, carbonatites and kimberlites.

Venusian LIP record: LIP-scale Venusian intraplate (mainly basaltic) magmatism is generally interpreted to have a plume origin, as with most terrestrial LIPs [2, 9]. Venusian LIPs can include the following major components: (1) large individual volcanoes with diameters >100 km, inclusive of flows (e.g. [10]); (2) coronae, which are quasi-circular tectonomagmatic structures that display (a) an annulus of graben that overlie dykes [11-12] with diameters averaging 300 km (up to 2600 km) and (b) topographic features such as a rim and/or moat [13-14]; (3) novae, which are radiating graben systems overlying laterally propagating dykes and extending up to 2000 km away from volcanoes or coronae at their foci [15-16]; and (4) lava flow fields of scale comparable to terrestrial flood basalt provinces and often associated with large volcanoes, coronae and rift/fracture belts [17].

In addition, there are other intraplate magmatic features that are likely minor components of LIP events: (1) intermediate size (20-100 km) volcanoes; (2) shield fields (fields of small shield volcanoes) that are up to a few hundred km across [3, 18-19]; and (3) narrow sinuous lava channels (e.g., canali) with lengths of hundreds or thousands of km [20-21].

All these tectonomagmatic features are superimposed on earlier plains volcanism that is usually interpreted as flood basalts and on highly deformed crustal plateaus/tesserae.

Flood basalts in tesserae: It is also possible that the oldest "basement" rocks, the tesserae, record a history of flood volcanism [7]. The recent recognition that there may be eroded flood basalts sequences in tesserae [22] and additional evidence for fluvial erosion [7] indicates that tesserae may preserve a much longer geological and LIP history than indicated by their crater counts. This can be explained if erosion removed older craters and only allowed preservation of craters once the climate had dramatically warmed and fluvial erosion had ended [7]. A long LIP history for tesserae would be comparable to the situation on Earth where a protracted LIP history is preserved in Precambrian basement rocks [23].

Plume centre region: The region within a few hundred km of the plume centre is of particular importance for both terrestrial and Venusian LIPs. Plume centres on Venus are marked by large volcanoes, corona centres and nova foci. On Earth, plume centres are poorly understood because of erosion and because they are typically on rifted margins that are deformed during subsequent continental collision.

Extent of flood basalts: On Earth, flood basalts are observed to extend out several hundred km from the plume centre, although the full original extent is difficult to determine due to erosion. However, on Venus, the absence of erosion, at least in the post-tessera record [7], allows the full extent of flood volcanism to be mapped, as well as revealing the sources (dykes or volcanoes) and flow patterns of individual lava flows.

Distal flows and sills fed via dykes from the plume centre region: On Earth, isolated lava flow packages and sill complexes are sometimes located at great distances from the plume centre (up to 2500 km), fed laterally by radiating dykes [24]. A link to a specific plume/LIP can often be demonstrated by precise U-Pb dating. On Venus, radiometric dating is not available. Nevertheless, distal flows can potentially be linked to a particular plume/LIP if the feeder dykes for the flows can be traced back to the plume centre.

Central caldera complex: On Venus, large volcanoes and coronae in the plume centre region often exhibit a large caldera complex (e.g. [25-26], thought to reflect roof collapse of an underlying magma chamber (e.g., [27]). Similar features may characterize plume centres on Earth, but the central calderas will usually have been lost to erosion. A terrestrial candidate for such a central caldera is the 1270 Ma Mackenzie LIP, where [28] inferred the presence of an apical graben (200-300 km diameter) at the focus of the Mackenzie dyke swarm.

Volcanic flows that emanate from the plume centre region may be associated with central caldera collapse by one of two possible mechanisms: (1) release of magma from circumferential fractures associated with the caldera rim (see [26]), or (2) lateral dyke injection from the underlying magma chamber (see small-scale Icelandic example of [29]).

Younger plume tail magmatism: On Earth, as noted above, short duration plume head magmatism is followed by lower flux plume tail magmatism, which can last up to 200 myr and yield a hotspot track due to plate movement. Plume tail magmatism is mostly apparent in oceanic areas. It is more rarely observed in continental areas because greater lithospheric thickness hampers magma generation and transport through the lithosphere.

Similar plume tail magmatism likely occurs on Venus. However, owing to the apparent absence of shifting of Venus's single plate, it will be superimposed on the plume head event itself. Detailed mapping and analysis may help distinguish the plume tail magmatism from more voluminous plume head magmatism. If plume tail magmatism is not observed, it might indicate that the lithosphere is too thick to permit such magmatism.

Link with climate change: On Earth, the link between LIPs and climate change has been greatly strengthened through precise U-Pb dating (e.g. [30]), which has shown that many LIPs are of short duration and many are associated with geological timescale boundaries [30,31]. The link is sufficiently strong in the Phanerozoic that it has been suggested that the selection of natural time scale boundaries in the Precambrian may be partly defined by LIPs [31,32].

On Venus, there is evidence for a Great Climate Change event from Earth-like conditions to the hyperwarm current conditions (450°C, 96% CO_2, dense atmosphere of 90 bars). The initial evidence was obtained from climate modelling [5-6]. More recent evidence comes from the geological record with the recognition of fluvial erosion in tesserae ([7]; cf. [22]). While a range of causes for this Great Climate Change event are suggested, the role of LIPs is a leading candidate [6]. To test this model, we need to better understand the LIP record during this critical time period.

References: [1] Gilmore M.S. et al. (2015) *Icarus*, 254, 350–361. [2] Head J.W. & Coffin M.F. (1997). *In: AGU GM* 100, 411–438. [3] Hansen V.L. (2007) *Chem. Geol*., 241, 354-374. [4] Ernst R.E. et al. (2007) In: Superplumes: Beyond Plate Tectonics., Springer, 537–562. [5] Way M.J. (2016) *GRL*, 43, 8376–8383. [6] Way M.J. & Del Genio A.D. (2020) *Adv. Earth Space Sci*. 125. [7] Khawja S. et al. (2020) *Nat. Comm., 11, 5789*. [8] Ernst R.E. (2014) Large Igneous Provinces. *Cambridge U. Press*. [9] Head J.W. et al. (1992) *JGR* 97, 13153–13197. [10] Crumpler L.S. & Aubele J.C. (2000) In: *Encyclopedia of Volcanoes*, Academic Press, 727–769. [11] McKenzie D. et al. (1992) *JGR*, 97, 15977–15990. [12] Buchan K.L. & Ernst R.E. (2019) In: Dyke Swarms of the World – A Modern Perspective. *Springer*, p. 263-314. [13] Stofan E.R. et al. (1992) *JGR*, 97, 13347–13378. [14] Gülcher A.J.P. et al. (2020) *Nat. Geosci.* 13, 547-554. [15] Grosfils E.B. & Head J.W. (1994) *GRL, 21, 701–704*. [16] Ernst R.E. et al. (2001) *Ann. Rev. Earth Planet. Sci.*, 29, 489–534. [17] Magee K. & Head J.W. (2001) *In: GSA SP* 352, 81–101. [18] Hansen V.L. (2005) GSA Bull., 117, 808-822. [19] Ivanov & Head 2004. *JGR*, 109, E100001. [20] Baker V. R. et al. (1992) *JGR*, 97 E8, 13421. [21] Komatsu G. et al. (1992). *GRL*, 19, 1415-1418. [22] Byrne P. et al. (2020) *Geology*, 49, 81-85. [23] Bleeker W. & Ernst R.E. (2010) *CJES*, 47, 695–739. [24] Ernst R.E. & Buchan K.L. (2001) In: *GSA SP*352, 483–575. [25] Head et al. (1991) *Science*, 252, 276-288. [26] MacLellan L.M. et al. (2021) *Earth-Sci. Reviews* (in review). [27] Mouginis-Mark P.J. (2016) *Icarus*, 277, 433-441. [28] Baragar W.R.A. et al. (1996). *J. Petrol*., 37, 317–359. [29] Gudmundsson M.T. et al. (2016) Science, 353, 6296. [30] Kasbohm J. et al. (2021) *AGU GM*, 255, 27-82. [31] Ernst R.E. & Youbi N. (2017) *PPP*, 478, 30-52. [32] Ernst R.E. et al. (2021) In: *AGU GM* 255, 3-26.