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Review Article

Lithospheric weakspots, not hotspots: New England-Quebec and Shenandoah anorogenic magmatism in the context of global plate tectonics, intraplate stress and LIPs

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ABSTRACT

We explore the origins of anorogenic post-breakup magmatism in two areas of the mid-Atlantic Appalachians: the New England-Quebec Province (ca. 130–120 Ma) and the Shenandoah Province (ca. 49–47 Ma). Radiometric rock ages and other data do not support claims that this magmatism occurred when these sites were located above postulated Great Meteor and Bermuda mantle hotspots/plumes. We propose instead that the sites are persistent lithospheric 'weakspots' favorable for magma ascent during relatively short intervals of a few Myr when global-scale plate motion reorganizes every 20–30 Myr. Magma ascends into the crust when compressive intra-plate stress is relaxed. Weakspots in the plate, not fixed mantle hotspots, can explain why anorogenic magmatism occurred at the same two sites also much earlier (by ca. 50 Myr in the New England-Quebec province and ca. 100 Myr in the Shenandoah Province), and why the Bermuda volcanoes formed not later, but coevally with the Shenandoah Province, 1400 km along the postulated hotspot trace. The plume hypothesis also fails to explain why the New England-Quebec magmas were emplaced at the same time as anomalously productive magmatism along the northern mid-Atlantic Ridge and coincident with the breakup of Iberia from the Grand Banks, sites almost 2000 km distant from the New England-Quebec Province. Moreover, New England-Quebec radiometric age distributions suggest that distant magmatic events and continental breakup affecting other plates were global plate reorganization events that may be 'recorded' by volcanism at weakspots. Shenandoah-Bermuda magmatism happened during the Pacific plate motion change recorded by the Hawaii-Emperor Bend. The ca. 720 Ma Robertson River Igneous Suite of anorogenic plutons in Virginia, USA, may be an old analog of the Shenandoah Province exploiting the same lithospheric weakspot. The New England-Quebec magmatic period 130–120 Ma is also the time over which the geomagnetic reversal frequency slowed, reaching zero at the onset of the Cretaceous normal superchron (C34n) at ca. 120 Ma. This event was recorded at the mid-Atlantic Ridge axis as the J-Anomaly Ridge and a large increase in spreading half-rate from 1 to 2.5 cm/a. Thus, geomagnetic reversal frequency may also be related to plate tectonics.

1. Introduction

Plate tectonics has dominated the Earth's geological evolution for a significant part of the planet's history (e.g., Whitmeyer et al., 2023). Evolving from Wegnerian continental drift, the plate tectonic paradigm shift revolutionized geoscientific thinking in just a few years starting in 1968. Not long thereafter, Morgan (1971) theorized that narrow hot mantle plumes, rising from the deep mantle cause excess crustal magmatism both along plate boundaries and in plate interiors. As plates move across such fixed, independent plumes, they leave magmatic traces, the Hawaii-Emperor volcano chain being the most notable. Like plate tectonics, the fixed hotspot concept also had its antecedent (Wilson, 1963). Unlike plate tectonics, however, Morgan's mantle plume hypothesis has remained mired in controversy for over half a century, from its beginning to the present day (e.g., Anderson, 2013; Foulger, 2010; Foulger, 2021). Rock magnetization data from the Emperor seamounts (Tarduno et al., 2003) showed that even the 'poster child' hotspot trace was not left by a hotspot fixed relative to Earth's rotation axis. Many model modifications and special pleading were subsequently proposed to fit these and other observed data to hotspot

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models. For example, in some cases the narrow, original-model plume morphed from ca. 50–100 km to thousands of kilometers in width (e.g., Merle et al., 2019) or was tilted to lie outside the volume beneath the surface volcanism where seismic studies failed to find it.

In the current paper we review data and theories for the anorogenic magmatic province comprising the ca. 47–49 Ma intrusives in Virginia, USA (the Shenandoah Province–SP) and the primarily 130–120 Ma volcanics in the New England-Quebec (NEQ) area (Figs. 1A,B and 2). We discuss *syn*- and pre-breakup magmatism (240–170 Ma, e.g., Eby, 1987; Eby et al., 1992; Kinney et al., 2022) only where it occurred in the same area as NEQ magmatism. We also omit Triassic rift-related magmatism, except to note that a) magmatism and rifting commonly precede breakup by many tens of millions of years, and b) post-breakup magmatism was not widely associated with older Central Atlantic Magmatic Province (CAMP) and Triassic structures and intrusives.

We evaluate the fixed hotspot/plume model and compare its performance with a 'weakspot' model whereby weak zones in the lithosphere permit opportunistic magmatism. The term 'weakspot' joins a growing family of 'spot' members that include 'crackspots','notspots' and 'wetspots' (e.g., Metrich et al., 2014). We also re-evaluate a suggestion made a half century ago that plate tectonics may have recorded changes in reversal frequency.

2. History, background and context

The NE seaboard of North America is host to *syn*-, co- and post-breakup magmatism, most notably in the SP and NEQ igneous provinces. This magmatism represents the Pangea breakup phase of the present Wilson cycle. Fragmentation and dispersal was preceded by ca. 50 Myr of episodic intracontinental rifting, and magmatism began when compression was replaced by extension during the Permo-Triassic (Ma et al., 2023). Final breakup of NW Africa from eastern North America is dated at ca. 175 Ma (e.g., Klitgord and Schouten, 1986).

Named after their first recognition by Wilson (1968), Wilson cycles are ca. 500–600 Myr cycles that involve continental fragmentation, dispersal, and reassembly as supercontinents. If our weakspot concept is valid, it should influence the dispersal stages of earlier Wilson cycles (e.g., Schiffer et al., 2020). We show that the ca. 720 Ma (late Proterozoic) Robertson River Igneous Suite pluton line (Southworth et al., 1993) in the same region may represent a possible weakspot activated during the breakup of Rodinia.

A number of different, and for many features conflicting, hotspot lists have been drawn up over the five decades since the paper of Morgan (1971). Anderson and Schramm (2005) graded each hotspot according to how well it conforms to the criteria. Courtillot et al. (2003) propose five diagnostic criteria: 1) a time progressive track; 2) associated flood basalts; 3) a large buoyancy flux; 4) high He^3/He^4 ratios; and 5) an underlying seismic anomaly. They apply these measures to 49 hotpots and find that only nine arguably arise from the core-mantle boundary. The Bermuda and Great Meteor hotspot (also called the Corner hotpot) fail to meet most diagnostic criteria.

Seismology has failed to resolve mantle plumes and in any case cannot resolve features smaller than ~1000 km in the deep mantle (Hwang et al., 2011). Anderson and Natland (2005) challenge the notion that narrow mantle convection is physically possible. According to Anderson (2004) mantle convection related to plate tectonics is confined to the upper 1000 km, while deeper mantle convection is sluggish with mantle overturns occurring only on the timescale of Ga. These views are inconsistent with plumes rising from the core-mantle boundary and with geomagnetic reversal frequency changes being caused by changes in core heat loss (e.g., Loper, 1992).

2.1. The weakspot hypothesis

We describe our weakspot hypothesis as follows. Magmatic activity is suppressed by horizontal compression. As continental fragments raft apart, passive margins and spreading ridges develop in the voids and new crust and mantle lithosphere forms. Ridge-push compression is then imposed on the developing passive margins. Anorogenic magmatism occurs at weakspots when this compression is episodically relaxed or replaced by extension.

To test this hypothesis we examine the geochronology and location of sporadic and localized post-breakup magmatism along the eastern margin of North America from Nova Scotia to Florida (Fig. 1; de Boer et al., 1988). This magmatism occurred mostly ca. 500 km west of the continent-ocean boundary. Structure there was created in four orogenies, reflecting complex ocean basin closures in the period ca. 500–250 Ma.

NEQ magmatism comprises Cretaceous magmatism in the Monteregian Hills (MH–Quebec), and White Mountains (WM–New England area; Fig. 2). The Lower Cretaceous magmatism (e.g., Eby, 1984, 1987; Eby et al., 1992; de Boer et al., 1988; McHone, 1988; McHone and Butler, 1984; Kinney et al., 2021) has been attributed to the Great Meteor Hotspot, a putative mantle hotspot above which the NEQ area was proposed to have been located at ca. 120–130 Ma (Morgan, 1972; Crough, 1981; Duncan, 1984). The current proposed hotspot location, below the Great Meteor Seamount (Fig. 1A), is about 4100 km from the NEQ. The SP is a small area of Eocene (47–49 Ma) intrusions in Virginia, USA (Fullagar and Bottino, 1969; Southworth et al., 1993; Mazza et al., 2014) which has been attributed to a Bermuda mantle plume (Morgan, 1983; Duncan, 1984).

Crustal and mantle lithosphere properties and contemporary seismicity in the NEQ and SP areas are all anomalous. Many characteristics, e.g., composition and structure, are likely preserved over time in plate interiors. Melt formation, segregation and movement could, however, alter properties over time. As a result of long periods of erosion (e.g., Portenga et al., 2013) outcrops sampled for radiometric dating are mostly intrusive. Original eruptives, if any, are largely gone but the many plutons we discuss likely once fed volcanoes. Original erupted volumes are difficult to estimate, however.

Oceanic- and intraplate magmatism east of the NE seaboard comprises the New England and Fogo Seamounts (NES and FS), Bermuda Volcanoes (BV), the paleo-spreading axis J-Anomaly Ridge (JAR) and the aseismic Southeast Newfoundland Ridge (Figs. 1A,B). The cost and challenges of ocean-floor basement coring has limited sampling and dating of those features so much of what is known is still based on DSDP Leg 43 (Tucholke and Vogt, 1979) drill sites and samples dredged from the NES (Duncan, 1984; Merle et al., 2019).

The magmatism of the NEQ-SP province is more consistent with periodic short-lived (ca. 1–5 Myr) intraplate episodes of stress relaxation (McHone, 1988; Southworth et al., 1993) than with a mantle plume. Some volcanism is longer-lived (10s of Myr) and uniform in composition, an observation also at odds with the plume model unless the plate is fixed relative to the hotspot. Short-lived passive margin tension episodes, likely associated with global plate motion re-organizations every ca. 25–30 Myr, better account for the observations as we argue below.

Several authors have suggested episodic tectonic time scales of 25–30 Myr, including for geomagnetic reversal frequency, paleoclimate, mass extinctions (Loper et al., 1988, and references therein), and Large Igneous Province emplacements (LIPs; Ernst et al., 2020). A full statistical study has not yet been done. Passive margin magmatism may be discouraged during long periods of ridge-push compression and occur only where the crust and mantle lithosphere are weak. The ages of NEQ-SP igneous activity should thus record intervals of tension that facilitate intrusion, but only where the crust and mantle lithosphere are weak along pre-existing fractures (e.g., Faure et al., 2006).

2.2. Recent advances in radiometric dating and geochronology

Our present review has been partly motivated by the explosive increase of accurate radiometric dating and other geological information over the last two decades. Many of these data, interpretations, reviews, and syntheses are published in topical volumes, notably GSA Special Papers 388 (2005), 505 (2014) and 511 (2015), GSA Memoir 229 (2023) and the recent Geologic Time Scale of Ogg (2020). References in these publications provide a road map to many earlier relevant works.

In examining the radiometric geochronology, we note correlations in time between anorogenic magmatism along the NE passive margin and magmatic events in other parts of the world (e.g., Chilwa alkaline province in Malawi). It is well known that orogenic tectonism and magmatism, e.g., along the Andes, correlates in time and space but it is unknown to what extent global correlations in time are simply co-incidences or imply common causes. Correlations in time between geomagnetic reversal frequency and changes in the kinematics of

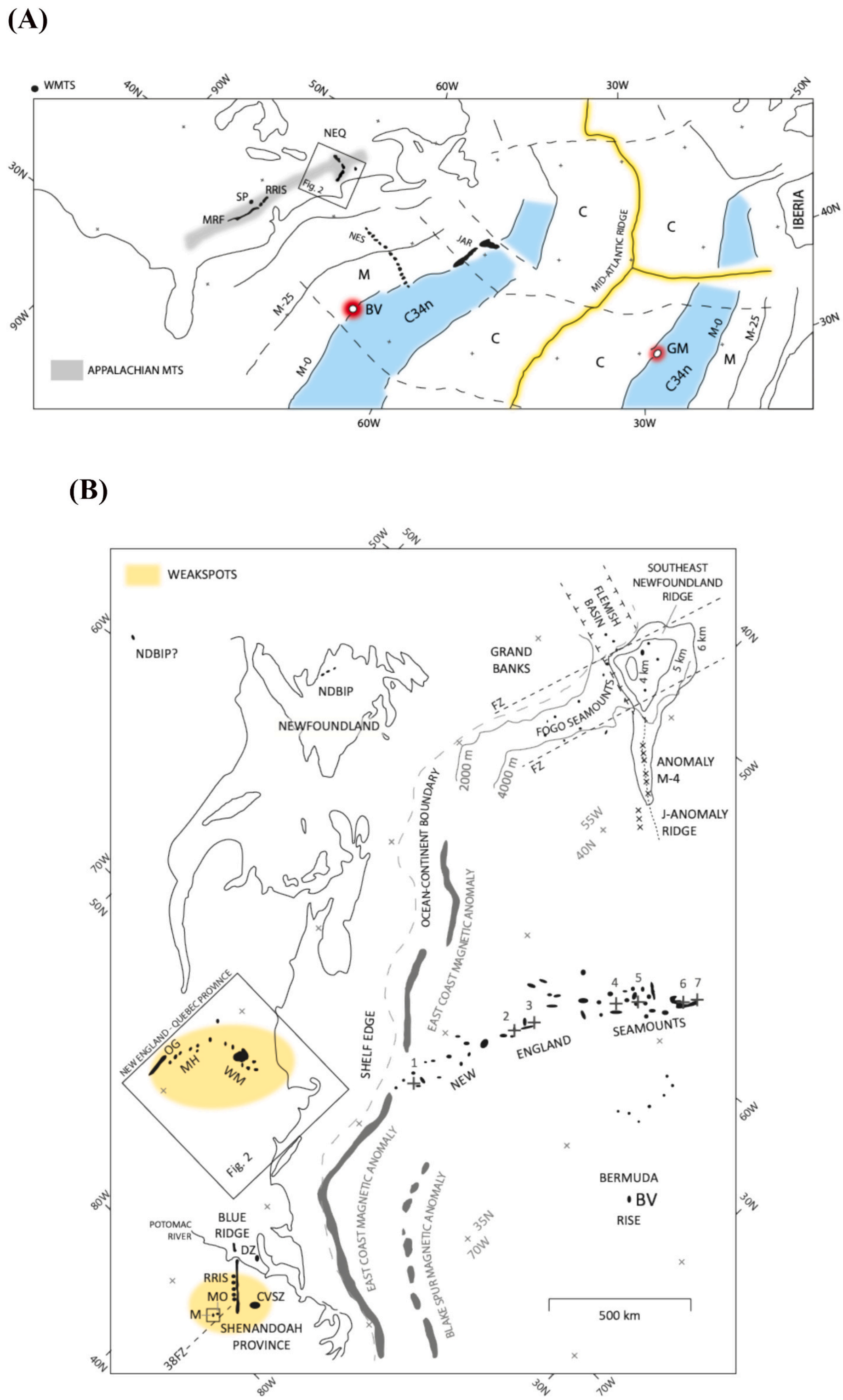

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Fig. 1. A. Regional relationships on transverse Mercator projection. MRF, Mt. Rogers Formation; SP, Shenandoah Igneous Province; RRIS, Robertson River Igneous Suite; NEQ, New England-Quebec Igneous Province; NES, New England Seamounts; M, M-Series aged oceanic crust; BV, Bermuda Volcanoes and present location of postulated Bermuda Hotspot; JAR, J-Anomaly Ridge and Southeast Newfoundland Ridge; Blue, C34n–oceanic crust generated during the Cretaceous Magnetic Quiet Zone; C, oceanic crust generated during C series of geomagnetic reversals; GM, Great Meteor Seamount and location of postulated Great Meteor Hotspot. Box shows area of Fig. 2. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.) B. Regional map showing New England-Quebec Cretaceous and Shenandoah-Bermuda Eocene anorogenic magmatism and related features. Grey lines, bathymetric contours labelled with depth; black lines around the Southeast Newfoundland Ridge, depth to igneous basement without sediment cover; NNDBIP, Notre Dame Bay Igneous Province, Newfoundland; FZ, Fracture Zone; BV, Bermuda Volcano; xxx, Anomaly M4; OG, Ottawa Graben; MH, Monteregian Hills plutons; WM, White Mountain batholith; DZ, Detrital Zircon; RRIS, Robertson River Igneous Suite; CVSZ, Central Virginia Seismic Zone; M, Monterey, Virginia; MO, Mole Hill; 38FZ, 38°N Fracture Zone. Age-dated New England Seamounts (Merle et al., 2019): 1, Bear, 104.3 Ma; 2, Atlantis II, 99.6 Ma; 3. Gosnold, 90.8 Ma; 4, Alleghany, 83.8 Ma; 5, Michael, 83.5 Ma; 6, Nashville, 82.7 Ma; 7, Nashville, 82.3 Ma. Modified from Fig. 1 of de Boer et al. (1988). Box shows area of Fig. 2. Yellow shading, proposed weakspots. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

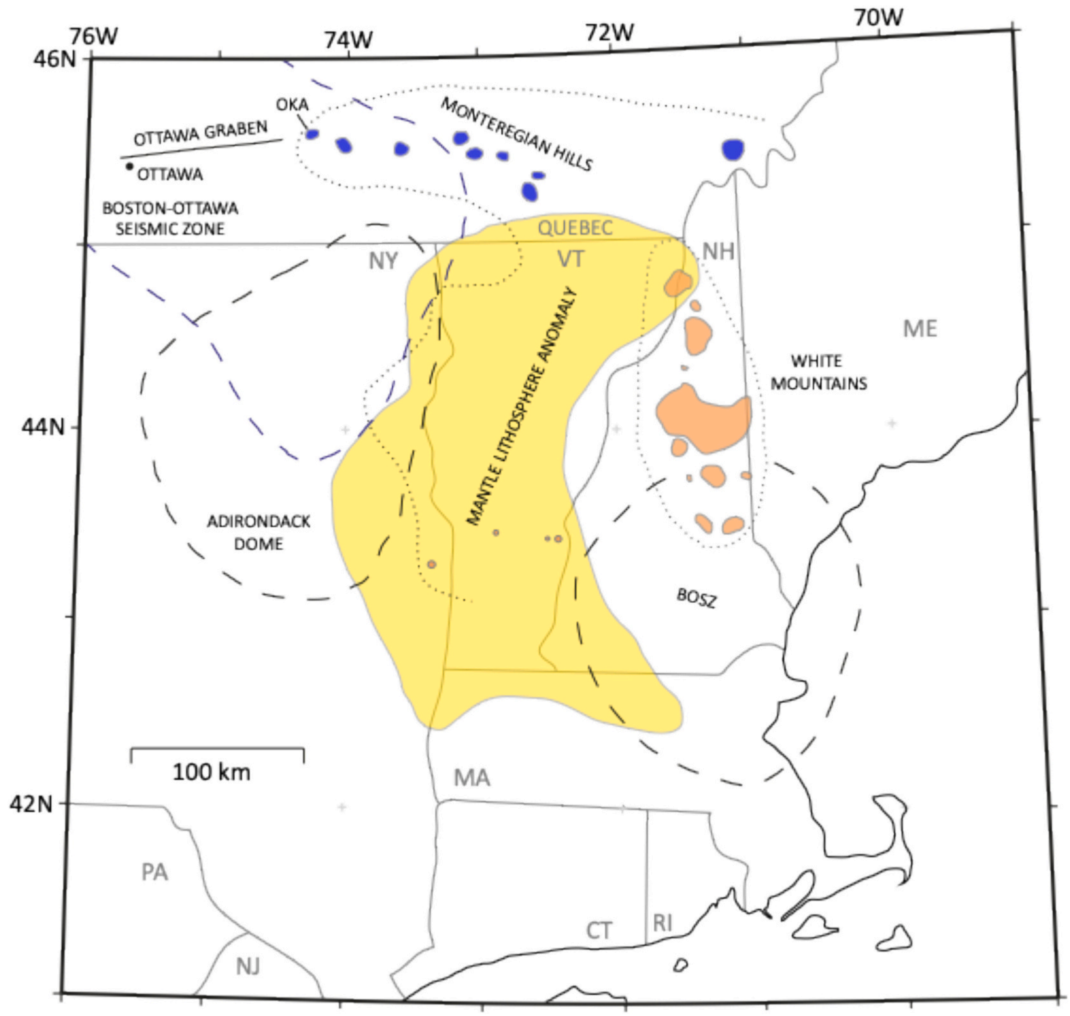

Fig. 2. New England-Quebec igneous features (pink–White Mountain, blue–Monteregian Hills) and modern non-magmatic features suggesting persistent regional lithospheric anomalies and weakness. Yellow shading indicates the area underlain by a mantle lithosphere seismic anomaly. BOSZ, Boston-Ottawa Seismic Zone (Frankel, 1995). Oka, Carbonatite Complex (Chen and Simonetti, 2014). Dotted lines outline White Mountain and Monteregian Hills magmatic zones based on scattered intrusions with compositional affinity with Monteregian Hills plutons. US State abbreviations: CT, Connecticut; MA, Massachusetts; ME, Maine; NH, New Hampshire; NJ. New Jersey; NY, New York; PA, Pennsylvania; RI, Rhode Island; VT, Vermont. Data from Bailey et al. (2017). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

tectonic plates have also been proposed (Vogt, 1975). It has been suggested that whole mantle convection changes (e.g., in the locations of cells, physical state, composition, or flux rates) could change the thermal boundary conditions at the core mantle boundary and influence the geodynamic dynamo (e.g., Vogt, 1975; Courtillot and Besse, 1987; Larsen, 1991a, 1991b; Larsen and Olson, 1991). If so, data from the NEQ-SP might reflect this.

3. Anorogenic magmatism and geophysical anomalies in the New England-Quebec-White Mountain region

Pre- and *syn*-breakup Triassic, early Jurassic and later post-breakup early Cretaceous anorogenic magmatism occurred in the NEQ igneous province (de Boer et al., 1988; Fig. 2). This province comprises the MH pluton line and the spatially and geochronologically complex WM magma series. Associated with these two linear provinces is a wider region of dikes and sills, extending into New York State (e.g., Eby et al., 1992; Bailey et al., 2017). We refer to this entire igneous province as the NEQ because of its location, largely in New England and Quebec.

The region affected by this igneous activity (Fig. 2) is about 350 × 400 km in area. Inclusion of kimberlites discussed by Crough et al. (1980) would enlarge the area, as would inclusion of the Notre Dame Bay Igneous Province (NDBIP, Fig. 1B; de Boer et al., 1988; Peace et al., 2013). The two NDBIP localities, separated by 700 km, are 1500 km from the NEQ. Radiometric dates for the NDBIP (Clarke, 1977) vary, but are all Cretaceous. If these two intrusion areas are included then there seems no reason not to also include the Fogo Seamounts (FS), J-Anomaly Ridge (JAR), and Southeast Newfoundland Ridge (Fig. 1B). That area is 2000 km from NEQ but only 800–1200 km from the NDBIP. Whether this much wider area of dispersed igneous activity is considered to be one large or several small provinces of coeval magmatism is unclear in the context of the hotspot/plume model.

The NEQ igneous province area itself (Fig. 2) comprises intrusions largely in two belts. This is inconsistent with the hotspot/plume model which predicts a single line built above a ca. 50–100 km diameter mantle plume. Special and unique variants of the plume model that could account for both the large area and localization include a large plume head with crustal intrusion guided by structures such as faults and the expected surface flood basalt now eroded away. Our weakspot model proposes control of occasional, small-volume magma ascent by crustal structures (Faure et al., 2006).

3.1. White Mountain magma geochronology

The WM igneous lineament trends ca. N07°W, extends about 160 km and is 10–60 km wide. It is dominated in its center by the WM batholith. The nearly orthogonal (N87°W) MH lineament extends for 250 km if the isolated Megantic pluton in the east is included. The other dozen or so MH plutons form a 140 km long line which can be subdivided into an east-westerly western segment and a ca. N52°W eastern segment. The 140 km long pluton line extrapolation intersects the WM line at a 110° angle. This suggests emplacement along a “ridge-ridge-ridge”-type, star-shaped tensional structure, with Megantic emplaced on the third arm.

The southern New England WM Magma Series comprises both early Cretaceous magmas and early Jurassic alkaline granitoid intrusives. The Cretaceous WM intrusions mostly date to 130–110 Ma (Fig. 2; Creasy and Eby, 1993). The youngest dated so far (the Cuttingsville complex) was intruded at 103 Ma (Boemmels et al., 2021). Three Cretaceous plutonic complexes have dates averaging 122.5 Ma (Van Fossen and Kent, 1992). Two highly precise ages (124.04 ± 0.3 Ma and 122.83 ± 0.30 Ma) were obtained by Kinney et al. (2021). Detrital zircons eroded from the White Mountains and from the Atlantic mouth of the Merrimack River record a primary abundance peak of ca. 123 Ma and a slightly younger, lesser peak (Gaschnig, 2019).

Recent radiometric ages from just the WM batholith (Kinney et al., 2021) mostly range from 198.5 to 180 Ma with the volcanics dated at 185–180 Ma. Most of the dates (Eby, 1984, 1987; McHone and Butler, 1984; McHone, 1996) fall in the range 190–160 Ma. The oldest intrusions (207 Ma) preceded the main (201 Ma) CAMP event (Kinney et al., 2021).

Cretaceous intrusions as young as ca. 115–135 Ma were emplaced in the same area as Jurassic WM granitoids as old as 207.5 Ma (Kinney et al., 2022). If the latter were caused by a fixed hotspot/mantle plume (e.g., Duncan, 1984) why was this site chosen, and why was the magmatism spread over more than 20 Myr? Crough (1981) and Morgan (1983) suggested that the plate crossed different hotspots at different times, special pleading that was disputed by McHone (1981). We argue that a persistent, repeatedly rejuvenated weakspot is a more credible explanation.

3.2. Monteregian Hills magma geochronology

The MH is a band of eight plutons, the western ones trending EW and the eastern ones NW-SE. Geochemistry shows MH pluton compositions mostly reflect mantle sources rather than fractional crystallization (Roulleau et al., 2012, Roulleau and Stevenson, 2013), but the locations of the melt sources are unknown. Foland et al. (1986) obtained an average MH age of 124 ± 1 Ma and two highly precise ages (125.47 ± 0.7 and 126.15 ± 0.3 Ma) were reported by Kinney et al. (2021). The WM Cretaceous and MH plutons, ca. 200 km apart, were thus emplaced close to the same time. Their locations are consistent with the plate motion direction (NW) but their ca. 2.5 Myr age difference is not consistent with the ca. 10 cm/a plate speed. There is a ca. 10 Myr age mismatch with the fixed-hotspot model (Kinney et al., 2021).

MH radiometric ages based on a consistent method are shown in Fig. 3 as a histogram (Bailey et al., 2017; Foland et al., 1986). Of ca. 60 dates (108–159 Ma) the few older than 150 Ma are suspect and thus not shown in Fig. 3. The dates cluster at 123–124 Ma (II in Fig. 3) with less well-defined clusters at 117–118 Ma (I) and 132–134 Ma (III). Chen and Simonetti (2014) also proposed three age clusters for the Oka complex at the western end of the MH, Oka1, Oka2 and Oka3 (Fig. 3), which they suggest involve separate magma reservoirs. Clusters Oka2 and Oka3 agree well with our stages II and III and there is a ca. 4 Myr mismatch with our poorly defined Stage I. The observations are, however, compatible with three separate stress relaxation episodes in the weakspot model.

3.3. J-Anomaly Ridge, Southeastern Newfoundland Ridge and Fogo Seamounts

The J-Anomaly Ridge (JAR; Figs. 1B, 3) is an anomalously high-relief oceanic basement ridge with a steep SW-dipping escarpment. It is associated with an anomalously high-amplitude positive magnetic anomaly (up to 1000 nT at sea level; Vogt, 1986a,b), suggesting a thickened lava layer or crustal thermo-remanent magnetizations six times typical. Analogies with high-amplitude magnetic oceanic lineations elsewhere (e.g., Vogt, 1979) suggest fractionated Fe—Ti ferrobasalts may be the source. The M-anomalies were first identified in the western Atlantic and thought to be Jurassic in age, hence the ‘J’ label.

The anomalous JAR crust formed during magnetic anomalies M-4 to M-2 (Tucholke and Ludwig, 1982) or M3-M0 (Pe-Piper et al., 2007). Association with the onset of superchron C34n and a dramatic spreading half-rate increase from 0.9 cm/a (M4-M0) to 2.4 cm/a after M-0 (Klitgord and Schouten, 1986) suggested a link between plate tectonics and geomagnetic reversal frequency (Vogt, 1975). This implies a change in location of the Nubia-North America Euler pole and/or an increase in rotation rate. DSDP Site 384 coring (Tucholke and Vogt, 1979) sampled JAR basement capped with Aptian reef sediment but recovered only altered basalt. The composition, magnetization, and thickness of the anomalous basalt thus remains speculative as does the possibly similar source of the older East Coast and Blake Spur magnetic anomalies.

Complicating interpretations of the JAR is its apparent age-progression southwestwards (at ca. 5 cm/a; Tucholke and Ludwig, 1982). Radiometric core dating of DSDP site 384 basement yielded an erroneously young (76 Ma) age due to basalt alteration (Merle et al., 2019). Based on magnetic anomalies, the JAR crust and at least part of the associated Southeast Newfoundland Rise (SENR), a ca. 2500 km^2, ≤ 2 km high, basement rise (Fig. 1B) formed at ca. 126–121 Ma on the Ogg (2020) time scale (Fig. 3). Thus there is no significant difference in age between the JAR and the MH age abundance peak of 124–123 Ma (Fig. 3) even though these features are ca. 2000 km apart.

Southwest of, but likely associated with, the JAR is a SW trending belt of scattered old volcanoes–the Fogo Seamounts (Fig. 1B de Boer et al., 1988; Pe-Piper et al., 2007). All–some of them guyots–were erupted on magnetic anomaly M-series crust, pre-JAR time, but only one (Frankfurt Seamount) has been radiometrically dated. This yielded an age of 130.3 ± 1.3 Ma from a dredged rock sample (Pe-Piper et al., 2007). This date lies in the MH age distribution (Fig. 3).

The JAR loses its topographic and magnetic identity towards the

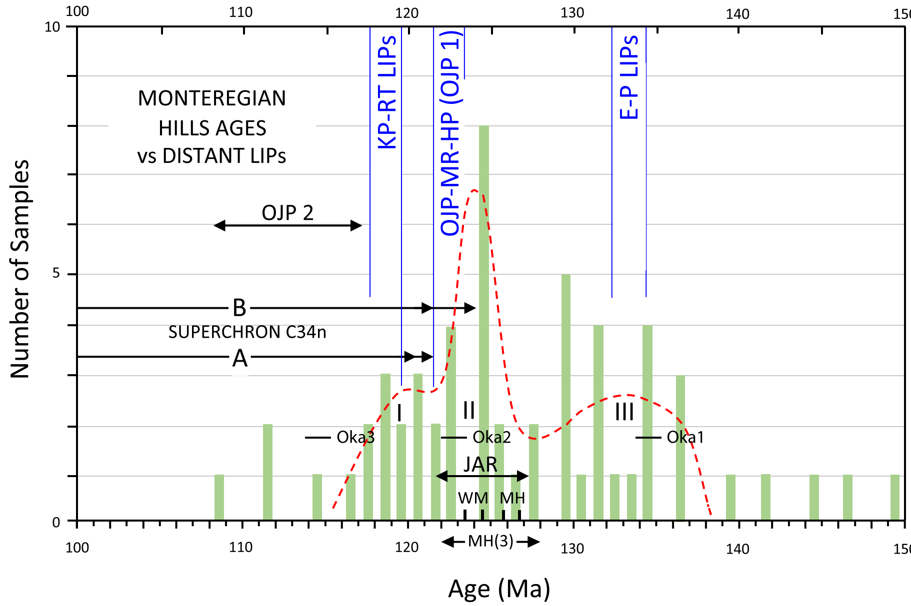

Fig. 3. Radiometric age distribution of Monteregian Hills plutons (Bailey et al., 2017) in relation to peak LIP eruption periods, associated continental breakups and related features. The histogram suggests three clusters of common dates, I, II (the most prominent) and III. OJP-MR-HP (OJP 1), Ontong Java Plateau-Manihiki Rise-Hikurangi Plateau "super-LIP" (Tejeda et al., 2002; Erba et al., 2015; Hoernle et al., 2010; Timm et al., 2011); *E*-P LIPs, Etendeka-Parana LIPs and first South Atlantic spreading (Macdonald et al., 2003); KP-RT LIPs, Kerguelen Plateau-Rajmahal Traps LIPs and Australia-India separation (Duncan, 2002; Kent et al., 2002); A, Cretaceous normal polarity superchron beginning ca. 121 Ma (Ogg, 2020) or B, beginning at 120 Ma (Li et al., 2023). OJP 2 range of dates from Davidson et al. (2023); MH, Monteregian Hills; WM, White Mountains; JAR, J-Anomaly Ridge (Tucholke and Vogt, 1979; Tucholke and Ludwig, 1982; Merle et al., 2019). Oka-1, Oka-2 and Oka-3 are three distinct magmatic episodes at the Oka Carbonatite Complex (Chen and Simonetti, 2014). MH(3), MH age range according to Boemmels et al. (2021). Data from Bailey et al. (2017).

northeast, where it merges into the SENR (Fig. 1B) This feature, which resembles a miniature Iceland, is sharply bounded on the NW by the major transform-type fracture zone, an extension of the SW-facing transform-type oceanic-continental boundary along the SE facing edge of the Grand Banks. A lesser fracture zone appears to separate the SENR from the JAR, which turns towards the north as it approaches and merges with the rise. The NE edge of the SENR is gradual, suggesting that anomalously high magma production continued into C34n superchron times.

The SENR is associated with chaotic high amplitude magnetic anomalies (Vogt, 1986c; Pe-Piper et al., 2007) and some of the Fogo Seamounts formed on it (Fig. 1B). Rise elevations suggest a crust thicker than typical for oceanic crust. Based on DSDP Site 384 and anomalously high basement topography, higher parts of the JAR and SENR stood above sea level. Along the southern (ca. 39°N) JAR seaward dipping reflectors were seismically imaged (Tucholke and Ludwig, 1982). In these respects the SENR resembles a smaller, older Iceland and Iceland-Faeroe Ridge (see Foulger et al., 2020 for a review). Nevertheless, construction by a hotspot has not been proposed for the JAR-SENR feature. Excess magma likely erupted over a wider area than is typical for a mid-ocean ridge axis and subaerial lava flowed laterally over larger distances northwest of the SENR.

Northeast of the SENR lies the deep, sediment-filled Flemish Basin (Tucholke, 1986). The crust below this basin plausibly comprises highly thinned transitional continental crust. If the steep SW-dipping JAR escarpment is extrapolated across the SENR, it is co-linear with the steep NE-dipping escarpment that extends from the Grand Banks down into the Flemish Basin. The J-anomaly has also been identified NW of the Flemish Basin, suggesting Grand Banks-Iberia breakup started just prior to 121 Ma along a non-volcanic margin (e.g., Srivastava and Tapscott, 1986; Barnett-Moore et al., 2018).

Taken together, the evidence suggests the events leading to continental breakup between Grand Banks and Iberia were preceded by enhanced mid-ocean ridge magmatism building the northern JAR and SENR around or soon after C5n (127.5 Ma on the Ogg (2020) time scale) and propagating both south along the new mid-ocean ridge (at 5 cm/a) and north into continental lithosphere. As continental rifting commenced around or soon after M0r (121.4 Ma), plate separation began, causing normal faulting along the new Grand Banks-Iberia break and the thin, stretched crust below the Flemish Basin. Similar processes have recently been proposed for Central Afar (Rime et al., 2024). Simultaneously, a rapid fall in the ratio of magma supply to opening rate along the JAR created its NE dipping slope. The timings of these events are based only on the magnetic reversal time scale, not directly on radiometric ages. Nevertheless, as discussed above, the timing is similar to MH magma emplacement ages (Fig. 3) some 2000 km away.

3.4. Post-NEQ magmatism in the wider region

Except for the SP and Bermuda, no magmatism significantly post-dating the youngest, 103-Ma, NEQ intrusives is known along the eastern US continental margin. The only magmatism of that age in the western Atlantic-eastern North America is the 1200-km-long New England Seamount chain (Fig. 1B). This chain is also the only plausible Hawaii-type time-progressive chain (Duncan, 1984).

On the basis of dredged and cored volcaniclastic debris, Duncan (1984) derived a propagation speed of 4.7 cm/a for the chain which agrees with the speed of that part of the North America plate. This speed is consistent with C34n superchron spreading rates (Klitgord and Schouten, 1986) if Africa is assumed fixed. Additional dates from these seamounts (Figs. 2) were reported by Merle et al. (2019). Nevertheless, an unambiguous age trend remains uncertain. The eastern seamount dates are 83.8 Ma (Alleghany), 83.5 Ma (Michael) and 82.7 Ma and 82.3 Ma at the two ends of the long Nashville seamount. Within the errors, these could all have erupted in a single short episode. The ca. 500-km-long western segment of the chain is on average older than the eastern segment and the three available dates (Bear–104.3 Ma, Atlantis II–99.5 Ma and Gosnold–90.8 Ma) suggest west to east younging at ca. 3–3. 5 cm/a, consistent with the hotspot hypothesis. However the data permit these seamounts to have all erupted in a single episode at around

95–100 Ma. The Cuttingsville pluton (103 Ma) was intruded at the same time as formation of Bear Seamount but more than 700 km away.

3.5. Relationship of NEQ magmatism to continental rifting and breakup

A possible connection of NEQ-region anorogenic magmatism to continental breakup has long been considered. Eby (1987) suggested that igneous ages of 200–165 Ma could relate to opening of the central North Atlantic and those of 140–110 Ma to breakup of Grand Banks and Iberia. These two post-Permian continental breakups both occurred in the general area of NEQ magmatism (e.g., Klitgord and Schouten, 1986; Müller et al., 2019). However, igneous activity at precisely 175 Ma is unknown, so any dates in the 180–170 Ma age range would satisfy the constraints. The continent-ocean transition is no closer than 500 km to NEQ magmatism, however. Africa-America breakup occurred in two phases, suggesting possibly two episodes of tension and magmatism. An early rifting episode produced a narrow Red Sea-sized ocean basin, after which the rift jumped into the continental margin, rafting a sliver of continent out into the ocean (Vogt, 1973) similar to formation of the Jan Mayen Ridge (Schiffer et al., 2019).

About 50 Myr later, breakup propagated north with Iberia separating from the Grand Banks (e.g., Srivastava and Tapscott, 1986; Barnett-Moore et al., 2018). This new breakup occurred ~2000 km away from the NEQ but was also accompanied by magmatism there. This is consistent with the weakspot concept. Widespread tension below the plate would have caused continental breakup as well as more scattered coeval anorogenic magmatism at intraplate weakspots.

Curiously, almost all NEQ magmatism clustered around the time of the more distant breakup. Magmatism plausibly coeval or just preceding Africa-North America breakup is recorded only in the 185–180 Ma dates for the moat volcanics of the WM batholith (Kinney et al., 2022) and a 179-Ma detrital zircon from the Maryland Coastal Plain (McCormick et al., 2017; P. Vogt and J. Saylor, unpublished).

3.6. Present geophysical anomalies in the NEQ lithosphere

Several geophysical features show that anomalous crust and upper mantle underlie the NEQ region, supporting the weakspot hypothesis. Processes appear to be ongoing. The Adirondack dome (Fig. 2) began to rise at ca. 180 Ma (Chiarenzelli et al., 2015) and has continued its slow ascent (0.5–1 mm/a) for the last tens of millions of years. In the same region low shear wave speeds are consistent with mantle upwelling ca. 100–300 km below the northern Appalachians around New Hampshire over the last few tens of Myr (Fig. 2) (Levin et al., 2018).

Schmandt and Lin (2014) used anomalous *P*- and *S*-wave speed based tomography to identify a larger area (Vermont, New Hampshire and Massachusetts) with anomalous mantle at 60–300 km. That body is 70–100 km wide at 75 km depth and widens at 125–200 km depth. The region exhibits low *S*-wave speeds at 195 km depth (Porter et al., 2016), low *P* and *S*-wave speeds at 60–300 km (Schmandt and Lin, 2014) and lacks the seismic anisotropy commonly observed elsewhere. It is not clear how passage over a postulated Great Meteor plume more than 100 Myr earlier could account for these anomalies (Schmandt and Lin, 2014), so the structure has been interpreted as a new mantle plume coincidentally impinging on vulnerable older mantle lithosphere. Alternatively, it may simply be part of the deeper mantle lithosphere, normal compositional variations, or result from edge convection (Levin et al., 2018).

Historical earthquakes (Fig. 2, Sbar and Sykes, 1977; Frankel, 1995) comprise a NW-SE band extending from Boston to Ottawa. This band is associated with the poorly dated pre-Cambrian Ottawa graben. It shows structural control of seismicity and likely also WM magmatism. The Boston-Ottawa seismic band is interrupted where it crosses the WM, however, suggesting either a strong brittle crust, long recurrent times, or aseismic slip. The seismicity gap corresponds to the anomalous underlying mantle (Fig. 2).

Present-day maximum horizontal compressive principal stress is orientated E to ENE (Zoback et al., 1986). This is unfavorable for magmatism. If this stress regime is dominated by ridge-push forces (e.g., Zoback et al., 1986; Zoback and Zoback, 1989) it must have formed as the young mid-ocean ridge developed in the Late Jurassic and persisted until the present. A long-term stress regime of this sort is supported by high-angle reverse faults at various sites along the mid-Atlantic continental margin (see Zoback et al., 1986 for a review). Many of these structures have been active along NE trending faults throughout post-Cretaceous time at low, long-term offset rates.

3.7. NEQ geochronology suggests regional or global synchronism of episodic intraplate magmatism

Early Cretaceous NEQ magmatism was essentially coeval with other anorogenic magmatism, anomalous mid-Atlantic Ridge magmatism, and the breakup of Iberia from the Grand Banks. Some of those features are 1000 to 2000 km from the NEQ igneous province but have ages coeval with the peak ages I, II and III shown in Fig. 3.

Within dating errors, MH episode I plutons (Fig. 3) were intruded at the same time the Kerguelen Plataeau and Rajmahal Traps Large Igneous Province (LIP; Sheth, 2007) basalts erupted (Duncan, 2002; Kent et al., 2002). Those and other LIPs are shown on a modern Earth in Fig. 4. MH episode III correlates within error with emplacement of the Etendeka and Parana LIP basalts (Macdonald et al., 2003). Given the time period in question was less than 5 % of the last 500 Ma, there was disproportionately copious anorogenic magmatism in this interval. Eruption of both the Kerguelen-Rajmahal Traps and the Etendeka-Parana LIPs likely exploited pre-existing weakspots (e.g., Foulger, 2018). A further example is the Chilwa province in Malawi (Eby et al., 1995), where magmatism began about 133 Ma, peaked again with instrusion of syenites 126 Ma, and ended with emplacement of granite plutons at 113 Ma.

The Ontong Java Plateau (OJP), thought to be the largest mass of submarine flood basalt, erupted close in age to the MH episode II (Tejada et al., 2002). Taylor (2006) suggested that the OJP, Hiturangi Plateau (Hoernle et al., 2010) and Manihiki Rise (Timm et al., 2011) were once part of a single OJP-Hikurangi-Manihiki "super-LIP" which was split apart by plate motion. Its origin is contested because the large amount of precursory uplift predicted for a plume is lacking (Korenaga, 2005) and the magma geochemistry is consistent with derivation from recycled ancient crust (Ishikawa et al., 2007).

There are two main views of the age of the OJP -Hikurangi-Manihiki super-LIP. The first is 121–123 Ma (OJP 1, Fig. 3). This is based on a date of 121 Ma for OJP (Tejada et al., 2002; Ernst et al., 2020), 121.6 ± 1.6 for the Manihiki Rise, and 118 Ma for the Hikurangi Plateau (Hoernle et al., 2010) (Figs. 3 and 4). A more recent determination of OJP age by Davidson et al. (2023), based on highly accurate Ar—Ar dating of plausibly the youngest 25 % of the lava pile, suggests dates of 108–116 Ma (OJP 2, Fig. 3). This is consistent with the normal polarity of OJP which implies eruption in the early part of C34n. In contrast, MH rock magnetizations of both polarities occur indicating that some were emplaced prior to C34n (Fig. 3). Much younger dates for OJP lavas (90 Ma; Tejada et al., 2002) and Hikurangi Plateau lavas (96 Ma; Hoernle et al., 2010) suggest that some magmatism continued over time.

If, as we propose, the crust and mantle lithosphere beneath the NEQ area has remained weak and vulnerable to brittle upper crustal failure, reactivation during major, global-scale plate motion changes and LIP emplacement is expected. The coincidence in age between NEQ and OJP magmatism provides qualitative support for this.

4. The Shenandoah-Bermuda Eocene magmatic episode

Two sites of Cenozoic magmatism are known from the eastern passive margin of North America and adjoining Atlantic basin west of the mid-Atlantic Ridge. They are the ca. Middle Eocene SP and the BV.

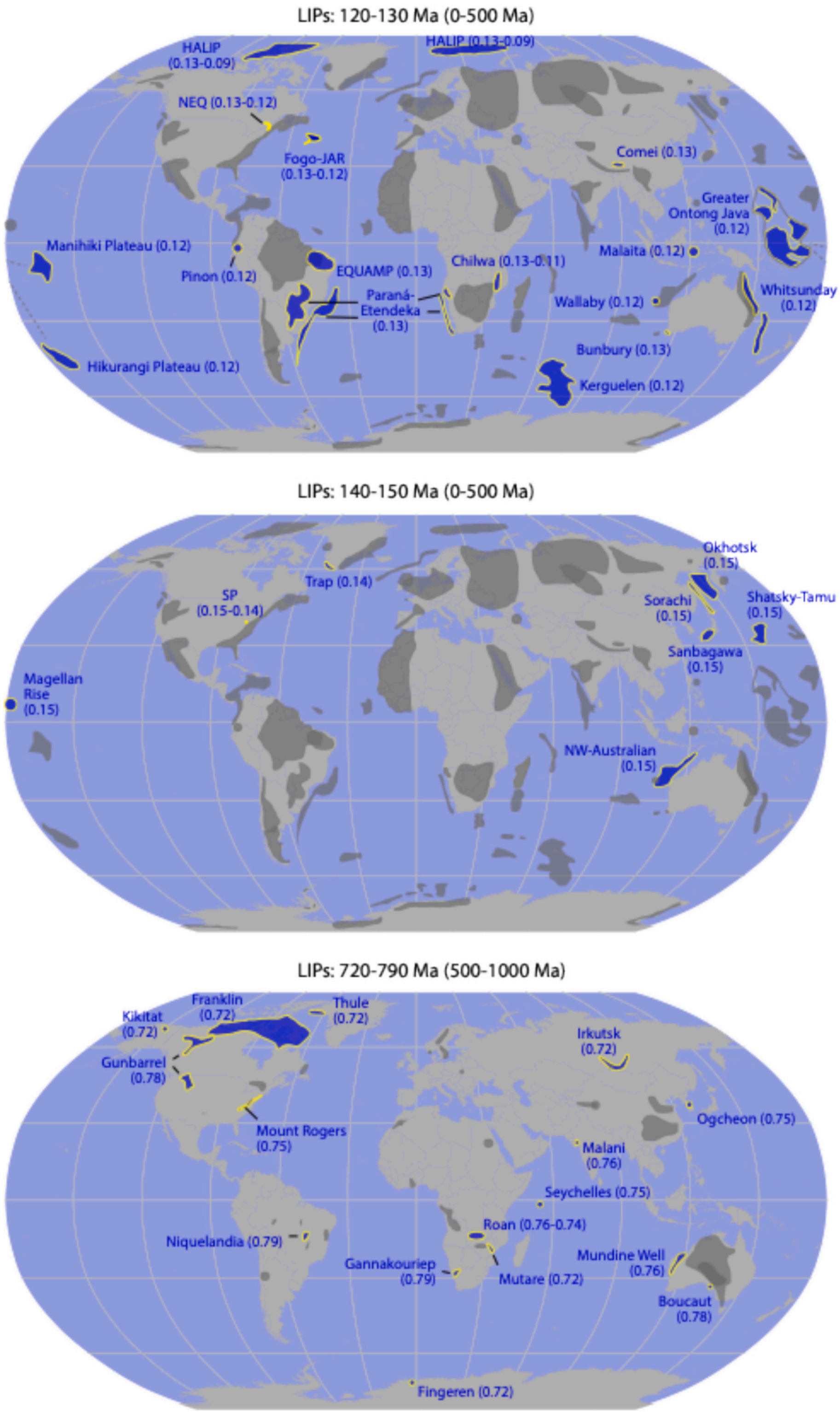

Fig. 4. Present-day world map showing three time intervals with frequent LIP events (blue with yellow borders), modified from Fig. 12.2 of Ernst et al. (2020). Upper, LIPs aged 120–130 Ma; middle, LIPs aged 140–150 Ma; lower, LIPs aged 720–790 Ma. SP, Shenandoah Province. New England-Quebec (NEQ) and Fogo-JAR-Anomaly Ridge igneous provinces are added. HALIP, High Arctic LIP; EQUAMP, Equatorial Atlantic Magmatic Province. Numbers in brackets indicate LIP ages in Ga. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Although SP and Bermuda (BV) edifice construction appear to be coeval (see review in Vogt and Jung, 2007), later authors attributed them both to a Bermuda plume even though they are some 1400 km apart (Morgan, 1983; Duncan, 1984). The calculated trace of this postulated plume approximately coincides in predicted age and location with ca. 65–100 Ma igneous activity associated with the Mississippi Embayment (Cox and Van Arlsdale, 2002). Cox and Van Arlsdale (1997) proposed that contemporary seismicity there is a tectonic relic of the Cretaceous location of the Bermuda hotspot.

The SP (Fig. 1B) is a compositionally diverse swarm of volcanic necks, dikes, and plugs, mostly within a ca. 200 km^2 area centered near Monterey, Virginia. If the easternmost intrusion is included (Mole Hill, Fig. 1B; Johnson et al., 2013) the SP extends east-west for 70 km. The SP is thus much smaller than the NEQ in areal extent.

SP intrusives were emplaced at ca. 48–42 Ma (Fullagar and Bottino, 1969; Mazza et al., 2014). North Carolina bentonites (dated at 45.7 and 46.2 Ma) were likely derived from SP volcanics (Harris and Thayer, 2020). Because most of the intrusives are reverse magnetized, they were plausibly emplaced during a single reverse chron (Lovlie and Opdyke, 1974). The 2020 Geomagnetic Time Scale (Speijer et al., 2020) would suggest a date from a later "C" chron, perhaps C21r (48.878–47.763 Ma). Intrusions during the relatively long C20r would match the bentonite ages better but conflict with the dates of Mazza et al. (2014). Nevertheless, intrusion evidently occurred during a brief period of time–less than ca. 1 Myr.

The SP and NEQ were never simultaneously active. However, they resemble one another in several ways. Both suites are anorogenic and mostly felsic or alkalic. Both were intruded into existing older (Appalachian and/or Pre-Cambrian) continental crust ca. 500 km west of the continent-oceanic crustal transition. NEQ magmatism was structurally controlled, at least in part, by the ancient Ottawa graben, and the EW trending SP by the postulated Thirty-Eighth Parallel Fracture Zone (Fig. 1B) (Dennison and Johnson Jr., 1971).

The SP intrusions were evidently not structurally controlled by CAMP- or earlier Late Triassic rifts, possibly because of a ca. 90° difference in the orientation of the axis of greatest tension (or least compression). This was NW-SE for the older features and NE-SW (Southworth et al., 1993) for the SP. As for the NEQ, recent earthquakes do not directly correlate with the area of SP magmatism, which is relatively aseismic. The Central Virginia Seismic Zone, an area of ca. 4000 km^2, is centered about 100–150 km ESE of the SP.

Both SP and NEQ magmatism were preceded by earlier magmatism in the same areas. Early NEQ magmas are dated at 190–160 Ma (McHone, 1996) and ages of 180–207.5 Ma are reported for the WM batholith (Kinney et al., 2021). Older SP intrusives (alkaline syenites) are dated to ca. 148 Ma (Zartman et al., 1967; Southworth et al., 1993). Hotspot/plume models fail to explain such long-term repeated magmatism.

4.1. Bermuda volcanoes and the Bermuda rise

The magmatic event that produced the BV has not been satisfactorily dated. This short volcano line is oriented NE-SW, so the intrusions followed the structural grain of the oceanic crust (Vogt and Jung, 2007) which accreted early during the 121–83 Ma superchron (Ogg, 2020). Because this crust formed about the same time as the J-Anomaly Ridge (Section 3.3) anomalous mid-Atlantic Ridge axial processes may have produced crust and mantle lithosphere more prone to subsequent failure, i.e. an oceanic weakspot. Drill cores through the Bermuda carbonate platform recovered post-edifice sills intruded into altered pillow lavas dated at ca. 33 Ma (Reynolds and Aumento, 1974). However, stratigraphic dating of volcanogenic and shallow water shell debris near the base of the main edifice (DSDP Site 386; Tucholke and Vogt, 1979) indicates that Bermuda volcano had reached sea level by Early to Mid-Eocene.

Oher DSDP Leg 43 cores suggest the Bermuda Rise formed at about the same time. The swell is elongated NE-SW, perpendicular to what the hotspot/plume model would predict (Vogt, 1991; Vogt and Jung, 2007). The Rise has not subsided, so does not have a thermal origin. Vogt (1991) suggested it might be supported by ongoing mantle convection associated with the continent-ocean boundary. However, it could also be supported compositionally by lower density material.

The implied ca. 47–40 Ma minimum stratigraphic age of the main BV and associated rise is within error coeval with the SP. This is inconsistent with both being produced by the same fixed hotpot (Vogt, 1991). The morphologic difference between the present BV and the SP is geologically superficial. The terrestrial SP was eroded down to its intrusive roots, while the submarine BV only suffered truncation by erosion of the subaerial extrusive parts, with reef carbonates then obscuring the summits.

4.2. Possible correlations of SP magmatism with magmatic and tectonic events elsewhere

The age for the older SP intrusives (148 Ma) is, within error, coeval with formation of the Shatsky Rise in the central Pacific Ocean (Fig. 5). The largest and oldest massif (TAMU) on this rise was cored at two IODP sites, with the older lavas dating from 147 to 143 Ma (Mahoney et al., 2005; Geldmacher et al., 2014). Shatsky Rise does, however, become progressively younger and less massive towards the northeast. Other ca. 140–150 Ma LIPs (Fig. 4; Ernst et al., 2020) include the Magellan Rise and the Trap, Okhotsk, and Sorachi LIPs.

There is no recognized LIP or major rift-drift breakup coeval with the BV-SP episode (e.g., Ernst et al., 2020). However, the major change in Pacific plate motion (relative to hotpots) recorded by the Hawaii-Emperor Bend (HEB) is, within the errors and uncertainties, coeval with the SP episode. Garcia et al. (2015) suggest the HEB initiated at 49–48 Ma, while O'Connor et al. (2013) argue for 50 Ma for the start of a more gradual change completed at ca. 47.5 Ma. More recently Jicha et al. (2018) calculated a HEB age of 49.4 $\pm$ 0.4 Ma but did not estimate the duration. K—Ar ages for the oldest seamounts formed after the HEB are 43.4 Ma (Yurijyaku Seamount) and 42.4 Ma (Daikakuji Seamount). Thus, the data permit a fairly wide range of possible dates and durations for the HEB.

The ca. 47.0–47.9 Ma ages for SP magmas (Mazza et al., 2014) thus match the age of the HEB. A dramatic change in motion of the Pacific plate must have affected other plates and thus perturbed intraplate stress. This may have influenced magmatism at the BV-SP for the time it took for plate motions to re-equilibrate. These time periods may be comparable in duration to the typical 1–2 Myr emplacement times of many LIPs (Coffin and Eldholm, 1994; Ernst et al., 2020). This needs study by additional radiometric dating and numerical modeling.

5. Possible correlations with geomagnetic reversal frequency

The possibility that geological processes such as magmatism, and more generally plate tectonics, might correlate with geomagnetic reversal frequency has been discussed and modeled in many papers (e.g., Vogt, 1975; Courtillot and Besse, 1987; Loper et al., 1988; Larsen and Olson, 1991). It has been proposed that changing plume flux from the core-mantle boundary could change the temperature gradient in the outer core and thus influence reversal frequency (e.g., Larsen and Olson, 1991). This would predict a correlation between surface magmatism and reversal frequency. A full review of the many relevant studies (e.g., Gubbins, 2008; Petrelis et al., 2011; Amit and Olson, 2015) is beyond the scope of this paper.

The most recent reversal time scale (Ogg, 2020) places the onset of superchron C34n at 120.964 Ma. Other proposals include 121.2–123.4 Ma (Olierook et al., 2019) and 119.4 $\pm$ 0.12 Ma (Li et al., 2023). These differing proposals reflect uncertainties in establishing the age of the Aptian-Barremian boundary. Within the uncertainties the beginning of superchron C34n thus closely, if not precisely, appears to coincide with the Greater Ontong Java magmatic event (Section 3.7).

There is a modest conflict (ca. 1–2 Myr) between the 120.964 Ma (Ogg, 2020) onset of superchron C34n and the majority of OJP radiometric ages (Fig. 3). However, most of this conflict disappears if the C34n onset dates are taken from early in the uncertainty range (123.8–121.8 Ma; Fig. 3 of Olierook et al., 2019). These point to an Aptian-Barremian boundary age such that superchron C34n may have started during the most generally accepted peak eruptions of the OJP. Thus, the data suggest that this exceptional magmatic event correlates with an exceptional geomagnetic event.

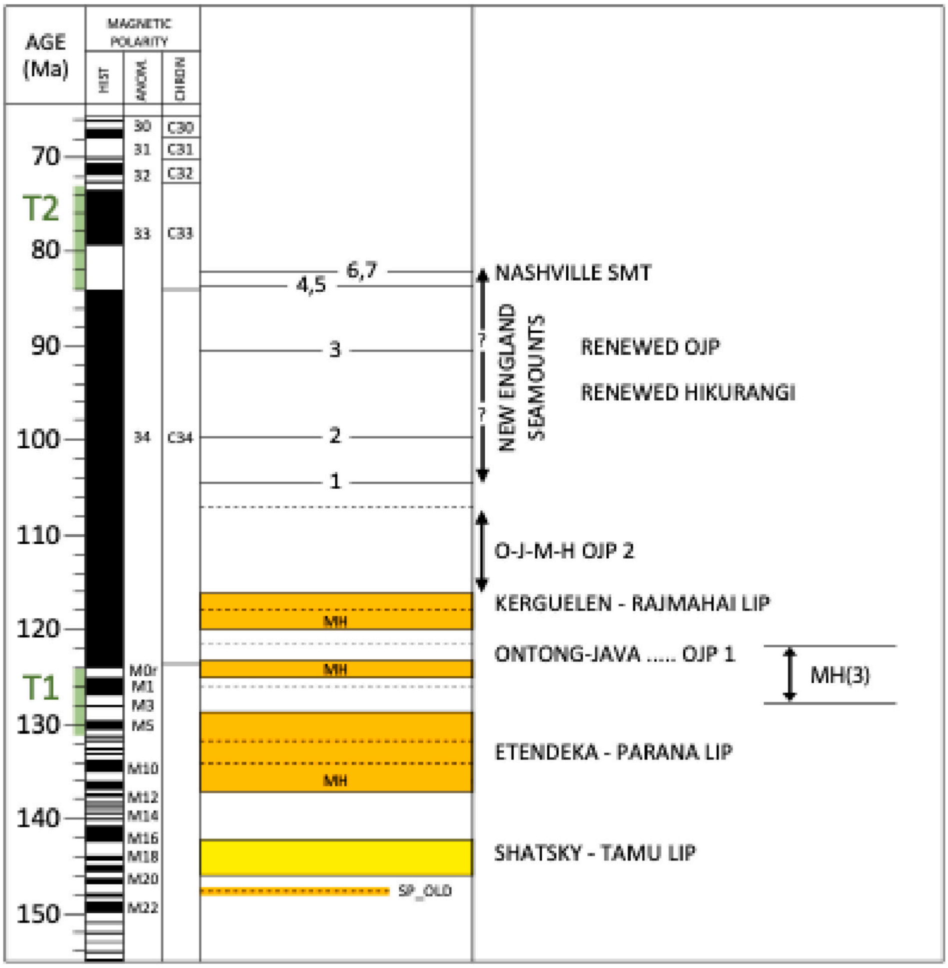

Fig. 5. Cretaceous magmatic episodes relative to the 2020 Geomagnetic Reversal Time Scale (Ogg, 2020). ONTONG-JAVA OJP 1, age of OJP according to pre-2023 publications; O-J-M-H OJP 2, range of ages according to newer Ar—Ar dating (Davidson et al., 2023); MH(3), MH age range according to Boemmels et al. (2021); SP_OLD, older, 148-Ma Shenandoah Province intrusives. 1–7 denote New England Seamount ages (Merle et al., 2019; Fig. 1B). T1 and T2 (shown in green), intervals of time with intermediate low reversal frequency suggesting transitional change from frequent reversals to the C34n superchron. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

As regards correlations also with MH magmatism, differences of a few Myr are permitted by the existing data. Four normal and reversed chrons are detected in the MH plutons and it is clear that some are reversely magnetized (Foster and Symons, 1979; de Boer et al., 1988). This implies that at least some and possibly all MH plutons predate C34n. All OJP lavas are normally magnetized and probably erupted during superchron C34n (e.g., Riisager et al., 2003). The most recent Ar—Ar dates of the OJP range from 108 to 116 Ma (Davidson et al., 2023), implying a 5–10 Myr lag between C34n onset and OJP magmatism.

5.1. Superchron onset and termination: Gradual or abrupt?

Did geomagnetic reversal frequency change abruptly at the onset of C34n, or is there an observable precursor (Hulot and Gallet, 2003)? The reversal chronology of Ogg (2020) and earlier ones suggest that the last six M-chrons lasted about three times longer on average compared to dozens of preceding ones (Fig. 5). Those six chrons, M5n to M0r, lasted 1.012, 1.797, 0.548, 0.386, 2.383, and 0.436 Ma respectively with a mean of 1.03 Myr (Ogg, 2020). By contrast, the previous 6.037 Myr experienced 19 chrons with a mean duration of 0.32 Myr. The longest and second longest lasted only 0.771 and 0.442 Myr. There may thus have been two reversal frequency changes, dated on the Ogg (2020) calibration at ca. 127.526 Ma and 120.64 Ma (Ogg, 2020) or 119.7 Ma (Li et al., 2023). Such a stepwise decline in reversal frequency can be considered a crude approximation to the nonstationary decline deduced by McFadden and Merrill (2000). Vogt (1975) suggested that the apparent 6-Myr precursor to C34n represented the lag in time for a mantle plume event near the CMB to rise to the surface.

A similar, also possibly stepwise, return to higher frequency occurs at the end of C34n with a polarity switch at 82.875 Ma (Ogg, 2020). This reversed chron was followed by a second long normal polarity chron (79.90–74.201 Ma). These two chrons, together lasting over 8.6 Myr, were followed by nine chrons over the following ca. 8.6 Myr. No chron pair lasting as long as 8.6 Myr has occurred subsequently. Cenozoic reversal rates have averaged about two or three per Myr. These observations suggest C34n was both preceded and followed by intervals of anomalously low reversal frequency ca. 5–10 Myr in duration. C34n termination coincided with with a sharp decline in Atlantic spreading rates and global oceanic crust production rate (e.g., Larsen, 1991a, 1991b).

No magmatic events dating to around the 82 Ma C34n termination have been found along the eastern continental margin of North America. The eastern New England Seamounts (Duncan, 1984; Merle et al., 2019), however, might all have formed coevally at ca. 82–84 as discussed above and thus relate to superchron termination.

5.2. Summary

Considering 1) the geomagnetic reversal time scale, 2) rock radiometric geochronology, 3) rock magnetic polarities and 4) the global plate kinematic history (not reevaluated here but see Larsen, 1991a, 1991b and Larsen and Olson, 1991), we conclude the following:

A) The C34n superchron was preceded (T1 of Fig. 5) and followed (T2) by ca. 5–10 Myr of transitional low but non-zero reversal frequencies.
B) The early Cretaceous NEQ intrusives and the JAR-Southeast Newfoundland Ridge magmatism largely correspond to period T1. There is no evidence for major global magmatism associated with T2.
C) The OJP super-LIP eruptions postdate C34n onset by ca 5–10 Myr based on most accurate dating (Davidson et al., 2023).
D) Global oceanic crust production by sea-floor spreading increased around C34n onset time (Larsen, 1991a, 1991b).
E) The 5–10 Myr lag between C34n onset and OJP super-LIP emplacement could be interpreted as (but does not require) a corresponding plume rise time from the core-mantle boundary as speculated by Vogt (1975). Model rise time estimates vary widely, from 5 to 100 s of Myr (e.g., Ito et al., 2003).

5.3. The Mt. Rogers Igneous Province–a Neo-Proterozoic analog to the New England-Quebec Igneous Province?

If the weakspot model is valid, other examples should be found on earlier passive margins. The Mt. Rogers Formation (Fig. 1A), also on the eastern margin of Laurentia, preserves a record of anorogenic magmatic events within continental crust just west of the incipient Rodinia breakup rift (Fig. 1 of Macdonald et al., 2023). This 900-km SW-NE line of ca. 40 magmatic centers (Table 1 of Macdonald et al., 2023) is aged 800–550 Ma (Tollo and Aleinikoff, 1996; Tollo et al., 2012; McClellan and Gazel, 2014). However, about 75 % of those ages span only 760–700 Ma. Ernst et al. (2020) assign an average age of 750 Ma to the Formation of which the ca. 100-km-long Robertson River Igneous Suite (Fig. 1A,B), a lineament of nine granitoid plutons, provides the closest analog to the WM and MH intrusives.

Both the MH-NEQ and Mt. Rogers igneous provinces are anorogenic and intraplate. Both occur in continental crust inland from rifted passive margins. Neither is directly associated with breakup. The WM-MH province is approximately orthogonal to the present continental margin whereas the Mt. Rogers pluton line roughly parallels the margin of Laurentia. The Mt. Rogers province could be the expression of a failed Rodinia (Laurentia) rift. Magmatism in both cases extended for many tens of Myr. WM magmatism occurred in two distinct episodes at ca. 200–165 Ma and 130–110 Ma (Eby et al., 1992). Many ages for the former episode are based on the older K—Ar method and more accurate dating is required.

Robertson River Igneous Suite U—Pb zircon ages (Fig. 6; McClellan and Gazel, 2014) cover most of the Mt. Rogers province ages and illustrate the scatter and errors. The granite and felsite pluton ages range from 745 to 702 Ma with an average of 722 Ma. Errors are ±2–9 Myr with an average of 4 Ma. The errors make it difficult to identify individual abundance peaks such as those suggested in Fig. 3. As far as can be said, the plutons thus lack an age trend and feature a magmatism age peak around 720–717 Ma.

About 6 % of detrital zircon from fluvial deposits of the Pliocene Potomac River in southern Maryland have been dated with a peak probability density of 723 Ma (McCormick et al., 2017). These zircons were most likely eroded from the northern Robertson River plutons and arrived in southern Maryland by way of the Shenandoah River, a major Potomac tributary. The detrital zircon ages probably only represent

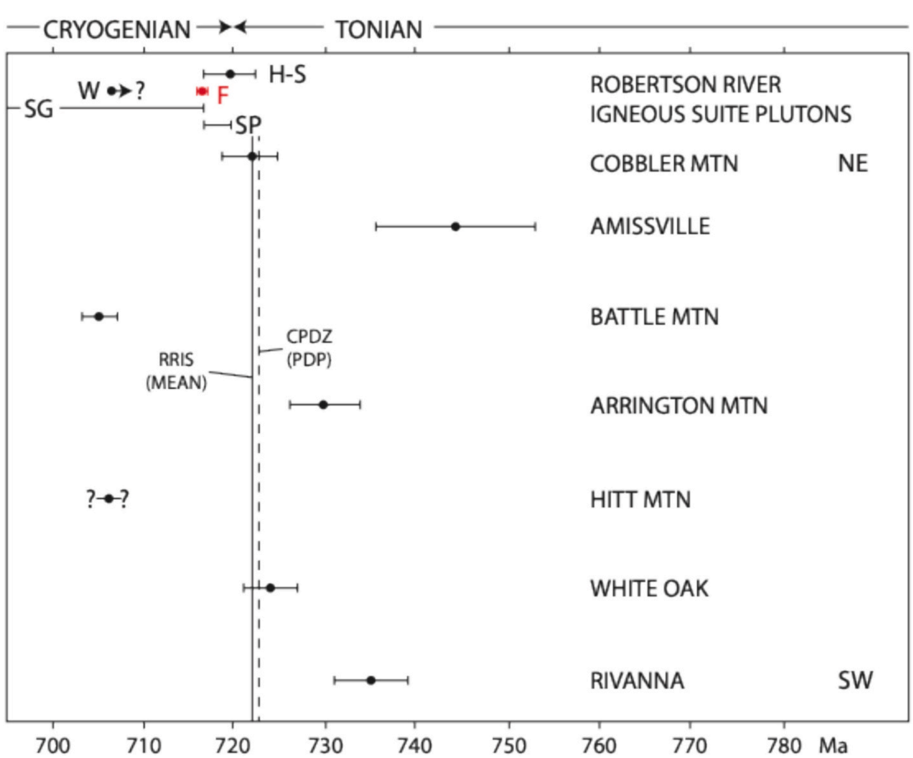

Fig. 6. Radiometric ages for the Robertson River Igneous Suite plutons arranged from SW (lower end of vertical axis) to NE (upper end of vertical axis) (McClellan and Gazel, 2014; Southworth et al., 2009). CPDZ, Coastal Plain Detrital Zircon age probability density distribution peak (PDP) (McCormick et al., 2017). RRIS (MEAN), Robertson River Igneous Suite pluton mean age. Distant magmatism of similar ages include F, Franklin LIP (in red); SG, onset of Sturtian glaciation (Macdonald and Wordsworth, 2017); W, Wichita Mtns volcanism (Hanson et al., 2016); H—S, Hubai-Shankxi LIP (Lu et al., 2022). SP is 'superplume' postulated by Lu et al. (2022). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

some of the Robertson River plutons and not the entire Mt. Rogers Formation. This distinction illustrates the challenges of detrital zircon age interpretations.

The best average age currently available for the Robertson River plutons, 720–717 Ma, is within error synchronous with the emplacement ages of the distant Franklin, Thule, Fingeren, Mutare and Irkutsk LIPs (Fig. 4; Ernst et al., 2020). Ernst et al. (2020) identify about 30 LIPs emplaced during the period 1000–500 Ma, and 790–720 Ma LIPs are scattered on all continents except South America. LIPs dated at 720–790 Ma are thus about twice as common as expected for that 500 Myr interval. Whether detailed analyses and additional data will enable a higher resolution search for shorter intervals of globally or regionally correlated magmatism remains to be determined.

The start of the Sturtian global 'ice house' age (Cryogenian) is also dated at 717 Ma and is not statistically different from the average age of the Robertson Rivers plutons (Macdonald et al., 2023). Macdonald and Wordsworth (2017) attribute a causative role to the voluminous equatorial Franklin volcanism triggering the Sturtian glaciation, the onset of which they date at 717.4 ± 0.1 Ma and 716.5 ± 0.5 Ma. Nevertheless, cold global climates had already begun around the time of emplacement of the southern Mt. Rogers Formation southwest of the Robertson River plutons (ca. 755 Ma). The correlation may thus be coincidental (MacLennan et al., 2020).

We suggest that the Mt. Rogers line of plutons comprises an intracontinental igneous province similar to the WM and MH plutons in composition, scale and relationship to breakup. All three pluton groups were emplaced in lines (Fig. 1B) and have lengths of ca. 100 km (Robertson River Igneous Suite, McClellan and Gazel, 2014), 160 km (WM, Fig. 2), and 140 or 250 km (MH). The Robertson River lineament parallels the inferred breakup rifts (Macdonald et al., 2023). Plutons in lines suggest structural control by faults.

The Mt. Rogers pluton line may thus be part of a global magmatic event related to the breakup of Rodinia (Ernst et al., 2023). If so, it is analogous to the WM-MH magmatism which appears to have a similar relationship to the following Wilson Cycle. No oceanic crust from this period survives so oceanic LIPs can only have been preserved where fragments were obducted onto, or emplaced into, continental crust.

We consider the older WM magmatism (207 Ma, Kinney et al., 2022) to provide evidence for a persistent NEQ weakspot. The Robertson River plutons are located in the same area as the proposed SP weakspot. Thus, this weakspot could have existed already in the late Proterozoic. If so, some weakspots might have persisted in some form throughout all the Appalachian orogenies.

6. Synthesis

A mantle plume model fails to explain NEQ and SP magmatism for the following reasons.

1) For both the NEQ and SP much older post-breakup intrusives underlie younger ones. SP Eocene eruptions occurred where early Cretaceous ones had been emplaced ca. 100 Myr earlier. WM early Cretaceous intrusions were preceded by others 75 Myr earlier. Plume models would have to require the plate to cross multiple hotspots at the same site.
2) Both the SP and NEQ areas are underlain by anomalously low seismic wave speed mantle. Mantle plume models would have to infer a recent plume rising below the old magmatic province or that the anomalous mantle is ca. 125 Myr old and has been transported 4100 km as part of the plate. On the contrary, plates are thought to move relative to the asthenosphere.
3) Radiometric dates of both igneous provinces correlate with near- and distant anomalous magmatism and tectonism including LIPs, continental breakup, and a major change in motion of the Pacific plate. Such correlations are not explained by plume models.
4) Intraplate seismicity occurs near the NEQ and SP. Plume models do not explain why this should happen 48 Myr (SP) and ca. 125 Myr (NEQ) after the plate passed over the putative hotspot.

In order to explain these observations with a mantle plume model, a situation would thus be required whereby more than one plume impacted the same region coincidentally with major volcano-tectonic events elsewhere, possibly crossed one another, emplaced low seismic wave speed mantle that was carried with the plate, and cause earthquake activity up to the present day several thousand kilometers away from the present plume locality. Such a model is contrived and implausible. No other postulated plume locality worldwide is suggested to exhibit this combination of features.

A weakspot model involving old structures that are easily reactivated by stress changes explains the observations better. Indeed, the repeated failure of persistent weak zones in the lithosphere is required by the Wilson cycle concept (Schiffer et al., 2020). Propagating rifting, accompanied by propagating volcanism, is a natural and predicted part of that model. Continental breakup on a scale of thousands of kilometers does not happen everywhere simultaneously but progressively forms new rifts. Underlying mantle seismic anomalies are also expected since structure may contrast with stable surroundings due to damage and fluid content (Foulger et al., 2013).

Low-level failure in weak zones under the present ridge-push compression can account for current ongoing seismic activity. Enhanced intraplate seismicity generally coincides with relatively thin lithosphere. For example, seismically active Britain has a lithosphere thickness of 83 km and is thought to be warmer, and with reduced mechanical strength. This may be compared with Ireland, which is seismically inactive and has a lithosphere thickness of around 100 km (Lebedev et al., 2023). Globally, Archean and early Proterozoic cratons are associated with thicker, colder lithosphere, higher shear wave speeds at 175 km depth (Mooney et al., 2012) and lower intraplate seismicity. Seismicity in the neighborhoods of the SP and NEQ is consistent with this picture.

The sites of NEQ and SP magmatism appear to have been particularly structurally vulnerable. More research is needed, however, particularly of paleostress, to explain why this should be. Structural features, mainly reactivated Paleozoic or even Pre-Cambrian faults, have been suggested (e.g., Faure et al., 1996) but the entire margin is underlain by networks of faults and other structures. Why the NEQ and SP areas are different remains to be clarified. The 750–720 Ma Mt. Rogers Formation and the Robertson River Igneous Suite magmatism provides an older, Proterozoic Laurentian margin analog to NEQ and SP magmatic behavior.

We infer an overall global pattern of relatively stable plate motions periodically interrupted by shorter LIP/continent-breakup intervals and major plate motion reconfigurations, rather than synchronized plume activity (Vogt, 1972). The characteristic ca. 25 Myr periods of stable motion may in part reflect the frequency with which large blocks of continental crust raft into one other. The shorter intervals of instability may last only ca. 1–5 Myr and account for less than 10 % of geologic time. One short interval of change is recorded by the HEB. Both relaxation of compressional stress and tension may be influential in bringing about these periods of instability (Nielsen et al., 2007). At these times, significant changes in stress would affect the plate interiors and passive margins. These events would be recorded at weakspots by changes in magmatism. Asthenosphere flow adjusts in accordance with changes in plate motions but is not the driving force. This model could be explored using numerical modeling that includes intraplate stress as an output and tested by field structural mapping.

Testing the possibility that anorogenic continental margin magmatism reflects major global tectono-magmatic events or major changes in plate motion requires additional accurate geochronological data. Some apparent correlations might be coincidences and older radiometric ages, especially K—Ar whole rock data, can have large errors (e.g., Garcia et al., 2015).

Oceanic lithosphere is recycled at subduction zones on timescales of 50–200 Myr. A global record of magmatism for nearly all Earth history is thus found only in the continental record. Establishing temporal connections, if any, between post-200 Ma continental intraplate igneous provinces (e.g., NEQ), oceanic LIPs (e.g., OJP) and magnetic reversal frequency changes (e.g., the onset of superchron C34n) offers the possibility of using continental anorogenic intraplate magmatism and/or reversal frequency as proxies for global magmatism and plate tectonic behavior. If confirmed, ancient continental anorogenic magmatic events might help to constrain ancient geomagnetic reversal frequencies and geomagnetic dynamo behavior. If verified, such correlations could provide clues regarding pre-Jurassic plate motions and oceanic LIPs for which essentially no oceanic crust remains.

Declaration of competing interest

Coauthor was on Editorial Board of this journal until December 2023. Neither author has known competing financial interests or personal relationships that influenced the work reported in this paper.

Data availability

Data will be made available on request.

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