Arc and slab-failure magmatism of the Taconic Orogeny, western New England, USA

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Abstract: During the 1970s, geologists considered that the Upper Ordovician Taconic Orogeny represented the collision of Laurentia with the Ammonoosuc arc, now largely exposed on the Bronson Hill anticlinorium. Subsequently, several researchers noted that magmatic rocks which intrude and overlie the Ammonoosuc arc are younger than the *c.* 455–451 Ma Taconic Orogeny. This led them to hypothesize that a Middle Ordovician collision was followed by westward-dipping subduction beneath the amalgamated Laurentian–Ammonoosuc zone to produce the younger arc rocks. In this model, the Taconic allochthons and foredeep were produced later in a retro-arc setting above westward-dipping subduction. However, those models prove inadequate due to the lack of ash beds, foredeep sedimentation and deformation on the Laurentian platform prior to the Upper Ordovician Taconic Orogeny.

Here, we resolve the dilemma by recognizing that the magmatic rocks, which post-date the 455–451 Ma Taconic Orogeny, are not arc rocks but, instead, typical post-collisional slab-failure rocks as old as 450 Ma, with $Sr/Y > 10$, $Sm/Yb > 2.5$, $Nb/Y > 0.4$ and $La/Yb > 10$. Thus, in New England and western New York, the Upper Ordovician Taconic Orogeny represents the collision of the Ammonoosuc arc with Laurentia followed by slab failure of the descending plate.

Soon after the plate tectonic revolution and the recognition by geologists that Kay's (1951) miogeoclinal–eugeosynclinal couplet represented the collision of a continental margin with an arc, insightful researchers (Stevens 1970; Chapple 1973; Rowley and Kidd 1981; Jacobi 1981; Stanley and Ratcliffe 1985) developed a model for the Upper Ordovician Taconic Orogeny in which the leading edge of the Laurentian passive margin was pulled beneath an arc complex, generally recognized to be located today on the Bronson Hill anticlinorium of western New England (Fig. 1). The Bronson Hill anticlinorium is a structure that formed during the Middle–Late Devonian Acadian deformation, and contains what is known as the Oliverian suite of Upper Ordovician gneissic plutons, mantled by Middle Ordovician rocks of the Ammonoosuc volcanics (Leo 1991; Moench and Aleinikoff 2003).

Tucker and Robinson (1990) argued that plutons of the 454–442 Ma Oliverian suite, as well as some volcanic rocks, were mostly too young to be arc rocks related to pre-Taconic arc magmatism. Subsequently, Karabinos *et al.* (1998, 2003) suggested that a different package of rocks (485–470 Ma), located west of the Connecticut Valley–Gaspé Trough (Fig. 1) and known as the Shelburne Falls arc, was more likely to be the arc that formed the upper plate during the Taconic Orogeny. Based on this inference, they proposed a more complicated model whereby the Shelburne Falls arc collided with Laurentia between 475 and 470 Ma above an easterly (current coordinates) subduction zone, followed by an outboard step and flip in subduction polarity, such that subduction dipped westwards beneath the amalgamated collision zone to generate the younger arc magmatism of the Bronson Hill anticlinorium. Additionally, Dorais *et al.* (2008, 2012) noted that rocks of the Ammonoosuc volcanics on the Bronson Hill anticlinorium contained isotopic signatures more compatible with Gondwanan lithosphere, whereas plutons of the Oliverian suite had isotopic signatures of Laurentian lithosphere, and so agreed that the Oliverian suite resulted from westward subduction beneath Laurentia after a 475–470 Ma collision. The flipped east-to-west subduction model was expanded and flushed out more recently in a set of companion papers by Karabinos *et al.* (2017) and Macdonald *et al.* (2017), as well as a contribution by Jacobi and Mitchell (2018).

Thus, there seems to be general consensus – based largely on the timing of magmatism and its isotopic compositions – that the Lower–Middle Ordovician Ammonoosuc arc collided with Laurentia over an eastward-dipping subduction zone at about 470 Ma and that younger Upper Ordovician magmatism of the Oliverian suite formed from westward subduction beneath the amalgamated Laurentian–Ammonoosuc

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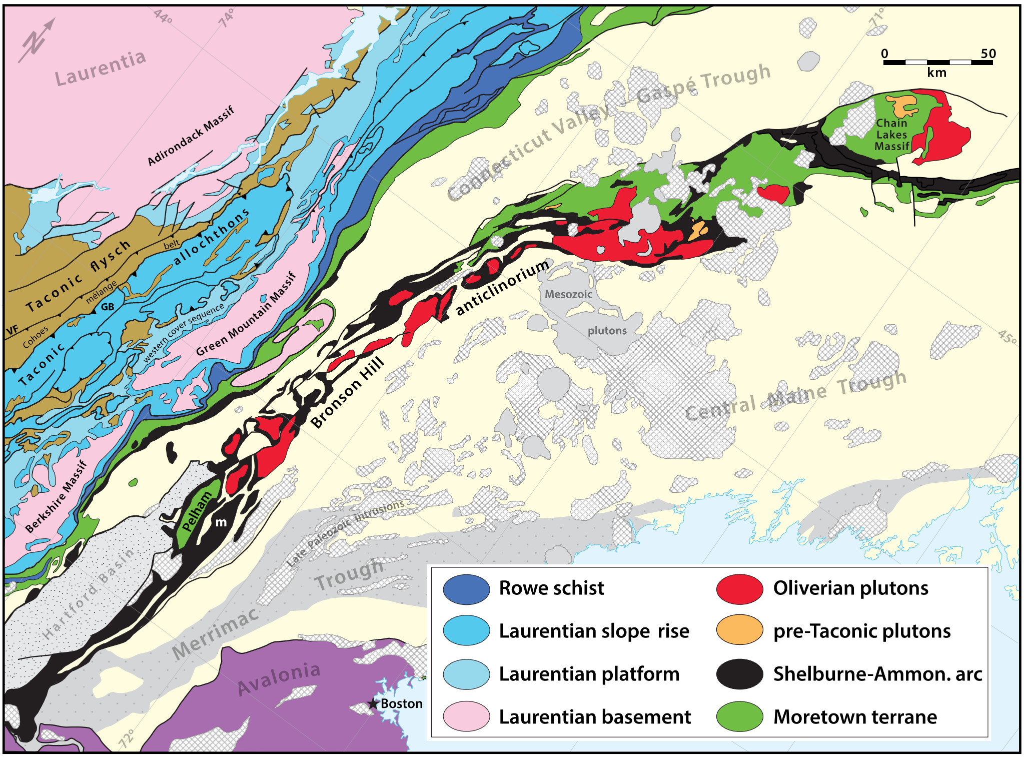

Fig. 1. Regional map illustrating the large-scale geological features of New England and northeastern New York from Hibbard *et al.* (2006) with modifications from Karabinos *et al.* (2017). Shelburne-Ammon. arc, Shelburne Falls–Ammonoosuc arc; VF, Vischer Ferry Fault; GB, Giddings Brook slice.

region of eastern Laurentia. In this model, geologists are forced to infer that the Upper Ordovician Taconic foreland basin and thrusts are retro-arc in nature (Karabinos *et al.* 1998, 2017; Macdonald *et al.* 2014, 2017; Jacobi and Mitchell 2018). These ideas gained acceptance despite the recognition that there is 'no record of Early Ordovician deformation, foreland basin deposits, or air-fall tephras on the Laurentian carbonate platform to the west' (Karabinos *et al.* 2017, p. 537).

Over the years we have examined the geochemistry from a spectrum of magmatic rocks in the North American Cordillera and discovered that those erupted and intruded in pre-collisional arcs had compositions distinct from those formed after collision by the process of slab failure (Hildebrand and Whalen 2014, 2017; Hildebrand *et al.* 2018). On the basis of these and much younger rocks, we developed a group of discrimination diagrams that separate the two suites (Whalen and Hildebrand 2019). For this study, we compiled and plotted geochemical data from Ordovician magmatic rocks of New England and found rocks with arc and slab-failure signatures.

Here we exploit the differences to show that there is no conflict between the relative ages of the Taconic Orogeny and magmatism of the Bronson Hill anticlinorium because the post-Taconic plutons were not generated by arc magmatism but, instead, represent post-collisional slab-failure magmatism, a model initially suggested by Hollocher *et al.* (2002). If we are correct, then there is no need for a hypothesis involving both easterly and westerly subduction to produce the Taconic Orogeny. Instead, and as originally understood, the Taconic Orogeny represents a collision between Laurentia and upper-plate rocks of the Bronson Hill–Ammonoosuc arc during the Upper Ordovician.

Laurentian autochthonous platform margin

Following Ediacaran–Cambrian rifting that led to the opening of the Iapetus Ocean, an easterly-facing passive margin formed on the cratonic margin of Laurentia as it thermally subsided (Cawood *et al.* 2001; Hatcher 2010; Landing 2012). Rocks of the margin occur today in both autochthonous and allochthonous successions. The autochthonous rocks lie unconformably upon Grenville-age igneous and metamorphic basement; however, to the east, rocks were folded, detached from their basement and thrust westwards during the Taconic Orogeny (Rowley and Kidd 1981). Rift-related rocks are generally absent from autochthonous sections and are scarce within the deformed allochthonous successions. Where preserved, a Middle Cambrian sandstone and conglomeratic formation – the Potsdam and its lateral correlatives – sits on crystalline basement but a slightly older succession of metasedimentary rocks occurs to the east within the western Green Mountains where they are known as the ‘western cover sequence’ (Fig. 1).

A series of carbonate platforms, collectively named the Beekmantown Group, sits atop the basal siliciclastic units. The margin was located in the tropics, where it faced south (Scotese and McKerrow 1990) and, according to Landing (2012), subsided slowly enough throughout the Cambro-Ordovician such that changes in eustatic sea level led to several sequences bounded by unconformities. Sandstone units are typically transgressive facies, whereas carbonate platforms represent the highstands.

Laurentian slope in Taconic allochthons

Rocks of the continental slope and rise are preserved in what are termed the Taconic allochthons (Zen 1967, 1972). The allochthons (Figs 1 & 2) are a stacked set of seven thrust slices where rocks of the continental margin were progressively folded, detached from their Laurentian basement and transported westwards over the carbonate platform on a basal detachment, typical of collisional fold-thrust belts worldwide (Boyer and Elliott 1982; Dahlen 1990).

Slope-rise facies rocks are dominantly siliciclastic shales, siltstones, grits, and greywackes with minor carbonate clast conglomerate and breccia. In the Giddings Brook slice (Fig. 1), which is the most extensive slice of the Taconic allochthons, Cambrian formations include the Rensselaer, Nassau, Browns Pond, Granville and Hatch Hill, in ascending order (Landing 2007). The eustatic changes in sea level noted on the platform directly correlate with units on the continental slope as weakly to unburrowed anoxic black shale alternate with more oxygenated burrowed, green shale (Landing 2012).

Along the western side of the allochthons, Rowley and Kidd (1981) mapped a conformable Lower–Middle Ordovician section through the Poultney, Indian River, Mount Merino and Pawlet formations. The Poultney Formation, or Deep Kill Formation as it is sometimes called, contains many 10 m-thick alternations of black and green mudstones that Landing (2007) found to be geographically and temporally persistent for some 1200 km along strike and biostratigraphically correlative to rocks on the adjacent carbonate platform.

The end of passive-margin sedimentation on Laurentia is first seen within the Taconic allochthons where it is marked by the Indian River Formation, a 50 m-thick, condensed sequence of red, strongly bioturbated mudstone that contains radiolarian-rich cherty horizons and volcanic ash, the oldest debris not of North American provenance on the passive margin (Landing 1988, 2012; Landing *et al.* 1992). The marked lack of quartz sand and carbonate sediments characteristic of the older slope units, coupled with the condensed and oxidized nature of the unit, led Landing (2012) to suggest that the unit was deposited as the continental slope rode up and over the peripheral bulge to the subduction zone (Jacobi 1981), such that sediment ponded on either side of the bulge. This idea is supported by neodymium isotopic data from slates of the Giddings Brook slice that show a clear break from consistent ε_{Nd} values of -15 to -12 from Cambrian units to the middle of the Indian River to less evolved values of -12 to -9 upwards through the Mount Merino and Pawlet formations (Macdonald *et al.* 2017).

Macdonald *et al.* (2017) also collected and studied detrital zircons from three locations within the Indian River Formation and found strong peaks dated by laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) of 453 ± 7, 451 ± 9 and 456 ± 7 Ma; however, they also found that individual prismatic zircons, analysed by chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS), yielded ages of 464, 466 and 468 Ma, which they interpreted to be the depositional age of ashes within the Indian River Formation. Given that the beds also contain abundant older rounded detrital grains without younger magmatic overgrowths, they are more reasonably interpreted as epiclastic, and thus the prismatic zircons do not accurately reflect the age of the Indian River Formation but, more likely, reworked ash. Additionally, Landing (1988) reported that specimens of early Caradocian *Nemagraptus gracilis* were collected from rocks of the formation and so are younger than 458 Ma (Fig. 3). Jacobi and Mitchell (2018) also noted the age discrepancy but invoked an unconformity at the top of the Indian River Formation to resolve it, which conflicts with field observations and with faunal evidence.

Rocks of the Indian River Formation are abruptly overlain by green to black pyritic and graptoliferous black shales of the Mount Merino Formation (Rowley and Kidd 1981; Landing 2012). The upper contact of the Mount Merino Formation with the overlying Pawlet Formation (also called Austin

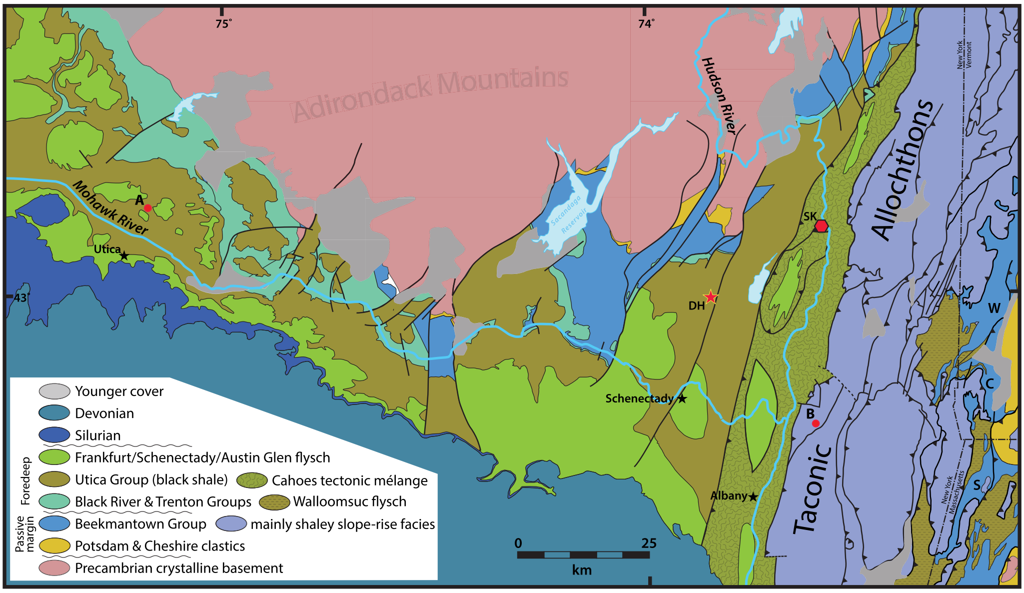

Fig. 2. Map illustrating the generalized geology of the Mohawk Valley and the area of the Taconic allochthon within eastern New York and westernmost New England modified from Fischer *et al.* (1970) with modifications from Bradley and Kusky (1986) and Jacobi and Mitchell (2018). The star marked DH is the location of the drill hole mentioned in the text; Starks Knob, a block of pillow basalt and the only known exotic clast in the Cohoes mélange (Landing *et al.* 2003), is shown as the small red hexagon labelled SK; WCS is part of the western cover sequence; and points AB mark the approximate end points for the cross-section in Figure 3.

Glen Formation) is gradational and is marked by the influx of thin silty to sandy greywacke turbidites, which thicken and coarsen upwards to contain detrital chromite, deformed metasedimentary clasts and devitrified volcanic fragments, which must have been derived from the east as turbidites are thicker, coarser and more abundant in that direction and contemporaneous rocks to the west on the Laurentian carbonate platform were hemipelagic shale grading westwards into limestone (Rowley and Kidd 1981). Both the Mount Merino and Pawlet formations contain graptolites (Fig. 3) characteristic of the *Nemagraptus gracilis–Climacograptus bicornis* and *D. multidens* zones (Berry 1962; Riva 1974). Rowley and Kidd (1981) pointed out that rocks of the Mount Merino and Pawlet formations have the same deformational history, which requires that the slope sequence was undeformed during turbidite deposition and not yet tectonically detached and transported. Based on the fauna, deposition of rocks in the Pawlet Formation ended abruptly at about 455–454 Ma (Fig. 3), when the unit was incorporated in the thrust stack. The turbiditic package then spread rapidly westwards in front of the migrating allochthons to form a blanket over the foundered autochthonous platform and its trench-slope deposits of black shale (Bradley and Kusky 1986; Bradley and Kidd 1991; Jacobi and Mitchell 2018).

Foreland basin

The top of the Beekmantown Group is marked by a major erosional interval, the post-Knox unconformity (Landing 2012), which some workers have suggested marks the passage of the passive margin over the flexural bulge to the trench and loading of the margin by an arc (Jacobi 1981), whereas others (Macdonald *et al.* 2017) attribute the unconformity largely to slab break-off during or immediately after deposition of the Indian River Formation. However, Hatcher and Repetski (2007) noted that the unconformity is far too widespread, recognizable on North America as far south as the Ouachitas, as far west as the Rocky Mountains and even on different cratons, so it must be related to a eustatic drop in sea level, perhaps ultimately due to processes in the upper mantle. Problematic for the break-off hypothesis of Macdonald *et al.* (2014, 2017) is that break-off should lead to rapid uplift and exhumation of the lower plate but rocks of the Indian River Formation are overlain by deep-water facies of the Mount Merino and Pawlet formations, as discussed above.

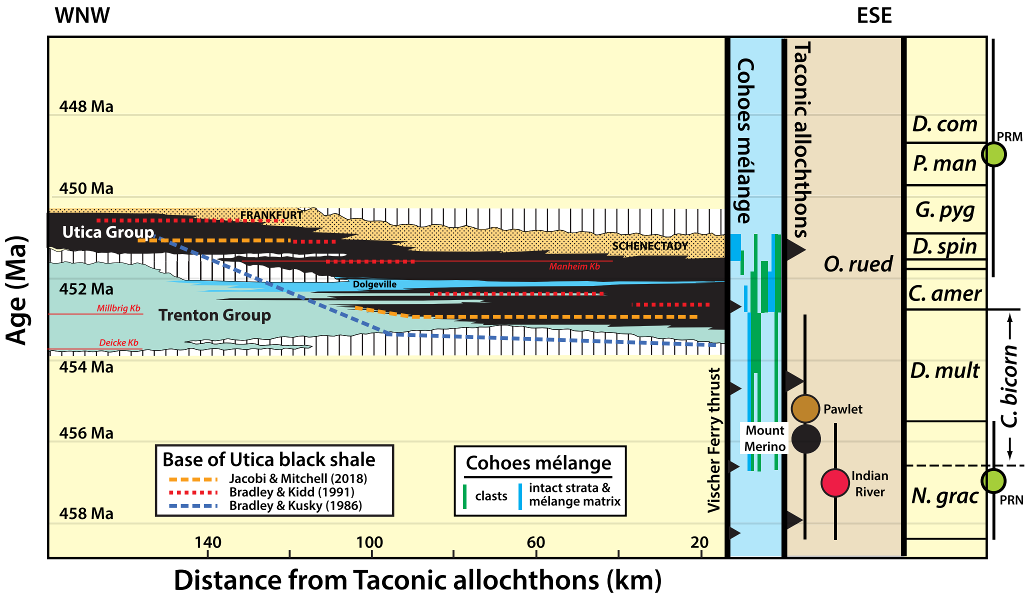

Fig. 3. Synthetic cross-section illustrating the geological and temporal relationships between units of the Taconic foredeep within the Mohawk Valley and selected units of the Laurentian slope rise within the Taconic allochthon. The figure is based on Rowley and Kidd (1981), Bradley and Kidd (1991) and Jacobi and Mitchell (2018). Coloured dashed and dotted lines show various authors' estimated time for the unconformity at the base of of the Utica Shale as recalculated by Jacobi and Mitchell (2018). The chronostratigraphic diagram illustrating stratigraphic correlations within the foredeep is from Brett and Baird (2002) with minor age corrections based on faunal ages cited in Jacobi and Mitchell (2018), as are faunal ages from the Cahoes mélange shown by coloured bars. PRM, U–Pb zircon age with errors of rhyolite within the Partridge Formation of central Massachusetts (Tucker and Robinson 1990); PRN, faunal age of rhyolite within the Partridge Formation of northern New Hampshire (Karabinos *et al.* 2017). Graptolite zones: N. grac, *Nemagraptus gracilis*; D. mult, *Diplograptus multidens*; C. bicorn, *Climacograptus bicornis*; C. amer, *Corynoides americanus*; O. rued, *Orthograptus ruedemanni*; D. spin, *Diplacanthograptus spiniferus*; G. pyg, *Geniculograptus pygmaeus*; P. man, *Paraorthograptus manitoulinensis*; D. com, *Dicellograptus complanatus*. Age of K-bentonites: Manheim (Macdonald *et al.* 2017), Millbrig and Deicke (Sell *et al.* 2013).

Additionally, the presence of the Caradocian graptolite *Nemagraptus gracilis* within the Indian River Formation place it well above the post-Knox unconformity, perhaps by 8–10 myr.

Within the Mohawk Valley area of New York, located just west of the Taconic allochthon, a more likely horizon to place the earliest effect of the Taconic Orogeny, and passage of the platform terrace over the peripheral bulge, is a younger erosional surface that underlies deposition of the marine carbonates of the Black River and Trenton groups, which were deposited during syndepositional normal faulting. Rocks of both formations pass eastwards and upwards into black marine shale of the Utica Group, which is commonly interpreted to have been deposited on the trench slope (Bradley and Kusky 1986; Bradley and Kidd 1991; Brett and Baird 2002; Jacobi and Mitchell 2002). Rocks of the Black River Group are peritidal, shallow shelf carbonates dominated by coral, whereas those of the overlying Trenton Group contain the earliest K-bentonites (Deicke and Millbrig 454–453 Ma, *C. bicornis* Zone) on the carbonate platform (Fig. 3), and are deeper-water mixed carbonate–siliciclastic facies with platform carbonates and associated slope facies deposited on the western flank of the foredeep basin (Sell *et al.* 2013; Cornell 2008). Seven other bentonitic ashes dated by Macdonald *et al.* (2017) from the Trenton Group are tightly constrained to be between 452.6 ± 0.06 and 450.68 ± 0.12 Ma.

Taconic mélange zone

The Taconic allochthons are underlain along their western side (Fig. 1) by a 10 km-wide zone of broad shear zones containing broken formation and mélange, derived largely from synorogenic deep-water clastics (Plesch 1994), and collectively termed the Cohoes mélange by Kidd *et al.* (1995, p. 8). They envisioned that:

> [T]he mélange formed by the progressive deformation of synorogenic flysch as an accretionary thrust wedge advanced over the Ordovician North American continental margin, creating, and supplying sediment to an active submarine foreland-type basin. The flysch was derived from, and was progressively accreted to and overridden by the Taconic Allochthon, resulting in the formation of belts of tectonic melange.

The mélange belt is bounded on the west by the Vischer Ferry Thrust (Figs 1 & 3), which separates it from non-deformed flysch (Plesch 1994).

Chert blocks within the mélange contain *N. gracilis* and *C. bicornis* faunas, which are consistent with lithologies of the Mount Merino Formation. The matrix ages reported from wildflysch in the mélange directly beneath the allochthons are *C. americanus* to *D. spiniferus* (Berry 1962, 1977; Jacobi and Mitchell 2018). According to Plesch (1994) and illustrated by Jacobi and Mitchell (2018) in their figure 7, and our Figure 3, the youngest blocks and matrix faunas in the zone are *D. spiniferus,* which is also the faunal age for the undeformed Schenectady Formation to the west of the Vischer Ferry Thrust. Thus, initial thrusting of the Giddings Brook allochthon must be younger than the uppermost *N. gracilis* and lower *D. multidens* zones found in the Mount Merino–Pawlet formations, and final emplacement must have been after or during the *D. spiniferus* Zone but before the *G. pygmaeus* Zone (Fig. 3).

The Rowe–Hawley zone and location of Iapetan suture

Another group of tectonic slices (from west to east and structurally upwards: Rowe schist, Moretown Formation and Hawley belt, collectively grouped as the Rowe–Hawley zone: Stanley and Hatch 1988) are located east of the Taconic allochthons and the Green Mountain Massif and west of the Bronson Hill anticlinorium. The Rowe schist comprises pelitic schist, commonly with garnet, staurolite or kyanite porphyroblasts, quartzite, and amphibolite with lozenges and blocks of ultramafic rocks located near the contact with the structurally overlying Moretown Formation (Stanley and Hatch 1988). The Rowe schist is probably best considered a tectonic, rather than a stratigraphic, unit as lithologies carry for great distances but are lenticular in detail, bounded by faults and are highly strained (Stanley and Hatch 1988). Rocks of the Rowe schist have detrital zircon ages believed to be characteristic of the North American Grenville with broad humps 1500–950 Ma with first-order peaks at about 1200 and 1000 Ma, and second-order peaks at about 2700, 1850 and 550 Ma (Macdonald *et al.* 2014).

Metamorphic rocks of the overlying Moretown Formation form a narrow, but continuous, thrust slice of granofels, quartzite, chert, carbonaceous pelite, and a variety of amphibolites, carbonates, pillow basalt, sparse rhyolite and volcaniclastic rocks (Stanley and Ratcliffe 1985), cut by an unnamed 496 Ma trondhjemite and the 502 Ma Newfane tonalitic pluton (Aleinikoff *et al.* 2011; Karabinos *et al.* 2017). Macdonald *et al.* (2014) examined detrital zircons in the Moretown Formation and found a prominent peak between 650 and 520 Ma, which they argued was different from Laurentian rocks found to the west, such as the Rowe schist, and more typical of Gondwanan sources. This led them to argue that the principal Iapetan suture was the contact between the Rowe belt and the Moretown Formation. The age of deformation within the package is younger than 475 Ma, the age of tonalitic gneiss that contains the same deformational fabrics as its wall rocks, and older than 445 Ma, the age of a granitoid pluton that cuts the Rowe–Moretown contact (Karabinos and Williamson 1994; Macdonald *et al.* 2014). Locally within and above rocks of the Moretown Formation is a discontinuous amalgam of black carbonaceous phyllite, quartzite and mafic volcanic to hypabyssal intrusive rocks collectively termed the Cram Hill Formation (Coish *et al.* 2015). Despite conflicting U–Pb ages in several places along strike, they suggested 466 Ma as a reasonable age for the formation.

The more easterly Hawley Formation is dominated by metabasalt and andesite, in places displaying relict pillow structures, and lesser amounts of black schist and light-coloured gneiss of likely siliceous volcanic origin (Stanley and Hatch 1988). The Hawley Formation and its equivalents along strike in Vermont fall in the age range 485–465 Ma, and are part of what Karabinos *et al.* (1998) termed the Shelburne Falls arc (Fig. 1). Rocks of the formation are cut by the 475 Ma Dell trondhjemite (Karabinos *et al.* 2017).

Kim and Jacobi (1996) examined amphibolites in the Hawley Formation of Massachusetts and found them to have upper-plate arc-like compositions, as well as boninitic geochemical signatures, which, based on much younger analogues, they argued represented forearc magmatism. They also noted the compositional similarity between amphibolites of the Hawley Formation and the subjacent Moretown Formation.

To the east, rocks of the Hawley Formation are overlain by Silurian–Devonian cover of the Connecticut River–Gaspé Trough and the Mesozoic Hartford rift basin (Fig. 1) but several domes, such as the Shelburne Falls and Goshen domes, poke through the younger cover and contain 475–470 Ma volcanic rocks with arc-like geochemistry (Karabinos *et al.* 1998). Still farther east, along the eastern side of the younger cover, arc rocks of the Bronson Hill anticlinorium crop out (Fig. 1).

Bronson Hill anticlinorium

The Bronson Hill anticlinorium extends for over 400 km from southern Connecticut to Maine, and is a north- to northeasterly-trending belt containing nearly two dozen structural domes (Hollocher *et al.* 2002) overlain both east and west by younger cover (Fig. 1). The domes formed during the Devonian Acadian Orogeny and have long been interpreted as giant fold nappes (Thompson 1956; Thompson *et al.* 1968; Robinson *et al.* 1979). Robinson *et al.* (1991) provided an excellent summary of geological research and ideas on these remarkably complex structures.

The domes are historically referred to as the Oliverian domes, which have gneissic cores mantled by Ammonoosuc volcanics and other metasedimentary and metavolcanic rocks (Billings 1937; Leo 1991; Tucker and Robinson 1990), but here we restrict the term ‘Oliverian’ to the gneissic plutonic rocks in the domes to avoid confusion. Because of complex deformation and high metamorphic grade, the geology is commonly bewildering and nearly always controversial. Nevertheless, detailed mapping, modern ICP-MS geochemistry and improved U–Pb zircon techniques have led to a general understanding of the geology and its temporal relationships. Most workers seem to agree that Ordovician metasedimentary and metavolcanic sequences of the Bronson Hill anticlinorium are stratigraphically, compositionally and temporally similar to those of the Moretown–Hawley–Shelburne Falls slices to the west, and so have collectively been interpreted to represent a single or composite arc complex (Tucker and Robinson 1990; Moench and Aleinikoff 2003; Karabinos *et al.* 2017).

The oldest rocks in the Bronson Hill anticlinorium occur within the Pelham dome of western Massachusetts (Fig. 1) and are 613 Ma gneisses of presumed Gondwanan affinity (Tucker and Robinson 1990). They, and adjacent gneisses, are structurally overlain by Late Ordovician intrusive gneisses characteristic of the cores of other domes, as well as a sequence of metasedimentary rocks as young as Devonian: all deformed into east-vergent recumbent fold-thrust nappes (Robinson *et al.* 1992).

Some Oliverian gneissic units in the domes have different age rocks within a single core. For example, the amphibolite-grade Monson Gneiss of central Massachusetts (Fig. 1) has high-Sr intermediate composition gneisses dated at 445 Ma and low-Sr gneisses dated at 454 Ma, as well as high- and low-Nb amphibolites of inferred similar ages (Tucker and Robinson 1990; Hollocher *et al.* 2002). In addition to the high-Sr in the younger rocks, Hollocher *et al.* (2002) reported that they had no Eu anomalies, are depleted in heavy REEs and have low Y contents. These features are similar to post-collisional slab-failure rocks in the Cordillera and elsewhere (Hildebrand and Whalen 2017; Whalen and Hildebrand 2019).

The Ammonoosuc volcanics range in thickness from 30 to 1200 m and consist of a lower member of dominantly tholeiitic basalt and andesite with minor low-K dacite, now with metamorphic assemblages of cummingtonite, gedrite and gedrite–cordierite, which are interpreted as hydrothermally altered seafloor basalt (Schumacher 1988; Robinson *et al.* 1998). Carbonate beds and marble–matrix volcanic conglomerate occur near the top of the lower member, whereas the overlying middle member comprises thin (<30 m) garnet–amphibole-bearing quartzites, interpreted as exhalative rocks, overlain by an upper member of metadacite, metarhyolite and minor amphibolite (Schumacher 1988; Hollocher 1993). A quartz-phyric rhyolite near the top of the formation in Massachusetts yielded a U–Pb zircon multigrain age of 453 ± 2 Ma (Tucker and Robinson 1990).

Overlying, and compositionally similar to the underlying Ammonoosuc volcanics, are metamorphosed sulfidic black shale and arkosic sandstone intercalated with bimodal volcanic rocks that constitute what is termed the Partridge Formation (Hollocher 1993). During the Acadian Orogeny, the volcanic rocks in the formation were regionally metamorphosed in the kyanite + muscovite + staurolite through to the sillimanite + orthoclase zones and are now gneisses, whereas mafic volcanic rocks are amphibolites with compositions resembling low-K tholeiites (Hollocher 1985, 1993). A rhyolitic tuff in the Partridge Formation of Massachusetts (Fig. 3) was dated by U–Pb zircon multigrain to be $449 + 3/-2$ Ma (Tucker and Robinson 1990). The Partridge Formation farther north in New Hampshire is older (Fig. 3) as it contained graptolites characteristic of the *N. gracilis* Zone (Karabinos *et al.* 2017). A recent study of detrital zircons from two locations of the Partridge Formation suggests a maximum age of 452 Ma on the basis of the single youngest zircon, as well as a Gondwanan source due to the presence of 650–560 Ma zircons (Merschat *et al.* 2016). They also found that younger metasedimentary rocks contained a mixture of inferred Laurentian and Gondwanan zircons.

Another group of metasedimentary and volcanic rocks, the Quimby sequence, unconformably overlies rocks of the Partridge Formation and is exposed on the northern Bronson Hill anticlinorium, where it is dominated by lithic metagreywackes grading laterally into metarhyolitic flow domes and metatuff with one tuffaceous unit dated at 443 ± 4 Ma (Moench and Aleinikoff 2003). Several plutons of the Highlandcroft plutonic suite, which are plutons similar to the Oliverian bodies but located on the NW flank of the anticlinorium, intrude the Ammonoosuc volcanics and are 450–441 Ma (Lyons *et al.* 1986, 1997; Moench and Aleinikoff 2003).

Over the years, siliceous metavolcanic rocks within the Ammonoosuc volcanics were dated by multigrain TIMS but, just as elsewhere, early multigrain zircon samples were plagued by xenocrystic cores, so ages might not be robust. More modern sensitive high-resolution ion microprobe (SHRIMP) work in northern New Hampshire and Vermont by Aleinikoff *et al.* (2015) indicate that Ammonoosuc volcanics and related plutons are as old as 477 $\pm$ 7 Ma and as young as 447 $\pm$ 5 or 449 $\pm$ 7 Ma.

Age of Taconic thrusting

Jacobi and Mitchell (2018) provided a detailed summary of the various constraints on the age of Taconic thrusting, and what follows largely utilizes their biostratigraphic controls on timing, with minor changes as noted. The basal part of the Mount Merino Formation contains only *N. gracilis*, which, according to Landing (1988), also occurs in the underlying Indian River Formation. As there is compelling sedimentological evidence that rocks of the Indian River Formation mark the passage of the continental slope over the outer bulge to the trench (Landing 2012), the continental slope was passing over the bulge at about 458–457 Ma (Fig. 3). Rocks of the Indian River Formation pass conformably upwards into green and black shales of the Mount Merino Formation, which reflect transit down the western and uppermost slope of the trench. Shales of the Mount Merino Formation grade upwards into Pawlet turbidites, which, according to Rowley and Kidd (1981), contain deformed metasedimentary and volcanic clasts, as well as detrital chromite, and become thicker, coarser and more abundant eastwards. Sedimentary rocks of both the Mount Merino and Pawlet formations contain graptolites characteristic of the *Nemagraptus gracilis–Climacograptus bicornis* and *D. multidens* zones (Berry 1962; Riva 1974), and have the same deformational history. This indicates that the units were involved in thrusting after deposition of the turbiditic Pawlet Formation, which, based on fauna, was most likely to be of late *C. Bicornis* age or about 455–453 Ma (Jacobi and Mitchell 2018).

The termination of thrusting appears to be younger than the Cohoes mélange beneath and immediately west of the Taconic allochthon, as the mélange contains sedimentary clasts and matrix with graptolites of the *D. spiniferus* Zone, as do turbiditic rocks of the Schenectady Formation that are located west of the Vischer Ferry frontal thrust (Jacobi and Mitchell 2018). Thus, the uppermost age of the mélange approximates the age of sedimentary flysch deposited on the autochthon and so marks the emplacement of the Taconic allochthon and end of thrusting (Fig. 3). Graptolites from the *G. pygmaeus* Zone are not reported in the mélange, so 451 Ma is a reasonable estimate for the end of thrusting. This fits well with estimates by Jacobi and Mitchell (2018) based on five soft-sediment deformation zones containing graptolites of the *D. spiniferus* Zone in drill core, as well as with decapitated westward-directed slump folds beneath a major unconformity at the top of the thin Dolgeville carbonate (Fig. 3) within the foredeep succession.

It would be expected for molasse to be found atop the foredeep flysch, which would mark the end of thrusting and exhumation of the hinterland belt, but the top of the Schenectady Formation is eroded and not present in the area. However, in southern New York, eastern Pennsylvania and northwestern New Jersey, the Upper Ordovician Martinsburg flysch, containing graptolites of the *D. spiniferus* Zone near the top, is unconformably overlain by a 600 m-thick shallow-marine to fluvial, conglomeratic, clastic wedge of the Shawangunk Formation, which is undated but older than about 440 Ma, the age of the overlying Bloomsburg Red Beds (Epstein and Epstein 1972; Epstein and Lyttle 1987; Epstein 1993).

Thus, while it could be a few million years younger, we will use 451 Ma as the age for the end of thrusting with the Taconic foreland. This means that Taconic thrusting was short-lived, some 2–4 myr.

Pre- and post-thrusting magmatism

Over the past few years we discovered that magmatic rocks erupted and intruded in arcs had compositions distinct from those formed after collision by the process of slab failure (Hildebrand and Whalen 2014, 2017; Hildebrand *et al.* 2018), and so we developed a suite of discrimination diagrams to separate the two suites (Whalen and Hildebrand 2019). Our overall model for the generation of slab-failure magmas involves tearing of the subducted slab, followed by partial melting of garnetiferous, plagioclase-free metabasalt and gabbro from subducted oceanic slabs at depths of >2 GPa to create siliceous magmas with the diagnostic trace element patterns, $Nb/Y > 0.4$, $La/Yb > 10$, $Gd/Yb > 2.0$, $Sm/Yb > 2.5$ and $Sr/Y > 10$.

During arc–continent collision, the leading edge of the continent is pulled beneath the arc and underplates it with continental lithosphere. In cases where the subcontinental lithospheric mantle (SCLM) is old and enriched, the rising slab-failure melts assimilate varying amounts of elements with large ionic ratios (LILs) and develop Sr and Nd isotopic signatures commonly considered to be of crustal origin (Hildebrand *et al.* 2018). Where the lower partially subducted plate is young and does not contain old, enriched SCLM, such as the Coast plutonic complex of British Columbia, radiogenic isotopic compositions remain relatively primitive and appear arc-like (see Wetmore and Ducea 2011).

While we were confident that our model was actualistic and robust, we were unsure how well our discrimination diagrams would work at higher metamorphic grades and/or in polydeformed rocks. Consequently, and even though we understood that the age of the Taconic Orogeny was becoming more precisely determined, we knew that for the most part we could not use field data on the relative age of the plutons to the Taconic Orogeny because in New England the rocks were deformed

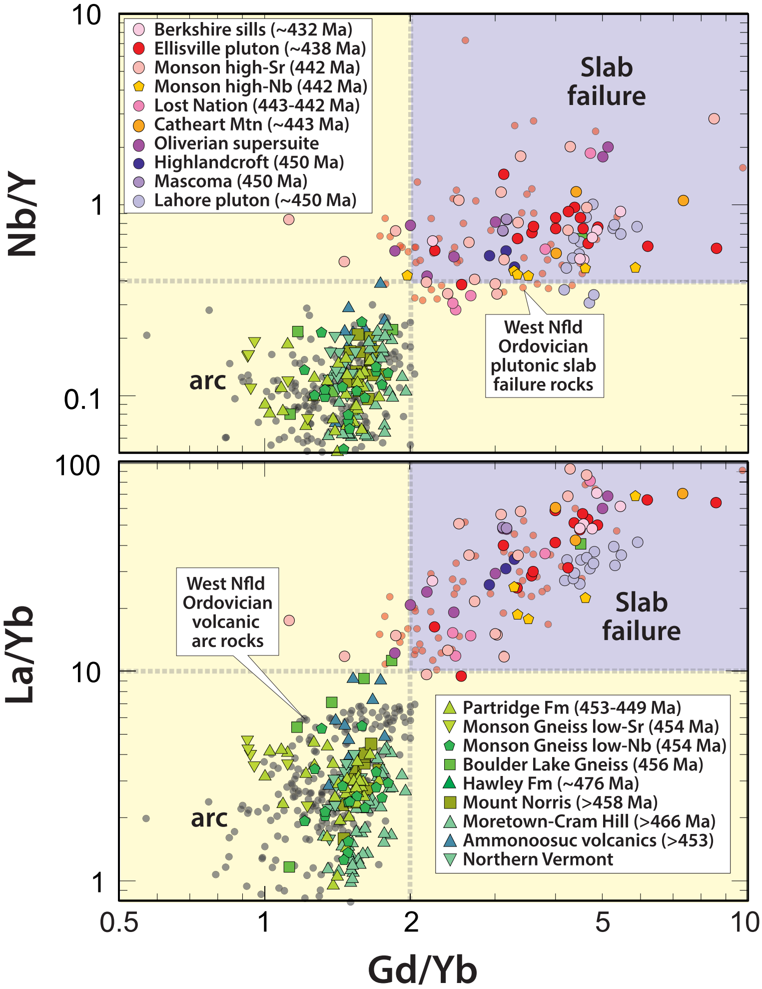

Fig. 4. Discrimination diagrams using Nb/Y and La/Yb v. Gd/Yb (Hildebrand and Whalen 2017; Whalen and Hildebrand 2019) for various pre- and post-Taconic magmatic units illustrating the two different types of magmatism: Catheart Mountain (Ayuso 1989); Moretown and Cram Hill (Coish *et al.* 2015); Oliverian (Dorais *et al.* 2008); Ammonoosuc volcanics (Schumacher 1988; Dorais *et al.* 2012); Berkshire sills (Karabinos *et al.* 2008; and unpublished data); Monson Gneiss (Hollocher *et al.* 2002); Hawley Formation (Kim and Jacobi 1996); northern Vermont (Kim *et al.* 2003); Lahore and Ellisville plutons (Pavlides *et al.* 1994; Hughes *et al.* 2013). Boulder Mountain Gneiss appears to be a composite unit based on the two different compositions in these and subsequent discrimination plots. As labelled, small pale dots show arc and slab-failure rocks from western Newfoundland for comparison (Rogers 2004; Zagorevski 2008; Whalen 2012).

and metamorphosed during several younger orogenies, such that we had to rely heavily on U–Pb zircon geochronology of the gneissic plutons to determine whether rocks were pre- or post-Taconic. Nevertheless, we decided that the Appalachian gneisses would provide a good test case to see how well our slab-failure model would work in higher-grade polydeformed rocks.

The gneisses in the cores of domes can be perplexing because they are commonly composite, which produces complexities that render them difficult to work with, at least with our present state of knowledge. For example, the Newtown Gneiss of Connecticut (Sevigny and Hansen 1995) contains quartz diorite, dated at 466 ± 2Ma, granodiorite with an U/Pb age of 436 ± 2 Ma, and 438 ± 2 Ma granite; whereas the Monson Gneiss of Massachusetts comprises magmatic rocks of 445 and 454 Ma (Tucker and Robinson 1990), each containing rocks of markedly different compositions and likely sources (Hollocher *et al.* 2002). Some core gneisses have a lot of existing geochemistry but are undated; whereas others are dated but geochemistry is scarce, non-existent or, in cases where there are analyses, does not include the trace elements necessary for useful discrimination. In some cases, the geochemistry obviously reflects multiple sources but one or both suites are undated, rendering them useless for our purposes.

The foregoing notwithstanding, the results of our examination (Figs 4 & 5) show that the magmatic rocks in western New England fall into two groups: arc and slab failure, depending on their age. Arc compositions predate 450 Ma, whereas many intrusive rocks on the Bronson Hill anticlinorium, such as some of the Oliverian suite, are as old as 450 Ma and have slab-failure signatures (Figs 4 & 5). These rocks were previously considered to be problematic because they were thought to be arc rocks that post-dated the Taconic Orogeny but they appear to be slab-failure rocks and not arc rocks, so they place a minimum age for Taconic deformation consistent with the faunal evidence discussed earlier.

The results validate the earlier ideas of Hollocher *et al.* (2002), who used the differences in compositions and sources to suggest that slab delamination and melting of plagioclase-free, garnetiferous metabasalt played a role in the origin of high-Sr, low-Nb gneisses, depleted in HREE, found in the Oliverian core gneisses of some domes. Additionally, they observed that while compatible elements in those rocks showed orderly trends, incompatible elements did not behave as regularly and that, therefore, they might reflect the source rocks rather than fractionation. Our examination of Cretaceous post-collisional granodioritic plutons in the Sierra Nevada and Peninsular Ranges batholiths (Hildebrand *et al.* 2018) found not only the same geochemical features, such as high Sr, HREE depletion and decoupling of incompatible trace elements, but also marked differences in Nd and Sr isotopic compositions between arc and slab-failure magmas.

In our earlier Cordilleran studies we concluded that slab-failure magmas are largely derived from the melting of garnet-bearing, plagioclase-free metamorphosed oceanic lithosphere, and as they rose through old, enriched SCLM, their source-derived Sr and Nd isotopic ratios and LIL element (LILE) concentrations were modified by fractional melting of the mantle lithosphere. In the Sierra Nevada batholith of California, post-collisional granodioritic plutons have mantle values for $\delta^{18}O$ with negative ε_{Nd} as low as -7 and initial Sr isotopic ratios of >0.705, generally considered to be of crustal origin but yet identical to garnet pyroxene xenoliths derived from the SCLM, and to much younger basaltic lavas of the Snake River Plain and Big Pine lavas (Hildebrand *et al.* 2018).

This explains why Dorais *et al.* (2008, 2012) found that rocks of the Oliverian, Highlandcroft

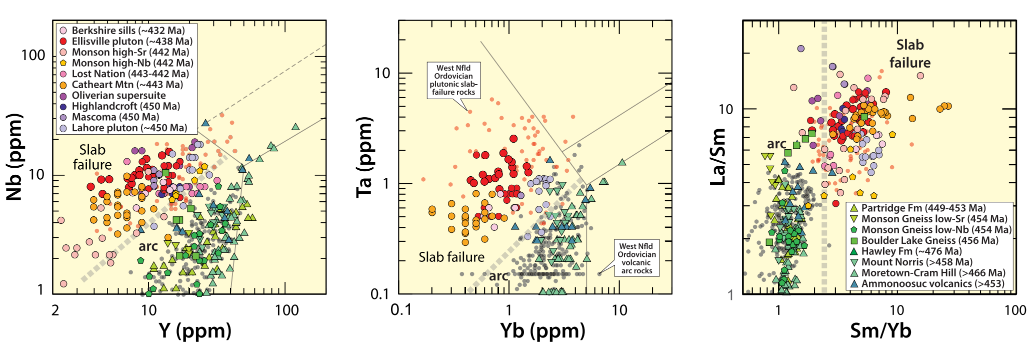

Fig. 5. Discrimination diagrams using Nb v. Y, Ta v. Yb and La/Sm v. Sm/Yb for the same pre- and post-Taconic magmatic suites as in Figure 4 (Hildebrand and Whalen 2017; Whalen and Hildebrand 2019) illustrating the two different types of magmatism. Data sources are as in Figure 4.

and Quimby suites – all about the same age and younger than the Taconic Orogeny – are compositionally and isotopically similar in that they have Laurentian isotopic signatures but contain entirely different signatures than rocks of the older Ammonoosuc volcanics, which have more primitive Gondwanan isotopic signatures. However, because they did not consider the possibility of cratonic underplating of the arc during collision, followed by slab-failure magmatism and its assimilation of old, enriched cratonic SCLM, they invoked early easterly subduction that ended at about 470 Ma with the accretion of the Ammonoosuc arc to Laurentia followed by outward-stepping, and westward dipping, subduction beneath the amalgamated Laurentian–Ammonoosuc block to generate magmatism of the Oliverian, Highlandcroft and Quimby suites (Dorais *et al.* 2012).

Conclusion

The observation that the 450 Ma and younger Taconic magmatic rocks are most likely post-collisional slab-failure rocks and not arc rocks means that arc magmatism is not younger than the 455–451 Ma Taconic Orogeny; thus, there is no conflict with a simple eastward subduction and continent–arc collisional model for the Taconic Orogeny: multiple collisions and flipping subduction zones are unnecessary and without basis. A simple collisional model for the Taconic Orogeny is further supported by the complete lack of 'Early Ordovician deformation, foreland basin deposits, or air-fall tephras on the Laurentian carbonate platform to the west' (Karabinos *et al.* 2017, p. 537). Given that within the proximal Upper Ordovician foreland basin of the Taconics (Sell *et al.* 2015), there are nearly 60 high-K bentonites in rocks of just two graptolite zones (*Corynoides americanus–Diplacanthograptus spiniferus*), it would be remarkable if an Ordovician arc approached the eastern shore of Laurentia, collided with it and left no evidence in the form of ash beds, let alone collisional deformation, metamorphism and foredeep sedimentation. Therefore, we favour a simple Upper Ordovician collisional model for the Taconic Orogeny, which like most reasonably well understood, young and ongoing continent–arc collision belts, such as Papua, New Guinea (Cloos *et al.* 2005) and Taiwan (Teng 1996), was short lived, lasting but a few million years.

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Data availability statement Geochemical data analysed in this paper are available from the publications listed as sources for Figure 4, except the data for Berkshire sills, which can be obtained from the cited author upon reasonable request.

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