

Article

Mantle Plume Dynamics: A Key Driver of Proximal and Distal Metallogeny

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ABSTRACT

Mantle plume-generated Large Igneous Provinces (LIPs) are well known for their genetic association with magmatic sulphide Ni–Cu–PGE deposits, exemplified by Norilsk in the Siberian Traps. Previous work has broadened genetic these links between LIP events and metallogeny to include other orthomagmatic and hydrothermal deposits, lateritic enrichment, fluid-flow barriers and reactive zones that enable the formation of mineral deposits, and distal mineral systems indirectly influenced through plate tectonics. Here, we further examine how plume-related stress and topographic changes affect metallogenesis. These drivers include those directly above plume heads where key processes include: (1) rapid $\sim 90^\circ$ stress transitions from domal uplift to lateral spreading, facilitating emplacement of chonolith-hosted magmatic sulphide systems; (2) regional uplift influencing hydrothermal vein orientations; (3) basin tilting that redirects mineralized brines and promotes sediment-hosted mineralization; and (4) localized ascent of plume material through thick lithosphere, supplying heat and fluids for crustal mineralization. Distally, lateral plume-head spreading and pushing against adjacent cratonic roots may reorganize regional stresses, alter plate motions, and enhance convergence-related mineralization, including orogenic gold and porphyry systems along remote plate margins. These dynamic plume processes expand the metallogenic significance of LIPs and strengthen their use as predictive tools in regional mineral exploration, whereby a single plume-generated LIP event may generate or trigger diverse mineral systems.

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Research Highlights

- Plume–LIP dynamics can influence both proximal and distal mineral systems.
- Proximal effects include stress change, basin tilting, fracture (radial and circumferential) development and thermal–magmatic input.
- These processes may affect magmatic sulphides, IOCG, and uranium and other sediment-hosted systems.
- Distal plume push can change plate-scale stress and motion, influencing orogenic gold and porphyry systems.
- Plume geometry and timing provide new predictive criteria for regional mineral exploration.

1. Introduction

1.1. Overview on LIPs

Large Igneous Provinces (LIPs) are massive intraplate magmatic events, typically mafic to ultramafic in composition [1–7]. They have total volumes exceeding 0.1 million km³, often surpassing 1 million km³. These events occur in both continental and oceanic environments and are generally short-lived, lasting less than 5 million years, although they may consist of several discrete pulses extending over a few tens of millions of years. LIPs are characterized by flood basalt volcanism and an extensive plumbing system that includes mafic dyke swarms, mafic sill complexes, layered mafic intrusions, and underplating at the base of the crust, although not all of these aspects of a given LIP may be preserved in ancient examples. They can also be associated with silicic magmatism—referred to as Silicic LIPs (SLIPs), which can include super-eruptions, and can be associated with economically important alkaline magmatism such as carbonatites and kimberlites which may host significant precious-, rare- and critical-metal mineralization [4, 5, 8, 9].

Since the Archean, continental LIPs have formed irregularly but on average are generated every 20–30 million years [5, 7]. Although oceanic LIPs are only well preserved over approximately the past 200 million years, because older oceanic lithosphere has largely been subducted and older oceanic plateaus partly accreted and deformed in orogenic belts, their original occurrence through geological time is expected to have been broadly comparable to that of continental LIPs. This implies a combined oceanic and continental LIP recurrence interval of approximately 15–20 million years extending back to at least 2.5 Ga [5]. The Archean LIP record has also been recently significantly expanded, resulting in the recognition of events back to about 3.5 Ga that were of an occurrence frequency comparable to that of the Proterozoic–Phanerozoic LIP record [10].

LIPs are also consistently linked to major geological and environmental processes, including continental rifting and breakup (e.g., [11]) influences on sedimentary basins

relevant to hydrocarbons and aquifers [5, 12–15], and dramatic shifts in global climate, sometimes culminating in extinction events [7, 16–18]. In the context of this manuscript there are also significant direct and indirect genetic links with a range of ore deposit types [12, 13, 19, 20].

The genesis of LIPs remains debated, with proposed mechanisms ranging from deep mantle plumes to lithospheric delamination, thermal insulation by supercontinents, decompression melting during rifting (above a hot plume), and small-scale edge-driven convection [1, 3–5, 17, 21–25]. However, more precise age dating (emphasizing the short duration of many of these huge events), the presence of giant radiating and circumferential mafic dyke swarms [5, 26–28], seismic tomography, and compositional data for elevated mantle potential temperatures provide a strong case for a link between LIP events and mantle plumes whose buoyancy is mainly thermal [5, 29].

1.2. Characteristics of Mantle Plumes Associated with LIPs

Mantle upwellings span a substantial range in scale, from relatively small upper-mantle convective instabilities associated with localized intraplate magmatism to the very large plume heads associated with major LIPs. Small-volume intraplate basaltic provinces may extend over only ~100–500 km and have been variously attributed to short-wavelength upper-mantle convection, “mini-plumes” originating near the 660-km discontinuity, lateral channeling or fragmentation of larger plumes, slab breakoff, or lithospheric delamination (e.g., [22, 30–33]).

In contrast, LIP events, as defined above, are associated with much larger plume heads (e.g., [5]). Theoretical and experimental models indicate that a plume rising from the deep mantle develops a head approximately 1000–1200 km in diameter during ascent and, on reaching the base of the lithosphere, flattens and spreads laterally to a diameter of approximately 2000–2500 km, corresponding to a radius of about 1000–1250 km [29, 34, 35]. Geological estimates from numerous LIPs yield comparable dimensions for mantle plume heads [28, 36–39].

Here we focus specifically on these large, LIP-generating plume heads, inferred to rise from the deep mantle, because their ~ 1000 -km-radius footprint provides the scale necessary to generate broad lithospheric uplift and horizontal stresses capable of influencing plate-scale tectonics and mineral systems. Smaller upper-mantle upwellings may also generate significant local thermal and tectonic effects, but their more limited dimensions make them less relevant to the regional- to plate-scale plume-push processes considered here.

The arrival of a buoyant mantle plume is associated with domal uplift that in addition to causing topographic and isostatic changes also typically produces a radial mafic dyke swarm, whereas outward spreading of the mantle plume head against the base of the lithosphere is associated with the formation of circumferential mafic dyke swarms (e.g., [27]). Numerous radial mafic dyke swarms have been recognized on Earth and are widely used to identify mantle plume centres [5, 26, 27]. Circumferential mafic swarms are also known, but only about 15 terrestrial examples have been recognized, perhaps because circumferential dykes may be thinner (<10 m) and harder to recognize [40] in contrast to the dykes of major radial swarms having typical individual dyke widths >10 m [41]. However, a much greater number of circumferential systems are expected on Earth by comparison with our neighboring planet Venus (e.g., [27, 40, 42]).

1.3. Links between LIPs, Mineral Systems and Ore Deposits

The links between LIP events and the rocks formed during these events and metallogeny are exemplified by magmatic sulphide systems. Nickel–Cu–platinum group element (PGE) magmatic sulphide deposits are a globally important source of Ni and PGE production; the most significant of them are associated with plume-generated LIPs (e.g., [12, 43–46]). LIPs can also have direct links with hydrothermal and lateritic deposits, as well as indirect links with orogenic and convergent-margin mineralization and broader tectonic controls on the distribution of mineral systems through geological time (e.g., [12, 13, 19, 47–49]).

More specifically, the relationships between LIPs and metallogenic, hydrocarbon and aquifer systems have been summarized into five partially overlapping categories [5, 12, 13, 50, 51], as listed below. (1) LIP-derived magmas may themselves generate economic deposits, such as magmatic Ni–Cu–PGE sulphide systems, including the Norilsk–Talnakh deposits of the Siberian Traps [52, 53], as well as Nb–Ta–REE and diamond mineralization linked to LIP-related carbonatites and kimberlites [5, 51, 54]. (2) LIPs can act as major thermal and geochemical engines, providing heat, fluids and metals that drive the formation of mineral deposit types including native copper, volcanogenic massive sulphide (VMS) and iron oxide–copper–gold (IOCG) deposits [12, 55, 56]. Prominent

IOCG hydrothermal ore systems, such as those of the Gawler Craton, have also been associated with large-scale felsic magmatism (representing a Silicic LIP, SLIP) and major mantle-derived thermal input [57, 58]. The thermal effects of LIP-related magmatism, particularly sill emplacement, can also enhance hydrocarbon maturation and locally produce over-maturation of source rocks [15, 59]. (3) Mafic intrusions related to LIPs—particularly extensive sill and dyke networks—can exert strong structural, hydrological and geochemical controls by acting as fluid barriers, conduits or reactive horizons [5, 12]. These features may localize mineralizing fluids and hydrothermal ore formation, create or modify hydrocarbon traps, and regulate groundwater flow and aquifer connectivity [5, 12, 59]. (4) At Earth's surface, prolonged weathering of LIP lithologies can produce lateritic Ni–Co deposits (although these are more commonly associated with exhumed ultramafic ophiolitic rocks) as well as major bauxite and other Al-rich residual deposits and locally residual Nb–Ta–REE enrichments [12, 60–62]. LIP activity is also closely linked to oceanic anoxic events and the accumulation of organic-rich black shales [16, 17, 63], which host hydrocarbons and are important components of several metalliferous and sediment-hosted Zn–Pb systems [64–66]. (5) Beyond their direct geological effects, LIPs can play an indirect role by promoting continental rifting and plate reorganization and, where associated with changes in plate motion, can influence subduction geometry, slab tearing, and regional stress patterns. These tectonic responses may in turn affect the development and localization of orogenic gold, porphyry, and related mineral systems [47, 48, 67]. LIPs also provide useful time-stratigraphic markers for supercontinent reconstruction, including the timing of continental rifting and breakup, and for correlating crustal and metallogenic provinces between now-separated continents [5, 8, 68, 69].

1.4. Research Goals

This manuscript evaluates how the dynamic evolution of mantle plumes and plume-generated Large Igneous Provinces (LIPs) may enable metallogeny and act as drivers for the formation of mineral deposits. Building on the five broad categories of LIP–mineralization links introduced above, we focus mainly on four: magmatic systems (#1), hydrothermal systems (#2, and #3), and indirect effects transmitted through the plate-tectonic circuit (#5). The aim is to assess how plume-related changes in stress, topography, heat flow, permeability, magma ascent and plate-boundary conditions may generate, enhance, focus or trigger diverse mineral systems. These concepts are summarized in Table 1 and developed below, with the broader goal of expanding the range of commodities and exploration criteria that can be evaluated using the LIP/plume paradigm.

Table 1. Mineral system responses to plume-induced stress reorganization associated with mantle plume emplacement.

Spatial Relationship to Plume	Plume-Related Process	Dominant Stress / Geodynamic Effect	Mineral System Response	Exploration Implications (Speculations)
Proximal (Section 3.1)	Transition from plume-head uplift to lateral plume-head spreading	Rapid reorganization of the regional stress field, involving $\sim 90^\circ$ rotation from radially to circumferential stress regimes	Controls timing, orientation, and emplacement of structurally focused magmatism, including chonoliths (e.g., Norilsk-type systems; Mid-Continent Rift)	Predict location and timing of magma conduits and associated Ni–Cu–PGE magmatic sulphide systems by reconstructing plume center location and pattern of stress evolution
Proximal (Section 3.3)	Domal uplift associated with plume emplacement	Gentle but regionally extensive tilting of continental crust and sedimentary basins	Reorganization of basinal fluid flow, including abrupt changes in flow direction and velocity, facilitating sediment-hosted mineralization (e.g., uranium, sediment-hosted copper; potentially Mississippi Valley-type (MVT) and clastic dominated/sedimentary exhalative (CD/SEDEX))	Identify basin margins and tilting directions relative to plume uplift to predict fluid pathways and redox traps
Proximal (Section 3.4)	Underlying plume head providing thermal input, structural input and magmatic input	Magma locally ascending along translithospheric zones above a mantle plume	Sed-hosted mineralization, orthomagmatic sulphides	Distribution of underlying plume head, comparison with lithospheric thickness, and location of translithospheric faults
Proximal (Section 3.5)	Development of plume-induced radial and circumferential fracture networks	Formation of crustal-to lithospheric-scale extensional fracture systems that act as first-order permeability pathways	Structural focusing of hydrothermal vein systems and related mineralization within and surrounding LIPs	Target hydrothermal, hydrothermal-magmatic and magmatic systems by mapping plume-related fracture geometries and their interaction with inherited structures

Table 1. (Continued).

Spatial Relationship to Plume	Plume-Related Process	Dominant Stress / Geodynamic Effect	Mineral System Response	Exploration Implications (Speculations)
Distal (Section 4)	Plume-head push and plume-related plate reorganization	Outward forces exerted by a spreading plume head against thick lithosphere modify intraplate and plate-boundary stress regimes; may induce rapid changes in plate motion	Reactivation of inherited structures and transient permeability creation in orogenic and arc settings, triggering mineral system activation, most clearly for orogenic gold and plausibly for porphyry copper	Assess temporal coincidence between plume emplacement and mineralization using plate reconstructions; target structurally prepared belts (orogenic belts) undergoing plume-driven stress perturbations

Notes: Given our focus on deep mantle plumes which cause LIPs, Proximal is approximately ≤ 1250 km away from the plume centre; Distal is approximately >1250 km and up to 20,000 km via plate scale stress transmission.

2. Methods

This is a literature-based synthesis that develops a conceptual framework for evaluating how mantle plume dynamics and plume-generated Large Igneous Provinces (LIPs) may influence diverse mineral systems. We compile and compare published information on LIPs, reconstructed plume centres, giant mafic dyke swarms, plume-head dimensions, plume-related uplift, stress-field evolution, plate reorganization, lithospheric structure, with selected ore-deposit types.

2.1. Conceptual and Stress-Field Framework

An important premise of our analysis is that geologically rapid changes in regional stress can exert a major control on the development and localization of mineral systems. Such changes can influence magma ascent and emplacement, creation and reactivation of permeability, hydrothermal-fluid flow, and the development of structural traps. Their importance of rapid stress changes has been recognized in a variety of mineral systems, including magmatic Ni–Cu–PGE, orogenic gold, and porphyry Cu systems (e.g., [70–74]). For example, the localization of magmatic Ni–Cu–PGE systems is strongly influenced by lithospheric and regional geodynamic architecture [71, 75], whereas structural studies of orogenic gold systems demonstrate the importance of changes in regional stress regime in controlling fault reactivation, fluid focusing, and mineralization (e.g., [72]).

Most of the stress changes outlined above are conventionally interpreted within the framework of plate tectonics. Orogenic and convergent-margin mineral systems, for example, have been linked to changes in subduction geometry and dynamics, including slab rollback, slab tearing and development of slab windows, and slab break-off.

These processes can reorganize regional stress fields and deformation patterns, modify mantle and crustal thermal regimes, promote magmatism, and influence the generation and focusing of hydrothermal fluids (e.g., [76–80]). For instance, Menant et al. [78] demonstrated links between three-dimensional subduction dynamics and changes in the distribution and character of Cu- and Au-rich mineralization, with rapid slab rollback and lateral slab-tearing favouring the development of Au-rich systems.

Here we evaluate an additional, and potentially widespread, control: rapid changes in stress, topography, heat flow, and lithospheric dynamics associated with the arrival and subsequent evolution of a mantle plume as part of a LIP event (Table 1). We extend this general concept by asking whether stress perturbations associated with plume-generated LIPs can generate, enhance, focus, or trigger processes already known to be important in a broad range of mineral systems.

For this purpose, we distinguish three successive components of plume–lithosphere interaction (Figure 1): (1) arrival of the plume head at the base of the lithosphere, flattening, and associated domal uplift; (2) active lateral spreading of plume-head material beneath the lithosphere; and (3) outward transmission of plume-related stresses, including the plume-push mechanism proposed here, particularly where laterally spreading plume material encounters thick cratonic lithospheric roots. Giant mafic dyke-swarm geometries (associated with LIPs) provide one means of reconstructing these processes. Radiating mafic dyke swarms commonly diverge from inferred plume centres and record regional extension associated with plume arrival and uplift, whereas giant circumferential mafic dyke swarms may record stresses associated with the subsequent lateral spreading of plume heads [27, 37, 38, 40].

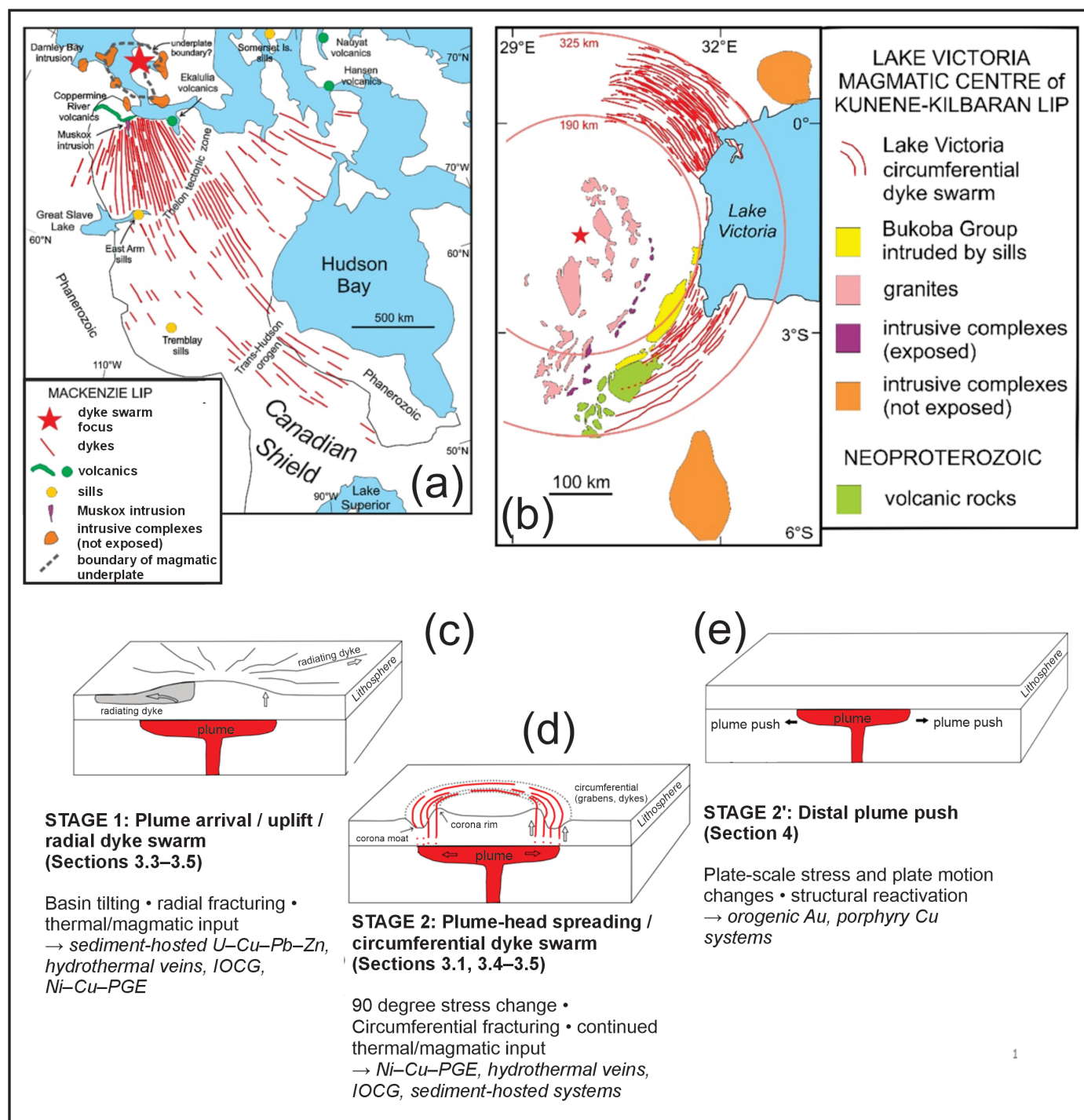


Figure 1. Radiating and circumferential dyke swarms of plume-generated LIPs. Including both schematic (parts **c**, **d**, **e**) and examples (parts **a**, **b**). **(a)** 1270 Ma Mackenzie radial swarm and **(b)** 1380 Ma Lake Victoria circumferential swarm, respectively. Stage 1 (parts **a**, **c**) is associated with arrival of a mantle plume causing domal uplift and development of a radial mafic dyke swarm. Stage 2 (parts **b**, **d**) is associated with spreading of the plume head and development of a circumferential mafic dyke swarm. The reference to “corona” relates to analogous features on Venus [27]. Outward “plume push” also occurs during Stage 2 (part **e**). part **(a)** modified after [27] and part **(b)** modified after [81], with the addition of the Bukoba Group (after [82]) and geophysically-recognized intrusive complexes (after [83]). The maximum horizontal stress vectors parallel the trend of the dykes: radial in the case of the radial dyke swarms and circumferential in the case of the circumferential swarms.

2.2. Spatial Framework

As a first-order spatial framework, we use idealized circular plume-head footprints based on starting-plume models [29, 35]. In this model, a large plume head rising through the mantle flattens beneath the lithosphere to form a disk approximately 2000–2500 km in diameter, corresponding to a radius of approximately 1000–1250 km. This provides a practical first-order estimate for regions potentially affected directly by plume-head dynamics, while recognizing that natural plume heads may be irregular or asymmetric because of lithospheric-thickness variations, lithosphere–asthenosphere boundary topography, cratonic roots and plate motion relative to the plume (e.g., [84–87]).

We distinguish two broad spatial domains of mantle plume influence (Table 1). The first comprises regions directly above or proximal to the plume head, where plume dynamics may influence mineralization through domal uplift and associated topographic gradients, changing regional stress fields, radial and circumferential fracturing, thermal input, mafic magmatism, and resulting modification of crustal permeability, and perturbation of sedimentary-basin hydrology.

The second comprises regions beyond the reconstructed plume-head footprint, where stresses generated by lateral plume-head spreading may be transmitted through the lithosphere and, potentially, through the broader plate-tectonic circuit. In this distal domain, plume-related changes in intraplate stress and plate motion may influence mineral systems developed at remote convergent or transpressional plate margins, including orogenic gold and potentially porphyry Cu systems.

3. Proximal Stress and Topographic Changes (Above and Near the Plume Head) and Links to Ore Deposits

Stress and topographic changes associated with mantle plumes are a potential trigger for a range of mineral deposit types (Table 1): (1) rapid $\sim 90^\circ$ stress transitions from uplift to lateral spreading, facilitating emplacement of chonolith-hosted magmatic sulphide systems (Section 3.1); (2) regional uplift and radial and circumferential fracturing controlling hydrothermal vein orientations (Section 3.5); (3) basin tilting that redirects mineralized brine flow, contributing to sediment-hosted mineralization (Section 3.3); and (4) the presence of an underlying mantle plume providing thermal, magmatic and structural input (Section 3.4).

3.1. Formation of Chonoliths Associated with the 90-degree Stress Change

Role of plume-generated stress changes in magmatic sulphide mineralization. Most genetic models for magmatic

Ni–Cu–PGE sulphide deposits require an anomalously hot mantle source, commonly attributed to the presence of a mantle plume, capable of generating large volumes of primitive, high-Mg mafic to ultramafic magma enriched in Ni, Cu and platinum-group elements through extensive partial melting. Ore formation then requires sulphide saturation, commonly promoted by crustal contamination or external sulfur addition, followed by segregation and accumulation of immiscible sulphide liquid within magma conduits, embayments, chonoliths or feeder systems, all of which often require the involvement of dynamic magma systems [45, 52, 88, 89]. These plume-generated “high-temperature” and “very high temperature” Ni deposits [90] are commonly associated with LIPs and with extensional tectonic settings that facilitate ascent and emplacement of mafic–ultramafic magmas. Holwell et al. [90] also recognize lower-temperature Ni–Cu deposits associated with orogenic settings, but these are outside the scope of the present discussion, which focuses on plume-related magmatic sulphide systems.

Our particular focus is on magmatic chonoliths, irregular tube- to blade-shaped mafic-ultramafic intrusions that form within dynamic magma-conduit systems and are important targets for magmatic sulphide Ni–Cu–PGE deposits (e.g., [89]). They develop where repeated magma pulses exploit faults, lithological contacts, and other structural weaknesses, with inflation, brecciation, erosion, and replacement of country rock producing highly irregular conduit geometries. Such intrusions have also been described using terms such as “channelized intrusions” [89, 91], “tube-like conduits” [92], “bladed dykes” [93], and “conduit-type intrusions” [56].

Sulphide saturation may result from crustal contamination, magma mixing, cooling, or volatile loss. Dense sulphide melt can accumulate in widened conduit segments and basal traps, while continued magma throughput promotes interaction with large volumes of silicate melt, enriching the sulphides in Ni, Cu, and the PGE (e.g., [94, 95]).

Previous research has explored the role of plume-generated stress changes associated with mineralized chonoliths [73, 96], as summarized below. We propose that the timing and location of mineralization within these systems is related to the stress transition caused by the underlying mantle plume and associated with a change from radiating (associated with plume-caused domal uplift) to circumferential (associated with outward spreading of the plume head beneath the lithosphere). Here, we describe two case studies: Norilsk mineralization associated with the 252 Ma Siberian Trap LIP [73] (Figure 2a) and mineralization of the ca. 1111–1085 Ma Keweenaw LIP of the Mid-Continent Rift (MCR) of North America (Figures 2b) [96].

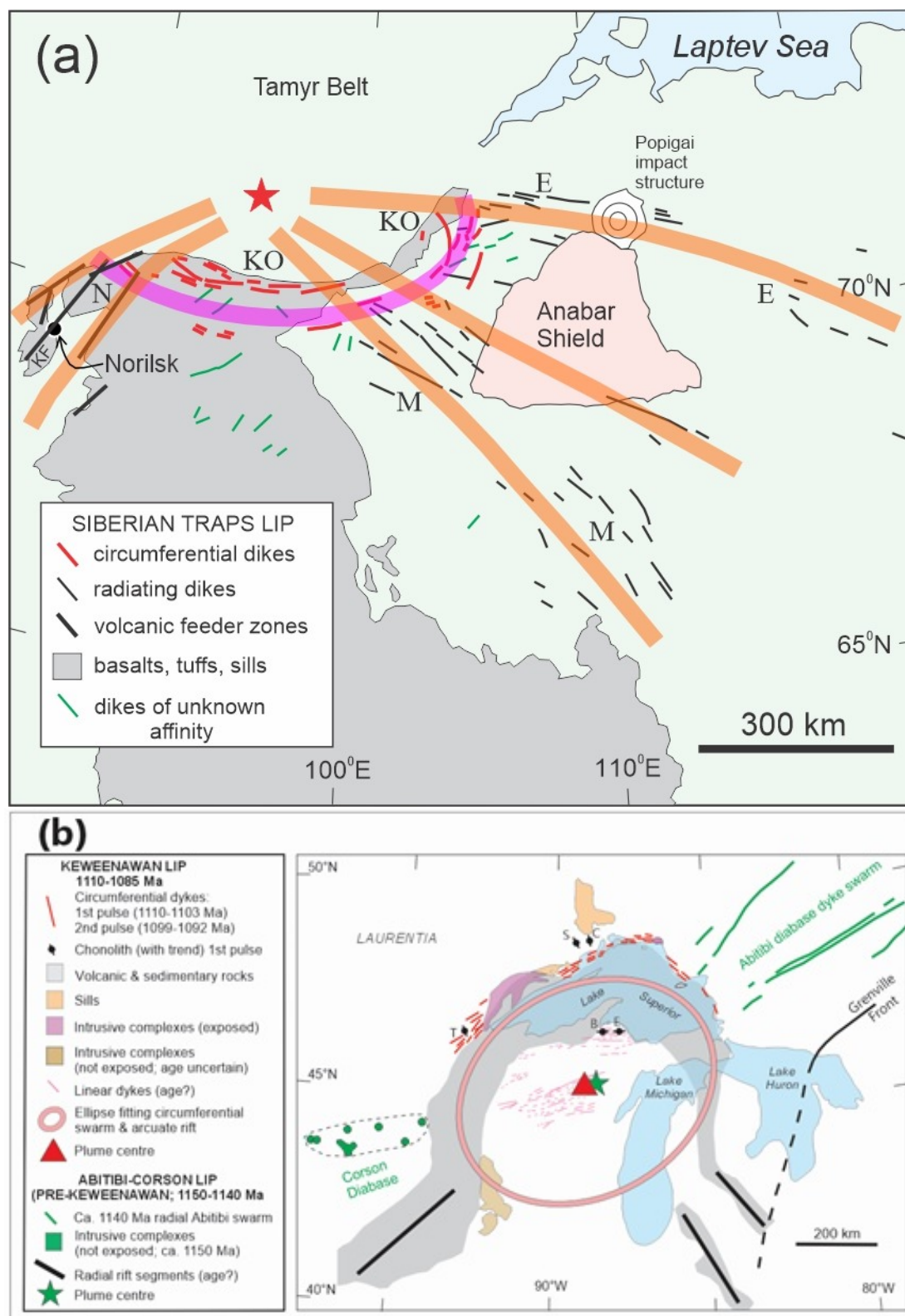


Figure 2. Stress changes associated with the emplacement of mineralized chonoliths. **(a)** Norilsk chonoliths of the 252 Ma Siberian Traps LIP [73]. **(b)** northern chonoliths associated with the arcuate (1110–1092 Ma circumferential swarm) that follows the arcuate the Mid-Continent Rift (MCR) of the Great Lakes region of North America. This figure also identifies a precursor stage featuring a ca. 1140–1150 Ma radial system in addition to the subsequent 1110–1092 Ma circumferential stress system (modified after [27, 96]), and with both sharing the same plume center.

3.1.1. Stress Change Origin of the Norilsk Mineralized Chonoliths of the Siberian Traps LIP

The 252 Ma Siberian Trap LIP (Figure 2a) is one of the most significant Phanerozoic LIPs given its link with the end Permian mass extinction (e.g., [97]). It is also important for its associated world-class magmatic sulphide deposits in the Norilsk region (e.g., [53]). A $\sim 90^\circ$ change in stress orientation associated with distant plate boundary stress changes was hypothesized as a trigger for emplacement of Norilsk chonolith mineralization [71]. We developed an alternative interpretation (Figure 2a) [73] where the emplacement of the mineralized Norilsk chonoliths occurred during a 90° change in regional stress caused by the transition from a radial horizontal sigma 1 stress (associated with domal uplift above the plume) to a circumferential horizontal sigma 1 stress (associated with spreading of the plume head) as revealed by the transition from a regional radial to circumferential mafic dyke swarm (Figure 2a). This implies an intervening period in which the stress is approximately isotropic (with low horizontal deviatoric stress), a period associated with emplacement of the Norilsk-Talnakh mineralization. The stress-drop during the transition period may have led to release of volatiles that allowed ascent and/or lateral emplacement of gas-buoyed magmatic sulphides (e.g., [94, 95, 98]). However, the specifics of the mechanism for mobilizing the emplacements are not constrained by our model. Also, while our model does not negate the Begg et al. [71] interpretation of external plate boundary stresses, it has the advantage of offering an explanation directly linked to the known dynamics of the underlying Siberian Traps mantle plume.

3.1.2. Stress Change Origin of Mineralized Chonoliths of the Keweenaw LIP of the Mid-Continent Rift

The >2000-km-long arcuate Midcontinent Rift of Laurentia (Figure 2b) records emplacement of the ca. 1111–1085 Ma Keweenaw LIP, centred on the Lake Superior region and dominated by voluminous tholeiitic basalts, mafic–ultramafic intrusions, dyke swarms, and subordinate felsic and carbonatitic magmatism [56, 92, 96, 99, 100]. Mineralized chonoliths are present both on the northern and southern sides of the arcuate Mid-Continent rift (MCR) (that follows Lake Superior); and here we focus on the origin of those on the northern side [56, 92]. This Keweenaw event was preceded by ca. precursor pulse of 1150–1140 Ma magmatism, including the Abitibi dyke swarm, lamprophyres, and the Corson diabase intrusions (Figure 2b).

Ernst et al. [96] investigated mantle-plume-generated stresses recorded by dyke swarm orientations associated with the 1111–1085 Ma Keweenaw LIP/Midcontinent Rift (MCR) and the 30 myr earlier 1140 Ma Abitibi (-Corson) event (Figure 2b). The Abitibi (-Corson) event includes a radial stress pattern, expressed by diabase and lamprophyre dykes that converge south of Lake Superior and define the plume center. We interpret that the

same plume centre was utilized during the subsequent Keweenaw plume/MCR activity with two main magmatic pulses at 1110–1103 and 1099–1092 Ma, both occurring during a circumferential sigma 1 stress time, revealed by an arcuate mafic dyke swarm located along the northern shore of Lake Superior. The first of the two arcuate dyke swarm pulses is associated with the mineralized chonoliths, and the latter pulse associated with 80 km rift opening of the Mid-Continent Rift [27, 96].

Mineralized chonoliths at Tamarack, Sunday Lake, and Current Lake, emplaced outside the circumferential dykes on the northern side of Lake Superior during the first Keweenaw pulse, have “wiggly” tails that are radially oriented and therefore approximately orthogonal to both the MCR and the circumferential swarm (Figure 2b). Ernst et al. [96] interpret these tails as recording local outward emplacement of magma from a circumferential dyke into a radial extensional fracture generated immediately outside the circumferential dyke as its opening produced circumferential crustal widening. In this model, magma rising through a circumferential dyke was locally diverted outward into these newly formed radial extension zones, where repeated magma flow promoted development of chonolithic conduits. This mechanism explains why radial-tailed chonoliths developed during the first Keweenaw pulse, but not the second, when opening of the MCR itself accommodated much of the required circumferential crustal widening. Chonoliths south of Lake Superior are not explained by this model and remain enigmatic in other interpretations as well [96].

3.1.3. Assessing Other Chonoliths

The proposed relationship between plume-generated stress field changes and the emplacement of mineralized chonoliths—particularly the proposed importance of circumferential stresses in both the Norilsk chonoliths of the Siberian Traps LIP [73] and the northern Keweenaw chonoliths [96]—suggests several paths for further study. These include evaluating other documented suites of mineralized chonoliths (e.g., [94]) for evidence of emplacement during plume-related stress reorganization and examining additional LIPs that contain circumferential dyke swarms (e.g., [40]) for a potential chonolith association.

3.2. Sediment-hosted and Fracture-controlled Ore Deposits: Potential Role of Mantle Plumes

Sediment-hosted and fracture-controlled hydrothermal mineral systems commonly overlap in terms of the general processes involved in the formation of these systems. Sedimentary basins may provide the fluid reservoirs, aquifers and reactive stratigraphic traps, whereas faults, fractures, breccias, dyke margins and basement-rooted structures provide focused permeability and sites of fluid discharge. These systems are exemplified by clastic-dominated (also known as sedimentary exhalative,

SEDEX) Zn–Pb–Ag, carbonate-hosted Mississippi Valley-type (MVT) Pb–Zn, sediment-hosted Cu, sedimentary and unconformity-related U, bedded barite, phosphate, Fe–Mn, evaporite-related and five-element Ag–Co–Ni–Bi–As mineralization [55, 101–112]. Despite their diversity, these systems commonly require suitable sources of metals and ligands, an effective hydraulic or thermal driver, connected fluid pathways, and structural, stratigraphic or chemical traps. Saline basinal brines are particularly important because they can leach and transport metals over substantial distances [108, 113–115]. Their circulation may be driven by burial compaction, topographic relief, tectonic deformation, thermal convection, density contrasts and transient overpressure associated with fault reactivation or seal breaching [116–120].

Fracture-controlled vein, stockwork and breccia systems similarly depend on focused and commonly transient permeability. Fault slip, hydraulic fracturing, brecciation and repeated fracture opening and sealing can generate short-lived fluid pathways, with fault-valve behaviour, seismic pumping and pressure fluctuations playing important roles in many hydrothermal systems [121–124]. These processes are relevant to orogenic and epithermal Au, polymetallic Ag–Pb–Zn–Cu, fluorite–barite $\pm$ Pb–Zn, U–Ag–Co and five-element Ag–Co–Ni–Bi–As veins, as well as fracture-, stockwork- and breccia-hosted components of IOCG, porphyry and skarn systems [19, 102, 111, 125].

For the present study, the key point is that both basin-hosted and fracture-controlled mineral systems can be highly sensitive to changes in topographic and hydraulic gradients, heat flow, regional stress and crustal permeability. These are precisely the parameters that may be rapidly perturbed during the arrival, uplift, lateral spreading and broader tectonic influence of an underlying mantle plume.

We consider three complementary ways in which mantle-plume dynamics may influence basin-hosted and

fracture-controlled mineralization:

1. plume-related domal uplift and basin tilting, which may generate regional hydraulic-head gradients and redirect basinal brine flow (Figures 3–5; Section 3.3);
2. thermal, magmatic and potentially fluid input from the underlying plume and associated LIP magmatism, particularly where focused along trans-lithospheric structures (Figure 6; Section 3.4); and
3. radial and circumferential fracture systems generated during plume-related uplift and plume-head spreading, which may provide pathways and depositional sites for hydrothermal vein mineralization (Figures 7 and 8; Section 3.5).

3.3. Plume-related Domal Uplift and Basin Tilting

Regional uplift is a well-established driver of basin-scale fluid flow and has been invoked in models for MVT Pb–Zn mineralization and, more tentatively, in sediment-hosted Cu–Co systems (e.g., [117, 126, 127]). In most published examples, the topographic gradient is attributed to orogenic uplift (Figure 5). We propose an alternative mode where plume-related domal uplift may provide a comparable hydrologic driver to mineralizing events in selected sedimentary basins.

Mantle plumes ascending beneath continental lithosphere may produce broad domal uplift of as much as approximately 1 km across regions approaching 1000–1250 km in radius (e.g., [34]). Sedimentary basins within this uplifted region would be tilted away from the plume centre. Although the resulting stratal dips may be less than 1°, even low-angle tilting may substantially influence fluid migration where permeability is strongly anisotropic and flow is preferentially channelled through bedding-parallel aquifers [115].

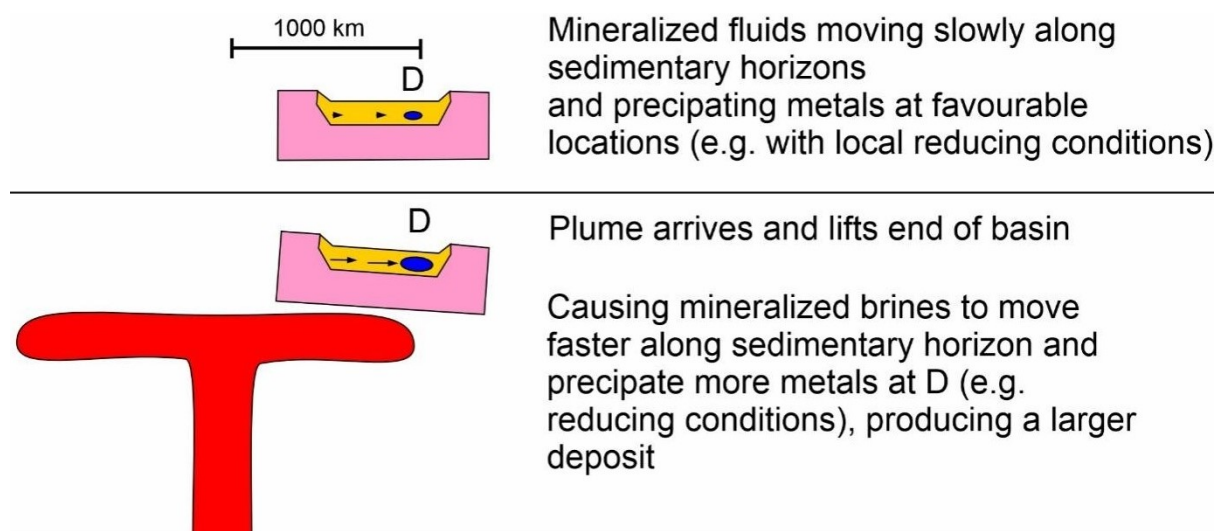


Figure 3. Model of plume head uplift causing tilting of a sedimentary basin, causing change in flow of mineralized brines and lead to new pulses of mineralization of a wide variety of sediment hosted ore types including U, Cu, and Pb–Zn (Table 1).

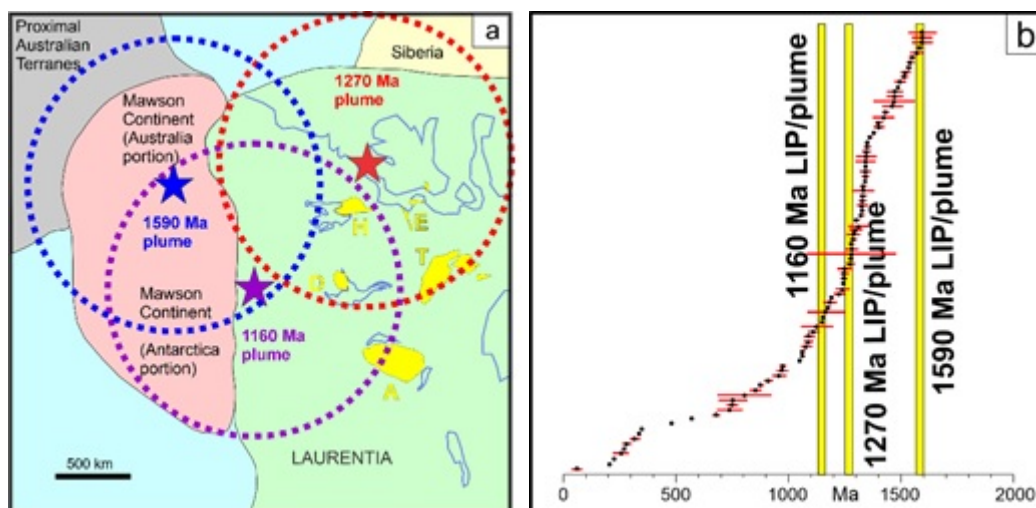


Figure 4. Proposed contribution of domal uplift associated with three mantle plumes to uranium mineralization in the Athabasca basin of northern Canada. (a) Paleogeographic reconstruction of Laurentia and Australia. The Mawson continent includes both the Gawler craton of south Australia and parts of Antarctica. Three Mesoproterozoic mantle plume centre (at 1590, 1270 and 1160 Ma) are shown with a star locating each plume centre and a dashed circle of radius 1000 km indicating the extent of the plume head. Herein, we propose that uplift caused by these plume head contributed to the timing of uranium mineralization associated with the Athabasca Basin (A) in northern Canada. Blue outlines in Laurentia outline the lakes (and Arctic Coast and Victoria Island). Base diagram modified after [128]. The Athabasca (A) and other basins (T = Thelon, H = Hornby Bay, E = Elu and D = Dessert Lake; added after [129]) are correlated in age and are interpreted as preserved remnants of a formerly much more extensive regional intracratonic sedimentary system [129–132]. (b) A compilation of U–Pb ages of uraninite and Ar–Ar ages of illite associated with U mineralization in the Athabasca Basin (after [133], modified after [134]) with the timing of the three LIP events shown as labelled vertical bars.

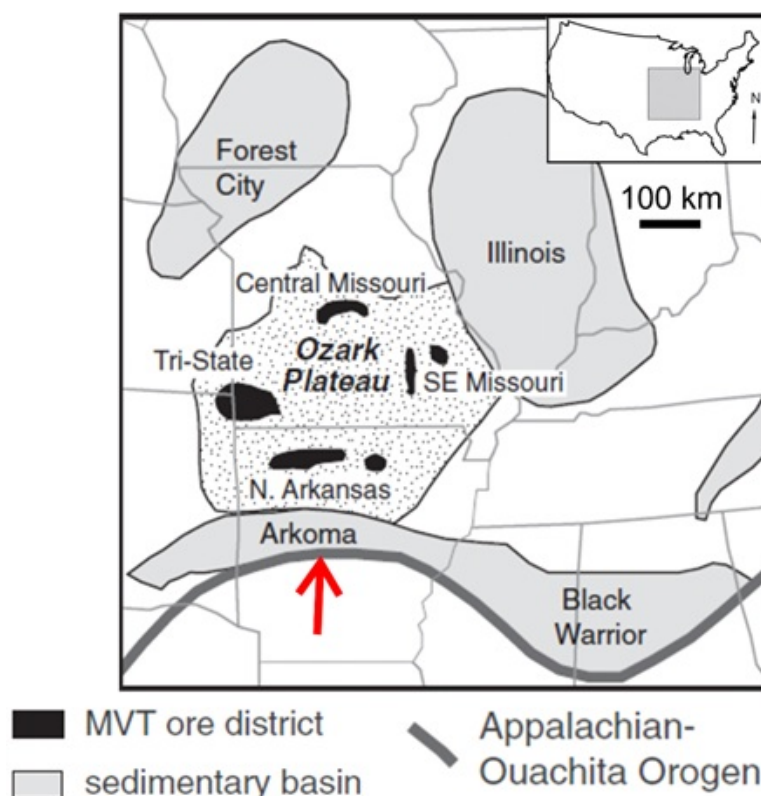


Figure 5. Uplift-driven MVT fluid migration in the Ozark Plateau, United States. Stoffell et al. [126] linked Ozark Plateau MVT mineralization to early Permian Ouachita-related uplift and tilting of the Arkoma foreland. Red arrow added to show downhill direction off the (orogenic) uplift.

Our hypothesis involves three linked effects. First, plume-related doming tilts sedimentary successions and establishes a regional hydraulic-head gradient. Second, this gradient increases or redirects the migration of saline, potentially metal-bearing brines down-dip through connected aquifers. Third, mineralization is localized where these migrating fluids encounter favourable faults, fracture zones, aquifer pinch-outs, compartment boundaries, reactive stratigraphic units or redox traps (some of which can also be generated by plume-related fracturing described in Section 3.5, or linked to lithospheric penetration by the underlying plume, Section 3.4). As noted above, this plume-caused basin tilting mechanism could potentially contribute to a range of sediment-hosted systems, including unconformity-related and sandstone-hosted U, sediment-hosted Cu, and MVT or other carbonate-hosted Pb–Zn deposits.

3.3.1. Uranium Mineralization in the Athabasca Basin, Northern Canada

The unconformity-related uranium deposits of the Athabasca Basin (Figure 4) constitute one of the world's major uranium provinces and have produced more than 500,000 t of uranium since the first discoveries in the late 1960s [133]. We hypothesize that the arrival of three mantle plumes sequentially uplifted and tilted the originally more extensive Athabasca sedimentary basin, contributing to the formation of multiple pulses of uranium mineralization.

A compilation of uranium-mineralization ages (Figure 4b) shows a broad and nearly continuous spectrum from approximately 1600 to 1100 Ma, followed by lower-frequency younger ages extending to approximately 50 Ma [133, 134]. An age of approximately 1590 Ma has been interpreted to represent the principal basin-wide uranium-mineralization event [135], although evidence for hydrothermal activity as early as approximately 1640 Ma suggests that development of the mineralizing system may have begun substantially earlier [134].

Chi et al. [134] also noted several regional magmatic events that may have provided thermal perturbations capable of influencing uranium mineralization, including ca. 1540 Ma basalts in the Kuungmi Formation of the Thelon Basin [136], the ca. 1267 Ma Mackenzie dykes [137], the ca. 1165 Ma Douglas River dyke [138], and the ca. 1109 Ma Moore Lakes intrusion [139]. Direct thermal effects of this magmatism generated by the underlying plume may have contributed to hydrothermal circulation, as per the model discussed below in Section 3.4. Here, however, we offer an alternative (and likely complementary) mechanism for affecting flow of mineralization basin flows, by considering plume-generated domal uplift and basin tilting by three plume-generated LIPs, whose plume centers are known and with inferred plume head radius of about 1000 km (Figure 4):

1. a ca. 1590–1550 Ma plume centre located in southern Australia when reconstructed adjacent to western Laurentia [128];

2. the ca. 1270 Ma Mackenzie plume centre [140]; and
3. a more speculative ca. 1160 Ma plume centre inferred from the westward convergence of three major mafic dykes [134, 138].

We propose that domal uplift associated with these three events sequentially tilted the broader sedimentary succession that originally included the Athabasca Basin remnant and thereby triggered or enhanced episodes of regional brine migration and uranium mineralization. This interpretation requires two principal assumptions. First, the Athabasca and Thelon successions were originally parts of a larger, interconnected intracratonic basin, consistent with paleocurrent evidence produced by [141] and discussed in [132]. Second, plume-generated hydraulic gradients could drive brine migration to the edge of uplift (marking the plume edge) (Figure 4) and even somewhat beyond through laterally connected aquifers.

We suggest that uranium mineralization in the Athabasca Basin includes three principal episodes centred broadly at ca. 1590–1550, 1270 and 1160 Ma, corresponding to the times of greatest plume/LIP-related uplift. The apparent continuity of mineralization ages rather than the presence of discrete pulses at these three times may reflect: (1) analytical uncertainties and partial resetting of mineral chronometers [133–135, 142]; (2) initiation of dynamic uplift by the ascending plume before the main LIP magmatic pulse [34, 35, 143, 144]; (3) gradual decay of uplift over several to tens of millions of years as the plume head spread, cooled or migrated laterally (e.g., [144–146]); (4) longer-lived uplift supported by lower-crustal magmatic underplating [5, 147]; and (5) later eclogitization and delamination of dense lower-crustal underplates, allowing renewed asthenospheric upwelling and additional uplift, as proposed in a plume/LIP context by [31] and supported more generally by models of lower-crustal instability and delamination [148–150].

A testable prediction of the plume-tilting hypothesis concerns changes in paleoflow direction. If the inferred plume centres are approximately correct, mineralizing-fluid flow may have been directed initially from the northwest during the ca. 1590–1550 Ma event, from the north during the ca. 1270 Ma Mackenzie event, and from the west-northwest during the ca. 1160 Ma event. Independent reconstructions of paleohydrology and fluid-flow directions could therefore provide a critical test of this model.

3.3.2. Mississippi Valley-type Pb–Zn Deposits

The Tri-State and Northern Arkansas districts of the Ozark Plateau are classic examples of MVT mineralization formed by continent-scale basinal-brine migration associated with uplift of the Arkoma foreland basin during the Early Permian Ouachita orogeny [117, 126]. Although basin tilting in this example (Figure 5) resulted from orogenic rather than plume-related uplift, it demonstrates that regional topographic gradients can reorganize basin-scale brine migration and contribute to MVT mineralization.

This raises the possibility that plume-related domal uplift, perhaps coupled with enhanced fluid circulation driven by LIP-related increases in geothermal gradients, could have hydrologically preconditioned some MVT systems. Such preconditioning could include hydrothermal dolomitization prior to mineralization, potentially enhancing permeability and establishing fluid pathways subsequently exploited by ore-forming brines (e.g., [151]). In addition, the reactivation or formation of faults by LIP-related uplift can also enable fluid flow, suggesting further potential links with MVT systems (e.g., [152]).

Testing this hypothesis requires deposit-specific reconstruction of mineralization ages, basin architecture, paleohydrology, uplift history and plume-centre geometry, all of which are commonly incomplete or difficult to resolve in ancient basins. A systematic assessment of additional MVT districts represents an important direction for subsequent case-history research.

3.3.3. Exploration Implications of the Uplift and Basin-Tilting Model

Building on the Athabasca uranium example discussed above, together with evidence from MVT Pb–Zn systems that regional uplift and basin tilting can redirect mineralizing brines, we suggest that analogous plume-induced uplift should be evaluated as a potential control on sediment-hosted mineralization more broadly (Figure 6). Within this framework, plume-related uplift may provide a first-order basis for predicting regional brine-flow directions, whereas basin architecture and the distribution of favourable stratigraphic, structural and geochemical traps would control the final locations of mineralization. Particularly prospective settings may occur where predicted down-gradient flow paths intersect basin-margin faults, aquifer pinch-outs, compartment boundaries, redox interfaces or reactive stratigraphic units.

3.4. Thermal, Magmatic and Fluid Effects of the Underlying Plume

The preceding Section 3.3 considered the indirect hydrologic consequences of plume-related surface uplift. A distinct but potentially complementary mechanism involves thermal, magmatic and possibly fluid input from the underlying plume and associated LIP magmatism (Figure 6).

This general relationship has been considered in previous studies linking mantle plumes and intracontinental rifting with elevated heat flow, lithospheric weakening, mafic magmatism, and the development of a broad range of hydrothermal and sediment-hosted mineral systems, including SEDEX, stratiform, IOCG, porphyry and related deposits [19, 20, 153]. Here, we build on this earlier work by considering how the geometry of an impinging and laterally spreading plume head, together with lithospheric thickness variations and inherited structures, may control where plume-derived heat, melts and fluids are preferentially transferred into the overlying crust and sedimentary basins.

Plume heads impinging on continental lithosphere can promote localized thinning or erosion of the lithosphere, mafic underplating and transfer of mantle-derived melts into the crust, particularly where rifts or inherited lithosphere-scale structures provide favourable pathways [19, 33, 145]. Such pathways may occur far from the inferred plume centre, especially where plume material migrates laterally along an inclined lithosphere–asthenosphere boundary and is focused beneath thinner or weaker lithosphere [154]. This process is widely invoked for high-degree partial melting and Ni–Cu–PGE mineralization along craton margins and lithospheric boundaries (e.g., [75, 89, 155]), and similar controls may determine where plume-derived heat and melts are transferred into sedimentary basins.

These effects differ from the plume-uplift/basin-tilting model (Section 3.3) by providing localized heat and deep structural connectivity rather than primarily affecting the regional hydraulic gradient. We first consider the example of the 1590 Ma mantle plume and its link with economically important Olympic Dam IOCG deposits.

3.4.1. Link between the 1590 Ma Plume and the Olympic Dam IOCG Deposit

Olympic Dam is a giant iron oxide–copper–gold–uranium (IOCG-U) deposit in the Gawler Craton of South Australia and one of the world's largest Cu–U–Au ore systems (Figure 7). It is hosted by the ca. 1594 Ma Roxby Downs Granite and formed within the broader ca. 1595–1590 Ma Gawler Range–Hiltaba magmatic and hydrothermal event [156, 157]. The Gawler Range Volcanics and associated Hiltaba Suite intrusions define a felsic-dominated large igneous province (a silicic LIP, SLIP) that has been interpreted in terms of mantle upwelling and major mantle-derived thermal input [57, 128].

A plume interpretation is supported by the reconstruction of [128], who correlated ca. 1590 Ma mafic dyke swarms in western Laurentia with contemporaneous magmatism in South Australia and reconstructed a common plume centre between Laurentia and the South Australian Craton (Figures 4 and 7). In this model, plume-related mafic underplating and heating generated a major crustal thermal anomaly, contributing to extensive felsic magmatism and hydrothermal circulation in both regions, including the Olympic Dam deposits in South Australia (and Wernecke breccia systems in Laurentia). Thus, the plume need not have been the direct source of the ore metals; rather, it may have supplied the heat and magmatic input required to establish and sustain the crustal-scale hydrothermal system.

Further support for a plume-head origin of the Gawler LIP comes from high-precision geochronology summarized and expanded by [156]. Integrating earlier age constraints from the Olympic Dam region, including those of [157], with new CA-ID-TIMS zircon ages across the province, they identified a systematic outward younging of magmatism (from ca. 1595–1586 Ma) away from the

Olympic Dam region [156]. This spatial–temporal progression is consistent with a plume head rising beneath the lithosphere and then spreading laterally outward, with magmatism initiated near the plume centre and subsequently migrating toward the margins of the province. The location of Olympic Dam close to the central and ear-

liest part of this ca. 1590 Ma magmatic system therefore provides a plausible example of a proximal mineral system in which an underlying mantle plume supplied the thermal and magmatic energy that promoted crustal melting, hydrothermal circulation, and IOCG mineralization [57, 128, 156].

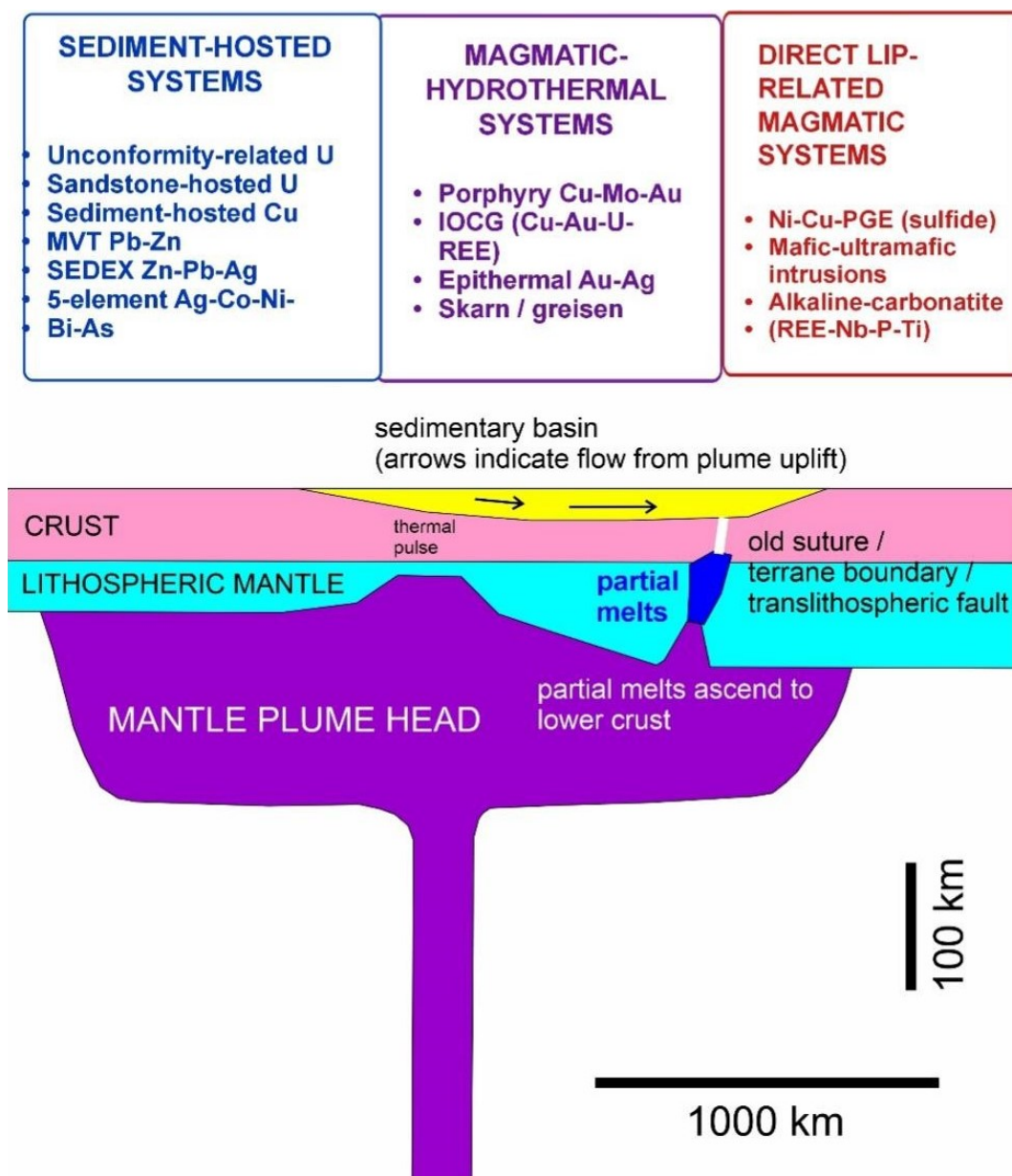


Figure 6. Schematic relationship between underlying mantle plume arriving beneath a variably thick lithosphere and the distribution of thermal pulses affecting mobilization of basinal fluids. As discussed in Section 3.3, plume-related doming and low-angle tilting can generate regional hydraulic-head gradients and enhance basinal brine migration through connected aquifers. At the same time, thermal pulses may affect portions of sedimentary basins where the plume reaches a shallower level owing to locally much thinner lithosphere. Also plume-derived melts, heat and stress may be focused upward along old sutures, terrane boundaries and other reactivated translithospheric structures, promoting mafic underplating, lower-crustal partial melting, fault reactivation and enhanced permeability. Mineralization is localized where these regional drivers interact with conventional ore controls, including redox fronts, reduced sulfur, reactive carbonate units, aquifer pinch-outs, compartment seals, and basement-rooted faults. Potentially affected systems include sediment-hosted U, Cu, Pb–Zn, Zn–Pb–Ag and five-element Ag–Co–Ni–Bi–As deposits, as well as porphyry, IOCG, epithermal, skarn, alkaline–carbonatite and Ni–Cu–PGE systems (Table 1).

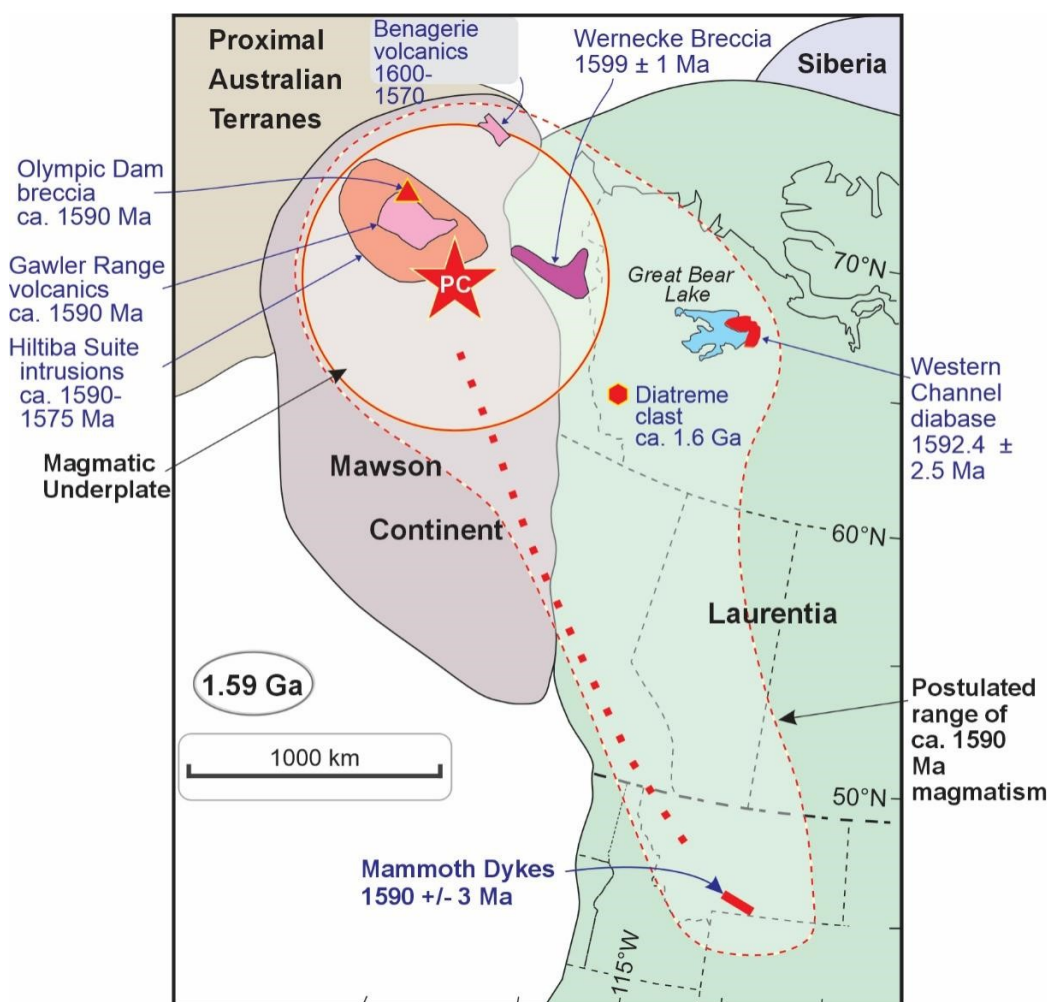


Figure 7. The spatial link between the 1590 Ma mantle plume and the Olympic Dam IOCG deposit in Gawler craton (South Australian craton) (modified after [128]). Note that the circle size radius is 500 km and it covers the central region of the mantle plume likely associated with the magmatic underplate [158]. The overall plume head size is inferred to be twice that radius (1000 km; see Figure 4). Note that according to [156] the plume centre would be closer to the Olympic Dam deposit.

3.5. Hydrothermal Vein Deposits Linked to Radial and Circumferential Fractures Associated with Plume-Head Dynamics

The preceding section considered plume-related effects on fluid flow within sedimentary basins. A third mechanism concerns fracture-controlled hydrothermal mineralization associated with the stress fields generated during plume-related domal uplift and subsequent plume-head spreading stages (cf. Figure 1). Domal uplift above a mantle plume can generate radial extensional fractures, whereas spreading of the plume head may produce circumferential / annular fractures (Figure 1). Many such fractures are filled by plume-derived partial melts and preserved as radial or circumferential dyke swarms. However, not all fractures are necessarily invaded by magma. Unfilled or incompletely filled fractures may remain available as high-permeability pathways for hydrocarbons, basinal brines and hydrothermal fluids and may therefore localize hydrothermal vein systems. We first illustrate this mechanism using two examples (Figures 8 and 9).

3.5.1. Hydrocarbon Mobilization within the Radial Fracture System of the ca. 17 Ma Columbia River LIP

The Columbia River LIP represents the youngest major continental flood-basalt province and is widely interpreted as recording impingement of the Yellowstone starting plume head beneath western North America [159, 160]. Magmatism began with the Steens basalts at ca. 17 Ma and progressed rapidly through the Imnaha, Grande Ronde, Wanapum and Saddle Mountains flood-basalt units, with high-precision U–Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology showing that much of the main eruptive phase was emplaced within <1 Myr [161]. The associated broadly radiating dyke swarm (e.g., [37]) and subsequent northeast-younging Snake River Plain–Yellowstone volcanic track are consistent with an initial plume-head phase followed by motion of North America over the plume tail to the present-day position of the Yellowstone hotspot.

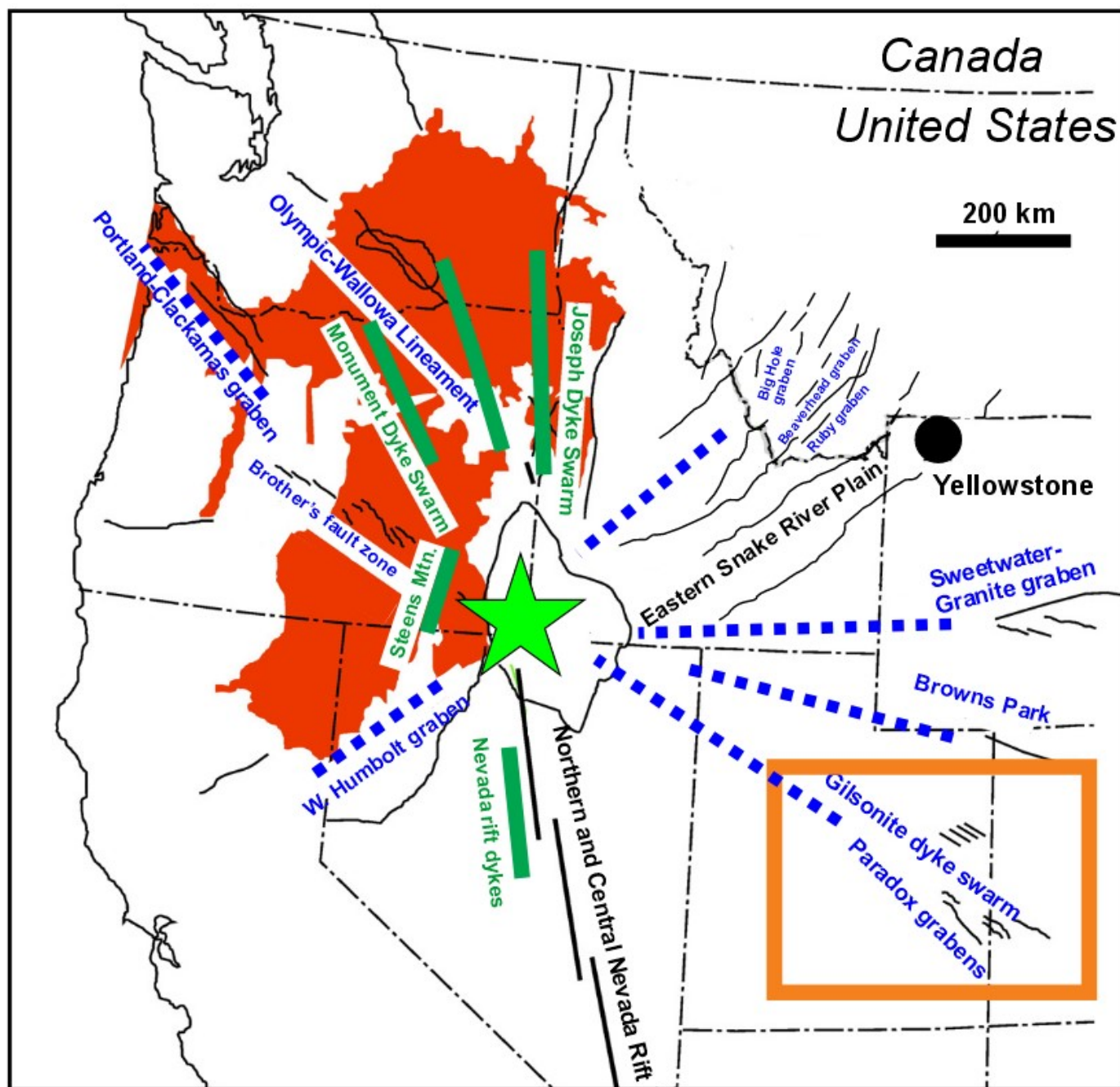


Figure 8. Columbia River LIP and associated radial dyke and graben systems. The ca. 17 Ma Columbia River LIP is linked to initiation of the Yellowstone hotspot (inferred starting-plume centre) (green star) located near the convergence of the Chief Joseph, Monument and Nevada Rift radial subswarms (green generalized lines); the Steens Mountain swarm does not form part of this radial pattern [37, 159]. East of the inferred plume centre, radial graben and fracture systems include the Paradox grabens, Brown Park, Sweetwater Granite graben, Big Hole graben, Beaverhead graben and Ruby graben; additional grabens occur to the south and northwest (modified after [162]). The orange box marks the area of the Paradox grabens and associated Gilsonite dykes. Red areas show volcanic rocks (flood basalts); the black dot marks the present-day position of the Yellowstone hotspot.

The radial dyke swarm is best developed north and west of the inferred plume centre (Figure 8). To the east, Sears et al. [162] identified Middle Miocene graben and fracture systems in southwest Montana that continue the radial structural pattern but lack associated dyke magmatism. They interpreted these grabens as part of a radial

extensional system developed on a broad dome centred near the region of initial arrival of the “Yellowstone hotspot outbreak area” [162] which corresponds to the region of initial arrival of the Yellowstone starting plume head (Figure 8). We suggest that the lack of associated dyke magma reflects a sector in which partial melts were unavailable or

unable to ascend. Sears et al. [162] also labelled a “Gibsonite Dike Swarm” near the Paradox grabens; this appears to refer to the Gilsonite dyke swarm farther south in the eastern Uinta Basin.

The Gilsonite dykes are long, subparallel hydrocarbon-filled fractures composed of brittle solid bitumen derived from organic-rich source beds of the upper Green River Formation [163]. Their forceful emplacement has been interpreted to reflect hydraulic-fracture propagation from overpressured underlying bitumen-rich strata. We suggest that plume-related uplift and extension may have helped generate or reactivate radial fractures that were subsequently exploited by overpressured hydrocarbons migrating into these high-permeability pathways.

The presence of radial grabens and fractures lacking magma in the eastern part of the Columbia River system indicates that plume-generated fracturing need not be accompanied by contemporaneous melt emplacement. Partial melts may have been unavailable or unable to ascend beneath this sector, potentially because thicker lithosphere inhibited decompression melting or melt transfer. Plume-related radial and circumferential fractures should therefore be regarded not only as potential magma conduits, but also as potential pathways for non-magmatic fluids.

3.5.2. Plumes Associated with Progressive Gondwana Breakup

A wide range of vein and breccia-hosted systems can potentially be influenced by plume-generated radial or circumferential fractures (Table 2) and examples can be sought through evaluating the orientation and ages of such systems in the vicinity of mantle plumes (i.e. within 1000–1250 km radius of the inferred plume centre). Here we consider the possibility of such systems associated with the mantle plumes associated with the Gondwana breakup.

Figure 9 illustrates the potential spatial relationship between Gondwana breakup-related mantle plumes [164, 165] and hydrothermal mineralization. Idealized ~1000-km-radius plume-head footprints (red superimposed lines) are used as a first-order approximation of the regions affected by topographic changes from an underlying mantle plume. Within these broad footprints, plume-related fracturing (radial and/or circumferential) could have enhanced permeability, focused hydrothermal fluid flow and influenced mineralization. If such links occurred during Gondwana breakup, the associated hydrothermal systems would be expected to broadly overlap in age and be spatially associated with the relevant plume head events at ca. 201, 183, 132, 130, 90, 67 and 30 Ma.

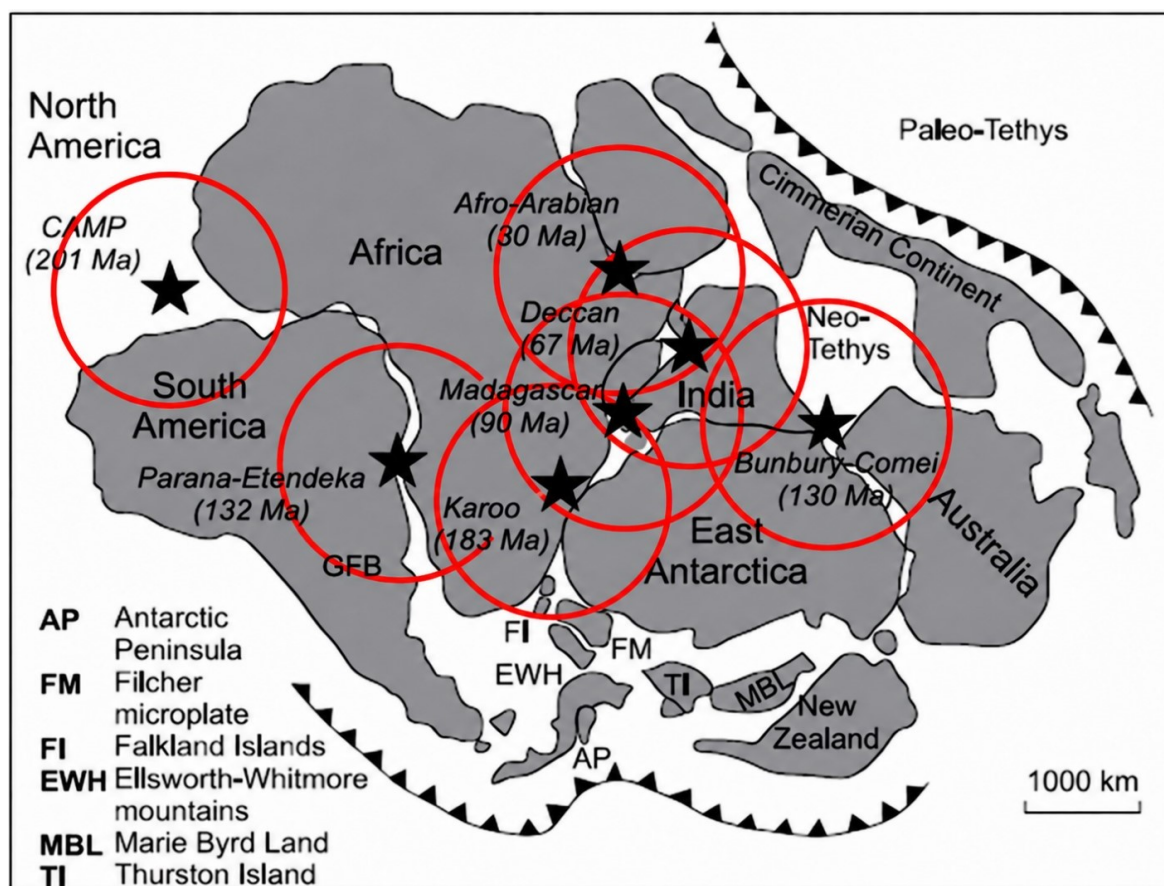


Figure 9. Gondwana reconstruction modified after [164, 165] with superimposed plume/LIP centres (black stars) (cf. Figure 11.1 in [5]) surrounded by 1000 km radius red circles indicating the inferred plume head size. The LIPs/plumes include: 201 Ma CAMP, 183 Ma Karoo, 132 Ma Parana-Etendeka, 130 Ma Bunbury-Comei, 90 Ma Madagascar, 67 Ma Deccan and 30 Ma Afro-Arabian.

Table 2. Vein systems that can potentially be linked to plume dynamics.

Vein-System Type	Possible Plume-Fracture Relationship
Epithermal Au–Ag veins	Plume- or LIP-related heating and magmatism may enhance shallow hydrothermal circulation, while radial or circumferential fractures may localize boiling, brecciation and vein corridors. Sections 3.4 and 3.5
Polymetallic Ag–Pb–Zn–Cu veins	Basinal, magmatic or metamorphic fluids may be focused into plume-related fracture corridors, dyke margins, fault intersections or annular structures. Sections 3.3–3.5
Five-element Ag–Co–Ni–Bi–As veins	Brittle faulting, overpressure, reducing hydrocarbons or methane, arsenide precipitation and repeated hydraulic fracturing are central to these systems and could be enhanced by plume-generated fracture permeability. Sections 3.4 and 3.5
U–Ag–Co and related arsenide veins	Oxidized basinal brines may interact with reduced basement rocks along reactivated plume-related faults and fracture corridors. Sections 3.3–3.5.
Fluorite–barite ± Pb–Zn veins	Basinal brines may be focused through extensional fractures, with plume-related heating and rifting enhancing circulation. Sections 3.3–3.5.
IOCG and breccia-hosted Cu–Au–U systems	Plume-related heat, stress changes and reactivation of basement faults may promote flow of oxidized saline fluids and hydrothermal brecciation. Sections 3.3–3.5.
Orogenic Au–quartz and quartz–carbonate veins	Reactivation of crustal-scale faults, fault-valve behaviour and repeated fracture opening and sealing; plume-related stress changes could influence the timing and orientation of fluid pulses. Link to plume push. Section 4.
Porphyry and skarn vein or stockwork systems	These systems are more directly magmatic-hydrothermal, but plume-related magmatism or stress reorganization may influence magma storage, fracture-controlled release and fluid focusing and also distal plume push. Sections 3.4 and 4.3.

3.5.3. Exploration Targeting Implications

As outlined above, plume-related doming and stress reorganization can generate radial and circumferential fracture systems that potentially focus hydrothermal fluids and may influence a wide range of fracture-hosted mineral systems (Table 2). Future exploration-oriented testing should compare the timing and distribution of such systems with reconstructed plume centres and plume-related fracture sets, including both magma-filled and unfilled structures, as potential pathways for hydrothermal fluids and as possible controls on fracture-hosted mineralization, including epithermal systems (e.g., [166–168]), located within about 1000 km radius of plume centres.

4. Distal “Sudden” Regional Stress Changes Caused by Mantle-Plume-Head Arrival and Ore Deposit Links

4.1. Plume-Head Push

The preceding sections focused on mineralization processes occurring above or near mantle-plume heads.

Here, we consider the more distal consequences of plume-head arrival, particularly the transmission of plume-related stresses through the lithosphere to remote plate boundaries and their possible influence on orogenic gold and porphyry systems. More broadly, mantle plumes represent a distinct mode of mantle convection capable of producing tectonic as well as magmatic effects on the overlying lithosphere [169].

Ernst and Jowitt [12, 13] proposed an indirect relationship between LIP–plume-related continental rifting or attempted breakup and broadly coeval episodes of compression or transpression at distant plate boundaries (Section 1.3). The concept follows from the plate-tectonic requirement that extension within one part of a global plate circuit must be accommodated by convergence, transpression or changes in plate motion elsewhere. Building on the BSc research of Kuzmich [170, 171], we further developed this hypothesis by considering how the arrival and lateral spreading of a plume head may rapidly reorganize intraplate stresses and affect plate-boundary kinematics. We then explore whether such stress perturbations could

trigger or enhance mineralization in distal orogenic / subduction settings.

The proposed plume push–mineralization linkage rests on three considerations. First, geologically rapid changes in regional stress, can reorganize fault slip, fracture permeability and fluid flow and thereby trigger short-lived mineralizing episodes (e.g., [73, 172, 173]). Second, plume-head arrival provides a mechanism capable of producing such abrupt stress changes and plume-push, and transmitting them through a plate circuit to boundaries thousands of kilometres from the plume centre (Sections 4.1.3–4.1.5). Third, plume-related forcing may produce geologically abrupt, plate-scale stress perturbations that contrast with the more gradual changes expected from background mantle convection or long-term plate-boundary evolution. Short-lived changes in regional stress or plate motion may therefore provide a useful indicator of plume push, particularly where they are broadly synchronous with plume-head arrival and cannot be readily explained by local tectonic processes alone. Rapid stress changes can also accompany terrane accretion, microcontinental collision, slab breakoff, or major changes in subduction geometry (e.g., [76, 174]). We further hypothesize that, in some cases, such apparently plate-boundary-driven reorganizations may themselves form part of the plate-scale response to externally imposed plume forcing.

Thus, unusually abrupt, regionally coherent changes in stress or plate motion provide particularly useful targets for testing the plume-push hypothesis, particularly where they coincide with independently dated plume-head emplacement, can be linked through plate reconstructions to a plausible pathway of plume-related stress transmission, and are accompanied by short-lived mineralization whose timing and structural setting are consistent with the same stress reorganization.

4.1.1. Evidence for the Magnitude and Reach of Plume-Head Push

The contribution of mantle plumes to plate dynamics may be much greater than is commonly recognized. Arrival of a large plume head can perturb or locally overwhelm pre-existing intraplate stresses and may influence the motion of an entire plate, with effects transmitted to boundaries thousands of kilometres from the plume centre [175, 176] and references therein.

Van Hinsbergen et al. [177], for example, concluded that arrival of the Morondava (Madagascar) plume beneath Madagascar–India generated a strong plume-related torque that rotated the Indian Plate and reorganized plate boundaries over very large distances, contributing to formation of a >12,000-km-long India–Africa plate boundary extending from the southern Indian Ocean to the eastern Mediterranean. The effects therefore propagated thousands of kilometres from the inferred plume centre. More generally, van Hinsbergen et al. [177, 178] showed that plume-head arrival can substantially mod-

ify both the direction and rate of plate motion and can influence plate-boundary kinematics far from the plume source. Stotz et al. [179] documented a marked slowdown of the North American Plate at ca. 17 Ma, broadly coincident with initiation of the Columbia River LIP and arrival of the Yellowstone plume. At the supercontinent scale, Zhang et al. [180] found that stresses associated with subcontinental mantle upwelling can exceed those generated by subduction retreat, implying that mantle upwelling may in some settings be a first-order driver of plate reorganization. Together, these studies indicate that plume-head arrival can generate stresses of sufficient magnitude to alter plate motion, rotate crustal blocks, initiate or reactivate rifts, and modify strain partitioning along distal plate boundaries.

4.1.2. Geodynamic Effects, Duration and Magnitude of Displacement

The principal geodynamic effects of plume-head arrival and spreading are illustrated schematically in Figure 10. These include outward plume-head push, continental rifting, translation or rotation of rigid lithospheric blocks, and changes in plate motion resulting from plume-related forcing. Although these processes may operate together, they represent distinct ways in which plume emplacement can modify the magnitude and direction of plate motion. Here we focus on two aspects: (1) the potential distance over which plume push stresses can be transmitted in lithospheric plates and (2) estimates of the magnitude and duration of plume push effect on causing displacement of lithospheric plates.

4.1.3. Spatial Scale Over Which Intraplate Stresses can be Transmitted

The spatial scale over which intraplate stresses can be transmitted is illustrated by the present-day World Stress Map project (<https://www.world-stress-map.org/>), which documents broad, coherent first-order stress provinces within plate interiors. Zoback [181] noted that broad-scale first-order regional stress patterns often extend over distances of approximately 20–200 times the ~20–25 km thickness of the upper brittle lithosphere, corresponding to spatial scales of roughly 400–5000 km. Such broad midplate stress fields were interpreted principally as responses to plate-scale forces, including ridge push and continental collision. More recent World Stress Map compilations confirm the persistence of long-wavelength lithospheric stress patterns, while also showing that they can be locally or regionally modified by lithospheric heterogeneity, topography, and other second-order effects [182–185]. Thus, transmission of tectonic stress over distances of several thousand kilometres is consistent with the behaviour of mechanically coupled lithospheric plates, although the local magnitude and orientation of stress may vary substantially along the transmission path.

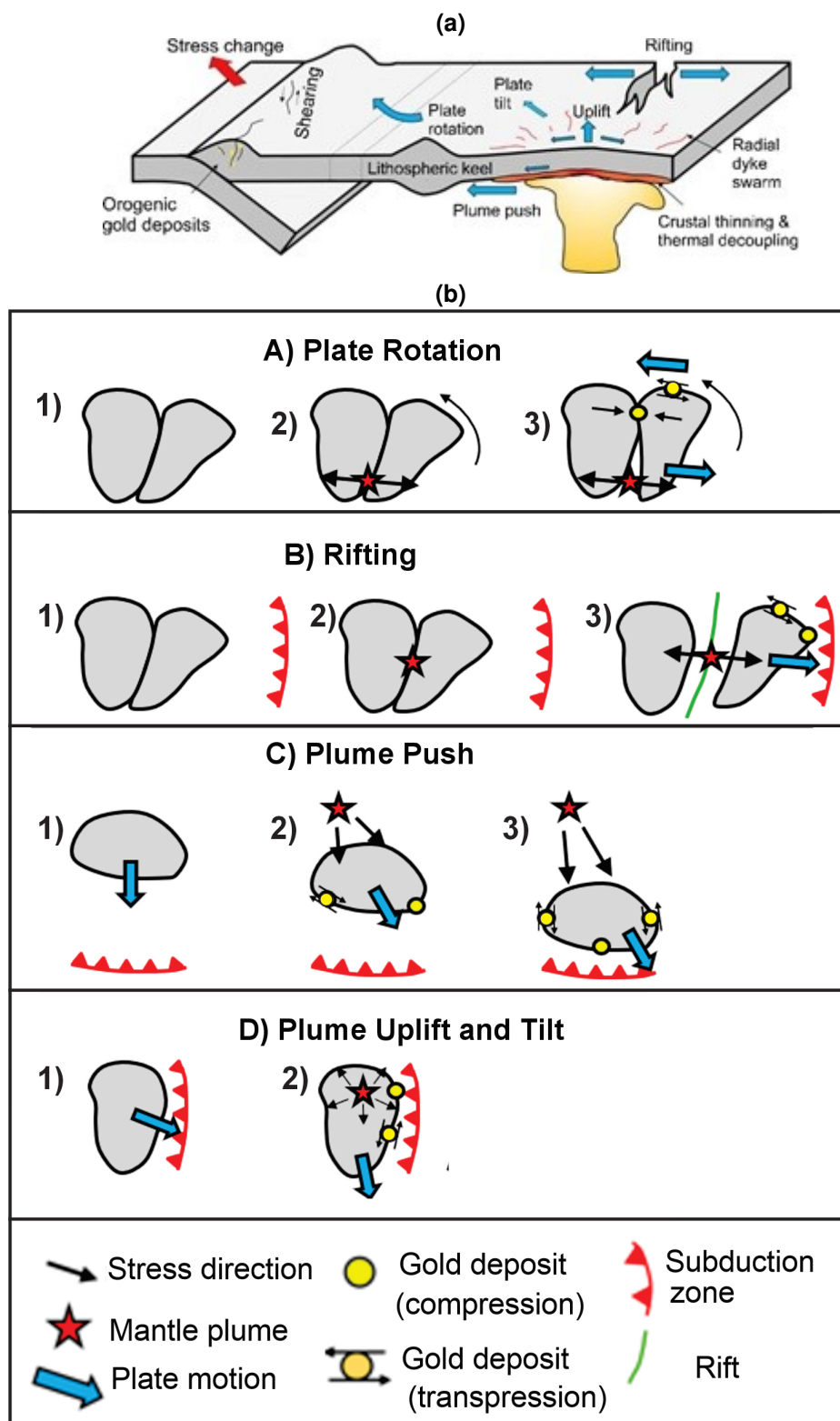


Figure 10. Role of mantle plumes in causing a change in distal plate motions, leading to distal stress changes facilitating emplacement of distal orogenic mineralization. (a) Simplified cross-sectional diagram of a mantle plume causing an abrupt increase in convergence and change in plate motion direction leading to transpression at the location of an orogenic gold deposit. The mechanisms for driving the plate motion are uplift, plume push, plate rotation and rifting. (b) Plan view of the different types of effects due to mantle plume arrival: plate rotation, plate rifting, plume push and plume uplift causing plate tilting. All these mechanisms can cause a geological rapid change in plate motion. Revised after [171].

In the context of mantle plumes, several mechanisms may generate such plate-scale stress perturbations. Plume-head arrival can dynamically uplift the overlying lithosphere, increasing its gravitational potential energy and generating outward “downhill” forces away from the uplifted region [178]. In addition, arrival and spreading of a plume head can generate a lateral pressure gradient within the weak asthenospheric layer at the base of the lithosphere, driving radially outward, pressure-driven (Poiseuille) asthenospheric flow that exerts basal shear stresses on the overlying lithosphere and can modify plate motion [179, 186]. Such outward asthenospheric flow should be particularly important where it encounters lateral variations in lithospheric thickness, including deep cratonic roots that can obstruct or deflect the flow and thereby modify the magnitude and orientation of stresses transmitted into the lithosphere.

Having established that lithospheric stresses can be transmitted over distances of several thousand kilometres, we next consider the magnitude and duration of plume push itself—specifically, the extent to which plume-head spreading can produce measurable displacement of adjacent lithospheric plates or crustal blocks. We first consider constraints from plume modelling and then examine three case studies—the Gawler, Réunion, and Keweenawan plume systems—as first-order estimates of the scale and duration of plume-related displacement.

4.1.4. Magnitude of Plate Movement Potentially Caused by Plume Push

The dimensions of model plume heads provide a first-order estimate of the maximum spatial scale over which plume push could operate. A deep-mantle plume head is commonly envisaged as approximately 500 km in radius before final impingement, subsequently flattening and spreading beneath the lithosphere to a radius of roughly 1000–1250 km [34]. Because horizontal forcing can begin as the ascending plume head encounters and spreads beneath the lithosphere, the maximum radial scale available for plume-induced plate displacement could approach the final plume-head radius of approximately 1000–1250 km, in the limiting case of highly efficient plume–lithosphere coupling. This scale is notably similar to the approximately 1000–1200 km of excess northward displacement of India estimated from the Réunion-related acceleration documented by [187, 188] (Section 4.1.5; Table 3).

However, the plume-head radius defines only the region of direct plume–lithosphere interaction, not the spatial extent of the resulting plate motion. If plume push against thick lithosphere produces net plate translation, that displacement is transmitted through the mechanically coupled plate and may therefore affect regions thousands of kilometres beyond the plume-head footprint, while plume-driven asthenospheric flow (Section 4.1.3) can provide an additional, more distributed source of basal traction over similarly large distances.

The nature and persistence of this plume–lithosphere coupling depend on how plume material interacts with the moving plate. Numerical models show that once plume material reaches the base of a moving plate, shallow plume material can become strongly entrained and advected in the direction of plate motion, while the deeper plume conduit may be deflected or become increasingly offset from the shallow plume head [189, 190]. Hot plume material can also remain ponded and spread laterally at sublithospheric depths for substantial periods [191]. More generally, plume heads have been proposed to become accreted or subcreted to the lithosphere and subsequently travel with their host plate, persisting as “fossil plume heads” that may continue to influence later magmatism [5, 192, 193]. Thus, beneath a rapidly moving plate, shallow plume material may be transported with the plate while continuing to spread laterally and becoming progressively offset from its deeper plume tail.

Not all plume-head expansion, however, is converted directly into translation of the overlying plate. A substantial component of plume material may spread laterally within the asthenosphere, while still exerting some basal traction on the lithosphere [179, 186, 194]. The resulting plate response therefore reflects both direct mechanical interaction with the spreading plume head and more distributed forcing associated with plume-driven asthenospheric flow.

The efficiency and direction of plume-induced plate motion should consequently depend strongly on lithospheric geometry and mechanical coupling. Where a spreading plume head or plume-driven asthenospheric flow encounters thick lithosphere or a deep lithospheric root, lateral flow may be impeded or deflected, potentially increasing the transfer of plume-related stresses into the lithosphere in that direction. Beneath thinner lithosphere, a larger proportion of plume material may escape laterally through the asthenosphere, although this flow can still exert some basal traction on the overlying plate. Plume push therefore need not be equally effective in all directions: its strongest focused component may develop where outward plume flow encounters adjacent thick lithosphere, while broader asthenospheric flow can transmit more distributed stresses beyond the immediate plume-head region.

4.1.5. Rate and Magnitude of Plume Push from Case Histories

Available geological and kinematic examples permit preliminary estimates of the duration and magnitude of plume-head spreading and associated plume push. Three complementary examples provide different constraints on this process.

Réunion Plume Head (Starting Plume): A particularly well-documented example of plume-generated horizontal forcing is provided by the Réunion starting plume. The Indian craton forms the continental core of the Indian Plate, and the arrival of the Réunion plume beneath India at ca. 67–65 Ma, associated with the Deccan LIP, has been linked to a marked acceleration of northward Indian plate

motion [187]. Cande and Stegman [187] showed that the unusually rapid motion of India between ca. 67 and 52 Ma was coupled with a contemporaneous slowing of Africa, and argued that both reflected the force exerted by the Réunion plume head.

Cande and Patriat [188] subsequently refined this kinematic history. India–Africa relative motion accelerated from approximately 60 mm yr^{-1} at ca. 71 Ma to approximately 180 mm yr^{-1} by ca. 66 Ma, remained exceptionally rapid for about 3 Myr, and then decreased sharply between ca. 63 and 62 Ma before declining more gradually. Importantly, their mantle-reference-frame reconstruction shows a closely comparable acceleration in the absolute northward motion of India, with the maximum absolute velocity reaching approximately 220 mm yr^{-1} during Chron 29R, coincident with the most voluminous phase of Deccan volcanism. Cande and Patriat [188] further noted that India's absolute velocity during this interval was approximately $20\text{--}30 \text{ mm yr}^{-1}$ greater than its velocity relative to Africa. Their reconstruction is technically based largely on Capricorn Plate rotations derived from Indian Ocean spreading data; for the Late Cretaceous–early Cenozoic interval considered here, Capricorn motion provides a close approximation to Indian Plate motion.

The total northward displacement of India during this interval cannot, however, be attributed solely to plume push because India was already moving rapidly before the Réunion plume-head event. A more appropriate first-order estimate is obtained by comparing the enhanced absolute velocity of India with its pre-acceleration absolute velocity, which was approximately $80\text{--}90 \text{ mm yr}^{-1}$. Integrating the excess velocity represented by the Capricorn absolute-motion (Cap–Abs) curve of Cande and Patriat [188, their Figure 15] provides a first-order estimate of approximately 1000 km of additional northward displacement,

with a plausible first-order range of roughly 1000–1200 km depending on the background velocity adopted (Table 3). This estimate is preferable to one based directly on the Capricorn–Antarctica (Cap–Ant) spreading-rate curve [188, their Figure 14], because Cap–Abs represents the anomalous translation of India itself, whereas Cap–Ant records relative spreading across the Southeast Indian Ridge. The resulting first-order displacement estimate is notably similar to the approximately 1000–1250 km maximum radial scale of plume-head spreading considered in Section 4.1.4.

For an intraplate plume such as the Réunion starting plume, approximately radial outward flow associated with plume-head spreading would not necessarily produce substantial net translation of the overlying plate if plume–lithosphere coupling were approximately symmetric, because tractions acting in different radial directions would tend to partly cancel. However, plume-related asthenospheric flow can be strongly influenced by variations in lithospheric thickness and by nearby spreading ridges [179, 194–196].

A substantial net northward plume-push force on India therefore implies an asymmetric plume–lithosphere interaction. Numerical modelling by Bredow et al. [195] shows that, following arrival of the Réunion plume head beneath the thick continental lithosphere of India, plume material preferentially migrated toward adjacent thinner oceanic lithosphere before melting, illustrating how lithospheric-thickness contrasts can strongly redirect plume-head flow. We suggest that such asymmetric confinement and escape of plume material could have preferentially transferred plume-related stresses into the thick Indian lithosphere, generating a net northward component of force consistent with the anomalous acceleration of India documented by [187, 188].

Table 3. First-order estimate of the Réunion plume-push contribution to anomalous northward displacement of India, based on the absolute India/Capricorn plate-motion history of Figure 15 in [188].

Interval	Approximate Absolute India/Capricorn Velocity	Excess above 90 mm/yr	Duration	Extra Distance
71–67 Ma	Rises $\sim 90 \rightarrow 180 \text{ mm/yr}$	Average excess $\sim 45 \text{ mm/yr}$	4 Myr	$\sim 180 \text{ km}$
67–65 Ma	Rises $\sim 180 \rightarrow 220 \text{ mm/yr}$	Average excess $\sim 110 \text{ mm/yr}$	2 Myr	$\sim 220 \text{ km}$
65–63 Ma	Falls $\sim 220 \rightarrow 180 \text{ mm/yr}$	Average excess $\sim 110 \text{ mm/yr}$	2 Myr	$\sim 220 \text{ km}$
63–62 Ma	Falls $\sim 180 \rightarrow 140 \text{ mm/yr}$	Average excess $\sim 70 \text{ mm/yr}$	1 Myr	$\sim 70 \text{ km}$
62–52 Ma	Falls $\sim 140 \rightarrow 100 \text{ mm/yr}$	Average excess $\sim 30 \text{ mm/yr}$	10 Myr	$\sim 300 \text{ km}$
Total 71–52 Ma				$\sim 990 \text{ km}$

Note. Values are first-order estimates read from the absolute India/Capricorn velocity history in [188]. Using an 80 mm yr^{-1} rather than a 90 mm yr^{-1} pre-acceleration baseline increases the integrated excess displacement to approximately 1180 km; hence the anomalous displacement associated with the Réunion-related acceleration is best expressed as approximately 1000–1200 km rather than as a precise value.

1590 Ma Gawler Plume: The ca. 1590 Ma Gawler LIP and its relationship to Olympic Dam were discussed previously in Section 3.4.1 and illustrated in Figure 7, where the event was considered as an example of plume-related thermal and magmatic input to a proximal mineral system. Here, we return to the same event because its regional geochronological pattern provides a particularly useful first-order constraint on the rate of plume-head spreading. High-precision geochronology indicates that intrusive and volcanic magmatism becomes broadly younger radially outward from the Olympic Dam region. The oldest magmatism occurs at ca. 1595–1592 Ma near the inferred plume centre, whereas younger magmatism reaches ca. 1587–1586 Ma toward the margins of the volcanic province. Jagodzinski et al. [156] interpreted this spatial–temporal progression as consistent with arrival of an ascending plume head followed by lateral flattening and spreading beneath the lithosphere. From the regional distribution of the dated magmatism, the outward propagation is of order 500–700 km over approximately 8–9 Myr, corresponding to a first-order radial propagation rate of approximately 55–90 mm yr⁻¹.

As noted in Section 3.4.1, the Gawler LIP is silicic-dominated, comprising the Gawler Range Volcanics and associated Hiltaba Suite intrusions, with mafic rocks representing a volumetrically subordinate component. Most of the dated units defining the outward-younging pattern are therefore silicic volcanic and granitic rocks. These felsic rocks can be interpreted to record high-temperature crustal melting, including partial melting of refractory lower crust, accompanied by substantial and repeated input of mantle-derived magma and hybridization between mantle- and crust-derived melts [197–199]. This distinction is important because the spatial progression of silicic magmatism may provide a particularly useful record of the outward migration of plume-head thermal influence. Unlike mafic magmatism, which may in part be supplied laterally through centrally fed radial dyke systems, felsic magmas generated by lower-crustal melting are more likely to reflect heating directly beneath the region where they are emplaced. Consequently, the outward decrease in silicic magmatic ages may approximately track the passage of the thermal pulse associated with the outer margin of the spreading plume head and progressive heating and partial melting of the overlying lower crust.

Keweenawan Plume: The Keweenawan–Abitibi plume system was discussed previously in Section 3.1.2 and illustrated in Figure 2b in the context of plume-generated stress changes and emplacement of mineralized chonoliths. Here, we return to the same system because the later Keweenawan magmatic pulse provides another useful constraint on the magnitude of plume-related crustal displacement. Ernst et al. [96] interpreted the ca. 1150–1140 Ma Abitibi–Corson event and the subsequent ca. 1110–1092 Ma Keweenawan magmatism as successive pulses of the same long-lived mantle plume system, utilizing essentially the same plume centre south of Lake Superior. The earlier ca. 1150–1140 Ma event produced

a large radial dyke system interpreted to record an initial plume head approximately 900 km in radius, whereas the later Keweenawan episode included two principal magmatic pulses at ca. 1110–1103 and 1099–1092 Ma associated with a smaller, 250 × 330 km elliptical plume-head (Figure 2b; Section 3.1.2).

During the younger 1099–1092 Ma pulse, approximately 80 km of opening of the Midcontinent Rift has been interpreted as resulting from lateral plume-head spreading and plume push against the thick lithospheric root of the Superior craton to the north [96]. In this interpretation, the plume-push interval lasted approximately 7 Myr and produced about 80 km of localized crustal extension, associated with the 250 × 330 km elliptical plume head pulse.

Summary: Taken together, these examples suggest that significant plume-head spreading and associated horizontal forcing may persist for approximately 7–15 Myr, and potentially longer, while producing very different amounts and styles of displacement depending on lithospheric geometry and plate-boundary conditions. The quantities constrained by the three examples are not directly equivalent: the Gawler record potentially tracks the outward propagation of plume-related thermal and magmatic influence; the Réunion example provides a first-order estimate of anomalous northward translation of the Indian Plate during plume-related acceleration; and the Keweenawan example records localized crustal extension associated with plume-head spreading. Nevertheless, all three indicate that plume-head evolution can exert sustained horizontal forcing over multi-million-year timescales, with effects ranging from tens of kilometres of localized extension to approximately 1000 km of cumulative plate displacement.

These estimates of the magnitude and duration of plume-related horizontal forcing provide a basis for considering, in the following sections, whether plume-induced changes in plate-scale stress and motion can contribute to the timing and localization of distal orogenic gold and porphyry mineral systems.

4.2. Link between Mantle Plumes and Orogenic Gold Deposits

4.2.1. Orogenic Gold Deposits

Orogenic gold deposits are structurally controlled lode systems formed mainly in convergent orogenic belts during compressional to transpressional deformation [200–203]. They are typically localized along crustal-scale faults and shear zones, where episodic deformation, seismic failure and fluid-pressure fluctuations generate transient permeability and focus gold-bearing fluids. Mineralization commonly forms at mid-crustal depths under greenschist- to amphibolite-facies conditions, with gold deposition promoted by pressure changes, fluid–rock interaction, cooling and sulfidation.

The mineralizing fluids associated with orogenic Au systems are generally low-salinity, CO₂-rich aqueous solutions (e.g., [202]). Although traditionally attributed mainly to metamorphic devolatilization, the sources of fluids and metals remain debated, with more recent models allowing

contributions from metasomatized lithospheric mantle and transfer of gold through mantle-derived fluids or alkaline mafic to ultramafic magmas [204–206].

Exploration models have therefore focused largely on favourable fault architecture and the structural, geophysical, geochemical and geochronological evolution of orogenic belts. Less attention has been given to the external geodynamic processes capable of rapidly changing regional stress and fault kinematics, despite their potential importance in triggering seismic pumping, fault-valve behaviour and mineralizing fluid flow [47]. Here we propose that mantle-plume arrival and lateral spreading can reorganize plate-scale stresses and thereby trigger or enhance orogenic gold mineralization at distal plate boundaries [170, 171].

4.2.2. Ontong Java Nui and the ca. 120 Ma Jiaodong Gold Event

Goldfarb et al. [47], and subsequently [203], proposed that emplacement of the Ontong Java Plateau triggered major reorganization of Pacific-region plate motions, changing the stress regime along eastern Asia and contributing to formation of the giant ca. 120 Ma Jiaodong gold province of eastern North China (Figure 11). More broadly, Sun et al. [207] linked large-scale Early Cretaceous gold mineralization in eastern China to a major change in Pacific Plate motion. This general framework has recently been further developed by [208], who document a short-lived ca. 122–120 Ma stress-field inversion closely associated with Jiaodong mineralization and discuss it in the context of contemporaneous Paleo-Pacific plate reorganization.

Ontong Java Nui (OJN) represents the originally contiguous Ontong Java, Manihiki, and Hikurangi plateaus, together with the Nauru Basin, with major initial volcanism at ca. 123–120 Ma [209–212]. Younger high-precision $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ca. 116–108 Ma from OJP and Manihiki lavas indicate substantial later or protracted volcanism [213], although stratigraphic and geochemical evidence indicates that OJN volcanism was already underway by ca. 120 Ma [212]. In the Goldfarb–Yang model, OJP/OJN emplacement reorganized the Pacific–Phoenix–Izanagi plate system, ultimately modifying the stress regime along eastern Asia and reactivating the Tan–Lu fault system, thereby facilitating ascent of auriferous fluids.

Alternative reconstructions of Pacific–Izanagi plate evolution are provided by [214, 215]. Torsvik et al. [214] do not require the pronounced change in Izanagi Plate motion envisaged in the Goldfarb–Yang model, but nevertheless show major Pacific plate-system reorganization at approximately the same time, expressed by breakup of the Pacific–Phoenix system into four plates—the Pacific, Manihiki, Hikurangi, and Catequil plates—during OJN emplacement. Wu et al. [215], using constraints from western Pacific mantle structure, likewise reconstruct substantial Early Cretaceous reorganization of the Pacific–Izanagi system, although with different details of plate geometry and motion. Thus, although the detailed kinematics differ

among these models, they independently indicate major reorganization of the Pacific plate system broadly contemporaneous with OJN emplacement.

Structural studies of the Jiaodong gold district also indicate that the principal gold event coincided with a marked change in regional stress and fault kinematics. A broader transition from NW–SE compression toward extension began at ca. 135 Ma [216], whereas Ref. [208] identify a sharper stress-field inversion at ca. 122–120 Ma, closely associated with mineralization, from dextral extensional shear during the pre-ore stage to sinistral compressional shear during the syn-ore stage. Earlier geochronological work also linked Jiaodong gold mineralization and magmatism to continental-margin transtension [217], while Ref. [218] relate the ca. 120 Ma mineralization to changes in Paleo-Pacific subduction dynamics and a transition from compression toward transpression or transtension, within the broader context of evolving Paleo-Pacific/Izanagi subduction beneath eastern Asia [219]. Recent geochronological studies provide additional support for concentration of gold mineralization at approximately this time, including the Jiudian deposit in the Jiaodong Peninsula [220] and the Baiyun deposit of the adjacent Liaodong Peninsula, interpreted as part of the broader Early Cretaceous Jiaodong-type gold mineralizing system [221].

These broadly contemporaneous changes support the possibility that OJN emplacement contributed a far-field plume-push perturbation to an eastern Asian margin already undergoing subduction- and rollback-related reorganization, thereby helping to trigger or sharpen the ca. 122–120 Ma stress change recorded in Jiaodong. Such a stress perturbation could have promoted reactivation of Tan–Lu-related structures, creating or enhancing transient permeability and focusing auriferous fluids. The temporal association between major OJN emplacement at ca. 123–120 Ma, Pacific plate-system reorganization, the Jiaodong stress inversion at ca. 122–120 Ma, and gold mineralization at ca. 121–119 Ma therefore provides an intriguing framework consistent with an OJN plume-related contribution to plate reorganization and metallogenesis, although the precise dynamic linkage remains to be established.

4.2.3. Ontong Java Plateau/subduction Effects at ca. 15 Ma

The metallogenic influence of the Ontong Java Plateau was not restricted to possible plume-push effects associated with initial emplacement of the OJP plume head at ca. 123–120 Ma. A distinctly different consequence occurred much later, during the Oligocene–Miocene, when the thickened OJP crust encountered the Melanesian/Kilinaillau subduction system. In the tectonic model adopted by [222, 223], southwestward subduction of the Pacific Plate ceased at about 15 Ma as a consequence of collision of the OJP with the Kilinaillau trench, producing plate rotation and regional stress reorganization. More recent regional reconstructions place initial OJP collision with the Melanesian arc somewhat earlier, during the late Oligocene to early Miocene, but likewise associate

the collision with disruption of the pre-existing subduction system and subsequent major plate-boundary reorganization [224].

Continued convergence was subsequently accommodated through subduction reversal and development of the north-dipping New Britain subduction system, together with progressive extension, transtension and microplate reorganization in the Bismarck–New Ireland region. These processes ultimately included back-arc spreading in the Manus Basin and extension within the

New Ireland Basin, promoting partial melting of previously subduction-modified mantle beneath the Tabar–Lihir–Tanga–Feni (TLTF) chain [222, 223, 225, 226]. The resulting alkaline magmatic systems include the giant Ladolam porphyry–epithermal Au deposit at Lihir and Cu–Au mineralization at Feni [223, 226, 227]. The structural framework of the TLTF chain has been interpreted in terms of regional extension and transtension related to the transition from subduction to collision and subsequent plate reorganization.

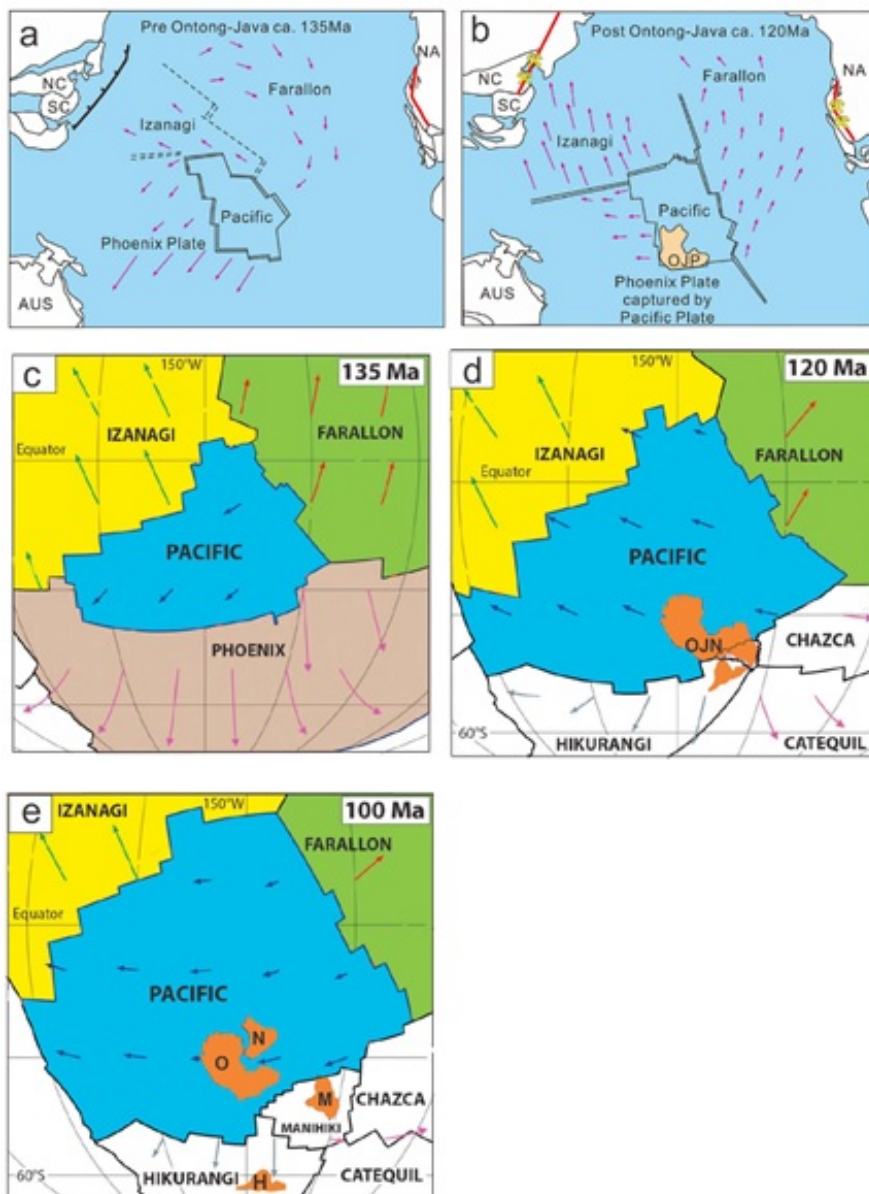


Figure 11. Alternative reconstructions relevant to possible links between Ontong Java Nui (OJN) emplacement, Pacific plate-system reorganization, and Jiaodong gold mineralization. **(a, b)** After [208], modified from [203]. This model links OJP emplacement with reorganization of the Pacific–Phoenix–Izanagi plate system, including capture of the Phoenix Plate by the Pacific Plate and broadly coeval ca. 125 Ma gold mineralization around the northern Pacific margin. **(c–e)** Alternative reconstructions modified after [214]. These do not show a comparable change in Izanagi Plate motion but show major reorganization and breakup of the Pacific–Phoenix plate system associated with OJN emplacement. OJN represents the reconstructed Ontong Java, Manihiki, and Hikurangi plateaus and Nauru Basin [209–211]. Orange polygons show fragmented Ontong Java (O), Manihiki (M) and Hikurangi (H) plateaus, and Nauru Basin (N).

Thus, the OJP illustrates two temporally and dynamically distinct ways in which a plume-generated LIP may influence metallogenesis: initially through plume-head emplacement and associated plate-scale stress effects (Section 4.2.2), and much later through collision and attempted subduction of the thickened oceanic plateau and consequent reorganization of the regional plate-boundary system (this section). Similar metallogenic effects have been discussed elsewhere for subduction of oceanic plateaus, aseismic ridges and other buoyant oceanic features, which can modify slab geometry, crustal thickness, magmatism and porphyry–epithermal mineralization (e.g., [12, 228–230]). We therefore introduce the OJP case here primarily as an example of this broader class of plateau (oceanic LIP)-subduction effects on mineralization, rather than attempting a comprehensive review.

4.2.4. Additional Candidates for Plume Push and Orogenic Gold

Preliminary analysis undertaken by [171] as part of an undergraduate thesis, and summarized by [170], identified several additional intervals in which plume/ LIP activity may be temporally associated with distal orogenic gold mineralization. These potential case histories include:

1. Progressive breakup of the Rodinia supercontinent between ca. 820 and 720 Ma [231] and broadly coeval ca. 820–650 Ma gold mineralization in the Yenisei Ridge of western Siberia;
2. plume interaction with the Nuna–Columbia supercontinent at ca. 1775–1730 Ma and approximately coeval gold mineralization in Laurentia, including the Homestake and Seabee districts; and
3. breakup or reorganization of an Archean supercontinent between ca. 2120 and 2060 Ma, broadly synchronous with major orogenic gold mineralization in the West African, Amazonian and São Francisco cratons.

These preliminary observations identify additional case histories for evaluation whether the plume push mechanism may have contributed to the activation of orogenic gold systems.

Each case requires more detailed evaluation of LIP and mineralization ages, paleogeographic reconstructions, predicted stress-transfer paths, changes in plate-boundary kinematics and the timing of shear-zone reactivation. We introduce these examples here to define directions for further evaluation.

4.3. Link between Mantle Plumes and Porphyry Copper Systems

We next consider whether plume push can contribute to the development of porphyry copper systems by generating rapid lithospheric stress changes that influence magma storage, ascent and hydrothermal fluid release at convergent plate margins.

4.3.1. Porphyry Copper Systems

Porphyry Cu systems are large magmatic–hydrothermal deposits formed mainly in subduction-related arcs and are the world's principal sources of Cu and Mo, commonly with economically important Au and Ag [232, 233]. Mineralization develops around shallow felsic to intermediate intrusions, where exsolved magmatic fluids produce stockwork veins, breccias, disseminated ore and characteristically zoned hydrothermal alteration [234].

Porphyry formation depends on magma fertility, volatile accumulation, deep crustal storage and structurally controlled ascent. Fertile systems commonly involve oxidized, hydrous arc magmas that undergo prolonged differentiation and volatile enrichment in lower-crustal or Moho-level reservoirs before ascending into the upper crust [235–237]. Deposit clusters are strongly influenced by crustal- to lithosphere-scale structures, including arc-transverse faults, inherited sutures, terrane boundaries and structural intersections, particularly where these are reactivated during changes in regional stress [238–240].

A particularly relevant model involves a two-stage stress evolution:

1. During prolonged horizontal compression, buoyancy-driven felsic dyke propagation may be inhibited, promoting storage, replenishment, differentiation and volatile enrichment of hydrous arc magmas and development of Cu-fertile geochemical characteristics [235, 236].
2. Subsequent waning compression, stress rotation, reversal or transient local extension may reactivate deep structures and permit rapid ascent of evolved, volatile-rich magma and fluids [173, 239–241]. Upper-crustal emplacement then promotes volatile saturation, overpressure, hydraulic fracturing, stockwork development and hydrothermal alteration. Thus, magma fertility may develop during prolonged deep storage, whereas ore formation can be triggered or facilitated by a later change in the regional stress regime.

4.3.2. Plume Push as a Possible Trigger for Porphyry Formation

We propose that abrupt stress changes associated with the second stage of porphyry development described by [235, 236] need not arise solely from local plate-boundary processes, such as changes in convergence, slab rollback or dip, slab tearing or breakoff, and ridge–trench or collisional events (e.g., [228, 229, 242, 243]), but may also be triggered or amplified by distal plate-scale forcing such as plume push.

In this model, plume-head emplacement generates a rapid stress perturbation, or plume push, that propagates through the plate and modifies the stress regime and kinematics of a distal convergent margin. Such changes could reactivate favourable crustal- to lithosphere-scale structures and facilitate release and ascent of evolved

magma along arc-transverse faults, inherited sutures, terrane boundaries and structural intersections [173, 239–241]. Upper-crustal emplacement would then promote decompression, volatile exsolution, hydraulic fracturing and focused hydrothermal fluid flow.

In this interpretation, plume push does not generate all components of a porphyry system. Magma fertility remains dependent on arc magmatism, source composition, oxidation state, water content and prolonged deep-crustal differentiation. Rather, plume push provides a possible external trigger that ‘suddenly’ (in a geologic sense) changes the timing of magma release and the permeability of the crustal plumbing system.

The mechanism is broadly analogous, in terms of the role of plume-induced stress perturbation, to that proposed above for orogenic gold systems (Section 4.2). In both cases, plume-head arrival and spreading may transmit a relatively rapid stress perturbation to a distant plate boundary and alter the kinematics of major structures. For orogenic gold systems, this may promote fault reactivation and mineralizing fluid flow; for porphyry systems, it may facilitate the release and ascent of differentiated magma and volatile-rich fluids from deep crustal reservoirs.

As a preliminary case study, we examine the possible temporal and geodynamic relationship between porphyry Cu–Au mineralization at Oyu Tolgoi in the Central Asian Orogenic Belt and emplacement of the ca. 380–360 Ma

Yakutsk–Vilyui LIP on the adjacent Siberian craton (Figure 12).

4.3.3. Oyu Tolgoi Porphyry Mineralization in the Central Asian Orogenic Belt

During the Late Devonian, southern Central Asia consisted of a complex assemblage of oceanic arcs, microcontinents, and intervening branches of the Paleo-Asian Ocean between the Siberian craton and Kazakhstan and other continental blocks (Figure 12) [203, 244–246]. In Figure 12, the TM label is retained from the [203] reconstruction for the Tuva–Mongol arc/microcontinental assemblage. Progressive subduction and closure of these oceanic domains generated widespread Devonian–Carboniferous arc magmatism [244–246] and associated porphyry mineralization [203]. Among these systems, Oyu Tolgoi and Tsagaan Suvarga are of particular interest here because they are well dated and record a closely timed Late Devonian porphyry-forming episode at ca. 372–370 Ma. Both deposits occur within the Gurvansayhan island-arc terrane, a segment of the broader Kazakh–Mongol arc system [70, 247, 248]. Geochemical and isotopic characteristics of the Devonian magmatic rocks indicate predominantly juvenile, intra-oceanic arc magmatism, although paleogeographic and zircon constraints suggest that this arc system lay relatively close to the southern margin of the Siberian craton [203, 248].

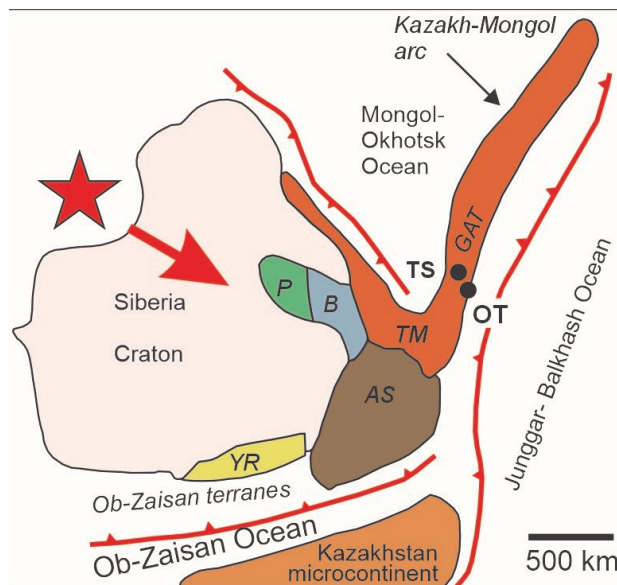


Figure 12. Schematic Late Devonian–Carboniferous tectonic reconstruction of the Central Asian Orogenic Belt, modified after [203], emphasizing the paleotectonic setting of the ca. 372–370 Ma Oyu Tolgoi (OT) and Tsagaan Suvarga porphyry (TS) systems. OT and TS are located within the Gurvansayhan arc terrane (GAT), part of the broader Kazakh–Mongol arc system developed during progressive subduction and closure of Paleo-Asian oceanic domains. The Junggar–Balkhash Ocean lay outboard of the arc to the east, whereas the Ob–Zaisan Ocean and associated terranes occupied the region between the Siberian craton and Kazakhstan microcontinent. The inferred centre of the ca. 380–360 Ma Yakutsk–Vilyui LIP/plume is superimposed after [249–251]. The large red arrow shows the approximate proposed direction of maximum plume-push forcing. See details in text. YR = Yenisey Ridge; AS = Altai–Sayán; P = Patom fold-and-thrust belt; B = Barguzin block; TM = Tuva–Mongol arc/microcontinental assemblage, retaining the label used by [203] for this part of their schematic reconstruction; GAT = Gurvansayhan arc terrane; OT = Oyu Tolgoi; TS = Tsagaan Suvarga.

The Oyu Tolgoi district comprises several giant porphyry Cu–Au–Mo centres distributed along an approximately 12-km-long mineralized structural corridor [70]. Mineralization is principally associated with Late Devonian quartz-monzodiorite intrusions. U–Pb zircon ages from pre- to syn-mineral intrusions span approximately 374–368 Ma, with the principal mineralizing episode concentrated at ca. 372–370 Ma [70, 248]. Younger, ca. 365 Ma post-ore granodioritic intrusions followed major deformation and the principal mineralizing event. Collectively, the Oyu Tolgoi deposits contain >42 Mt Cu and approximately 1850 t Au [70, 252].

Farther along the same Gurvansayhan island-arc terrane (Figure 12), the separate Tsagaan Suvarga Cu–Mo porphyry deposit has been described in detail by [253] and has a well-constrained molybdenite Re–Os age of 370.4 ± 0.8 Ma [254], providing a particularly close temporal match with the principal Oyu Tolgoi mineralizing interval. Together, the Oyu Tolgoi and Tsagaan Suvarga districts define a significant ca. 372–370 Ma porphyry-forming episode within the southern Central Asian Orogenic Belt.

4.3.4. Yakutsk–Vilyui Large Igneous Province

The Yakutsk–Vilyui LIP (YV-LIP), located near the eastern margin of the Siberian craton, comprises a giant radiating mafic dyke swarm, mafic sills and volcanic rocks, and a multi-armed rift system converging toward a well-defined plume-centre region [249, 250, 255, 256]. The southwest-trending Vilyui rift is the principal arm, and the province is also associated with carbonatites and highly diamondiferous kimberlite fields. Two main magmatic pulses are recognized at ca. 380–375 Ma and ca. 365–360 Ma, including precise U–Pb ages of 377.72 ± 0.12 Ma and 366.14 ± 0.28 Ma, respectively [250, 256–259]. These two pulses may record separate episodes of plume-related forcing, with initial plume-head arrival followed by renewed or plume-assisted rifting and magmatism, consistent with geological and geodynamic modelling of the Vilyui rift system [256, 260].

4.3.5. Proposed Yakutsk–Vilyui Plume-push Link to Oyu Tolgoi Mineralization

The ca. 372–370 Ma Oyu Tolgoi–Tsagaan Suvarga porphyry episode falls between the two principal magmatic pulses of the ca. 380–360 Ma Yakutsk–Vilyui LIP and therefore overlaps the broader interval of inferred plume-head activity beneath the Siberian craton (Figure 12). We hypothesize that lateral spreading of the Yakutsk–Vilyui plume head imposed an additional far-field stress perturbation on an already active subduction-related arc system and thereby contributed to this short-lived episode of porphyry emplacement and mineralization.

At ca. 370 Ma, the Oyu Tolgoi-bearing Gurvansayhan arc terrane had probably not yet accreted to the Siberian continental margin but lay offshore within the Paleo-Asian Ocean [245, 246, 261–263]. The intervening oceanic sep-

aration was likely relatively narrow, perhaps a few hundred kilometres, although its precise width is reconstruction-dependent [70, 248, 264, 265].

As discussed in Section 4.1, plume-head spreading can generate approximately radial asthenospheric flow, but the overlying lithosphere responds as a mechanically coherent plate. Where the expanding Yakutsk–Vilyui plume head encountered the thick lithospheric root of the Siberian craton (e.g., [266]), the resulting plume push would have been asymmetric, with the strongest net force directed toward and through the cratonic root. We infer that this preferential forcing was approximately toward the southern margin of Siberia and the adjacent Oyu Tolgoi-bearing arc system. The red vector in Figure 12 therefore shows a first-order inferred direction of maximum plume push rather than a precisely reconstructed plate-motion vector.

Such a transient plume-push perturbation could have modified the broader regional plate-motion and stress regime, potentially including a change in the motion of the Siberian craton (red arrow in Figure 12) and consequent reorganization of the surrounding plate-boundary system. Given uncertainties in the ca. 370 Ma paleogeographic configuration and the intervening oceanic domains, we do not specify the exact pathway by which this forcing would have been transmitted to the Oyu Tolgoi–Tsagaan Suvarga arc system. Nevertheless, such regional reorganization could have influenced structural reactivation, magma storage and release, porphyry emplacement, and hydrothermal focusing at Oyu Tolgoi and the coeval Tsagaan Suvarga system. The porphyry magmatism itself remained fundamentally subduction-related; plume push is proposed here as a possible additional tectonic trigger, not as the source of the magmas. This example is therefore presented as a plausible geodynamic hypothesis, with the close temporal association between Yakutsk–Vilyui plume activity and the short-lived ca. 372–370 Ma porphyry-forming episode providing a basis for further evaluation.

4.3.6. Testing the Plume-push Hypothesis in Other Porphyry Systems

The broader hypothesis is that some short-lived porphyry-forming events may have been triggered or assisted by distal stress perturbations associated with plume-head arrival and spreading. Simple temporal comparison between porphyry ages and the global LIP record is not diagnostic, because LIPs are numerous, many plume centres have been proposed, and paleogeographic reconstructions become increasingly uncertain with age [7, 10]. More rigorous tests should demonstrate a plausible mechanical linkage between a specific plume event, the resulting plate-scale stress or kinematic change, and a temporally corresponding porphyry-forming episode. Oyu Tolgoi is therefore highlighted as a plausible example and a framework for future testing, rather than as proof of the mechanism.

4.4. Plume-push Model and Exploration Implications

We propose that mantle plume-head arrival and spreading can generate geologically rapid stress perturbations that are transmitted through lithospheric plates and, in some cases, expressed at plate boundaries thousands of kilometres from the plume centre (Section 4.1). Such externally imposed stress changes may modify fault kinematics, crustal permeability, and the timing of magma or fluid release, thereby providing a potential trigger for mineralization in already favourable tectonic settings.

For orogenic gold, this model builds on the proposal of [47] linking Ontong Java plume-head arrival with Early Cretaceous plate reorganization and broadly coeval gold mineralization in eastern Asia; other possible plume–gold associations identified by [170, 171] remain to be tested. For porphyry systems, we propose a possible plume-push link between the Yakutsk–Vilyui plume of eastern Siberia and Oyu Tolgoi in the Central Asian Orogenic Belt.

Plume push does not replace established mineral-system controls. Rather, it may act as an external trigger that promotes structural reactivation, magma release and hydrothermal fluid flow where favourable structures, fluid or magma sources, and appropriate depositional conditions already exist. Similar stress-triggering effects may also be relevant to other stress-sensitive mineral systems, including epithermal Au–Ag, skarn, IOCG and structurally controlled base-metal deposits. From an exploration perspective, the model can be tested by integrating high-precision LIP and mineralization ages with paleogeographic reconstructions, reconstructed plume centres and plume-push directions, and the orientation and reactivation history of crustal- to lithosphere-scale structures in prospective mineral belts.

5. Conclusions

This study expands the potential metallogenic significance of plume-generated Large Igneous Provinces (LIPs) beyond their well-established links with magmatic sulphide deposits. We emphasize the dynamic effects of plume emplacement—including domal uplift, lateral plume-head spreading, stress reorganization, thermal and magmatic input, and plume-head push—and identify several ways in which these processes may influence mineral systems both above and near plume heads and at distant plate boundaries.

Proximally, plume-related stress changes may control emplacement of chonolith-hosted magmatic sulphides, while domal uplift and basin tilting can redirect basinal fluids and influence sediment-hosted mineralization. Plume-derived heat and magmatism may also be focused along translithospheric structures, enhancing crustal melting and hydrothermal activity, and plume-generated radial and circumferential fracture systems may provide large-scale pathways for magma and mineralizing fluids.

Distally, plume-head spreading and associated plume-driven asthenospheric flow may perturb plate-scale

stress and motion, reactivating major structures and potentially contributing to the timing of orogenic gold mineralization and, more tentatively, porphyry systems.

These relationships are not proposed as universal genetic models, but as additional plume-related controls that can be integrated with established mineral-system frameworks. Testing them requires robust plume-centre identification, high-precision geochronology, paleogeographic reconstruction, structural analysis and geodynamic modelling. Thus, plume-generated LIPs may represent an underutilized framework for linking mantle dynamics, lithospheric stress, and the timing and localization of diverse mineral systems.

Author Contributions

R.E.E.: Conceptualization, Methodology, Visualization, Writing—Original Draft Preparation; Writing—Review & Editing; H.E.B.: Conceptualization, Methodology, Visualization, Writing—Original Draft Preparation, Writing—Review & Editing; S.M.J.: Conceptualization, Methodology, Writing—Review & Editing. All authors have read and agreed to the published version of the manuscript.

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Conflicts of Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Use of AI and AI-assisted Technologies

During preparation of this manuscript, the authors used OpenAI ChatGPT (GPT-5.6 Sol) to assist with language editing, organization, shortening, and presentation of concepts, and to help identify potentially relevant literature. All AI-assisted text and references were reviewed and verified by the authors. All scientific concepts, interpretations, figures, and conclusions were developed and critically evaluated by the authors, who take full responsibility for the manuscript.

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