Triggering of the largest Deccan eruptions by the Chicxulub impact

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Abstract

New constraints on the timing of the Cretaceous-Paleogene mass extinction and the Chicxulub impact, together with a particularly voluminous and apparently brief eruptive pulse toward the end of the “main-stage” eruptions of the Deccan continental flood basalt province suggest that these three events may have occurred within less than about a hundred thousand years of each other. Partial melting induced by the Chicxulub event does not provide an energetically plausible explanation for this coincidence, and both geochronologic and magnetic-polarity data show that Deccan volcanism was under way well before Chicxulub/Cretaceous-Paleogene time. However, historical data document that eruptions from existing volcanic systems can be triggered by earthquakes. Seismic modeling of the ground motion due to the Chicxulub impact suggests that the impact could have generated seismic energy densities of order 0.1–1.0 J/m^3 throughout the upper ~200 km of Earth’s mantle, sufficient to trigger volcanic eruptions worldwide based upon comparison with historical examples. Triggering may have been caused by a transient increase in the effective permeability of the existing deep magmatic system beneath the Deccan province, or mantle plume “head.” It is therefore reasonable to hypothesize that the Chicxulub impact might have triggered the enormous Poladpur, Ambenali, and Mahabaleshwar (Wai Subgroup) lava flows, which together may account for >70% of the Deccan Traps main-stage eruptions. This hypothesis is consistent with independent stratigraphic, geochronologic, geochemical, and tectonic constraints, which combine to indicate that at approximately Chicxulub/Cretaceous-Paleogene time, a huge pulse of mantle plume–derived magma passed through the crust with little interaction and erupted to form the most extensive and voluminous lava flows known on Earth. High-precision radioisotopic dating of

the main-phase Deccan flood basalt formations may be able either to confirm or reject this hypothesis, which in turn might help to determine whether this singular outburst within the Deccan Traps (and possibly volcanic eruptions worldwide) contributed significantly to the Cretaceous-Paleogene extinction.

INTRODUCTION

The hypothesis that the ca. 66 Ma Cretaceous-Paleogene mass extinction event was caused by an extraterrestrial impact (Alvarez et al., 1980; Smit and Hertogen, 1980) stands in contrast with a competing hypothesis that the Cretaceous-Paleogene extinction was caused by the Deccan Traps continental flood basalt eruptions (Courtillot et al., 1988, 2000; Courtillot and Renne, 2003). The impact hypothesis is supported by the presence of impact deposits at the Cretaceous-Paleogene boundary, the discovery of the ~180-km-diameter Chicxulub crater in Yucatán, Mexico (Hildebrand et al., 1991; Schulte et al., 2010), and increasingly precise radioisotopic dating (Renne et al., 2013) showing that the Chicxulub impact iridium layer and the Cretaceous-Paleogene boundary are essentially coincident in time. The Deccan continental flood basalts started several million years prior to the Cretaceous-Paleogene boundary (Courtillot and Renne, 2003) and were therefore not initiated by the Chicxulub impact. However, the volcanic hypothesis for the Cretaceous-Paleogene extinction has persisted for several reasons: Massive continental flood basalt eruptions from the Siberian Traps and Emeishan Traps were approximately coincident with the extinctions at end-Permian and end–middle-Permian (end-Guadalupian) time, respectively, and the Central Atlantic magmatic province was erupted at approximately the Triassic-Jurassic boundary (Courtillot and Renne, 2003). Also, there is no compelling evidence to date for other impact/extinction associations (Alvarez, 2003). Moreover, constraints on the likely “kill mechanisms” for these mass extinctions (e.g., CO_2 outgassing, launching of sulfur gas leading to sulfate aerosols) are compatible with both impact and volcanic causes (Self et al., 2006, 2014; Black et al., 2014).

Recent high-precision ^{40}Ar-^{39}Ar dating of the Cretaceous-Paleogene boundary in the Hell Creek area (Montana, USA) and of the Chicxulub (Yucatán, Mexico) impact ejecta shows that these two events are time-coincident within ~32,000 yr precision at ca. 66.04 Ma (Renne et al. 2013), and both fall within paleomagnetic chron 29R (Chenet et al., 2007; Ogg, 2012). Radioisotopic constraints on Deccan-related volcanism are much less precise, but they show that early- and late-stage alkalic eruptions preceded and postdated Cretaceous-Paleogene time by at least several million years (Basu et al., 1993). Moreover, the Cretaceous-Paleogene boundary is thought to have occurred within chron 29R, whereas the “main-phase” tholeiitic basalt eruptions of the Western Ghats Province, during which at least 90% of Deccan lavas are thought to have erupted, began during (or before) chron 30N and ended during chron 29N (Chenet et al., 2008, 2009). Thus, Deccan volcanism was well under way at the time of the Cretaceous-Paleogene extinction and Chicxulub impact time, and it continued after Cretaceous-Paleogene time (Fig. 1A).

Recent studies of Deccan volcanic stratigraphy (Fig. 1A) have indicated an extraordinary pulse of basaltic volcanism that accounts for more than half the total volume of Deccan volcanism and includes the huge Poladpur, Ambenali, and Mahabaleshwar Formations within the Wai Subgroup of the Deccan continental flood basalt (Self et al., 2006). Magnetostratigraphic constraints have been interpreted to suggest that much of the Wai Subgroup volcanism occurred prior to the 29R/29N reversal during a brief interval of time, perhaps as little as ~100,000 yr (Chenet et al., 2008, 2009), but not likely more than several hundred thousand years, as deduced in part from the small amount of paleomagnetic secular variation recorded by these formations. Micropaleontological evidence from “intertrappean” sediments between lava flows suggests that the Cretaceous-Paleogene boundary may lie within or perhaps just below the Wai Subgroup flows, as discussed later in this paper. Pre-impact Deccan volcanism may also have played a role in pre–Cretaceous-Paleogene climate oscillations (Barrera and Savin, 1999; Li and Keller, 1998; Wilson, 2005; Wilf et al., 2003). In any case, it seems reasonable to assume that if the Deccan Traps eruptions contributed to the main Cretaceous-Paleogene extinctions, then the Wai Subgroup pulse of eruptions was the likely culprit. Although the timing and duration of the Kalsubai, Lonavala, and Wai Subgroup formations are not well constrained, it also appears likely that the Cretaceous-Paleogene boundary and the Chicxulub impact may have occurred at or just before the onset of the huge Wai Subgroup eruptions (e.g., Keller et al., 2012).

Chicxulub-size impact events occur perhaps only several times per billion years, and main-phase continental flood basalt eruptions such as the Deccan Wai Subgroup lava flows have occurred on average only about every 20–30 Ma through the Phanerozoic (Courtillot and Renne, 2003). Therefore, if we take the typical durations of main-stage flood basalt events to be ~2–3 Ma, the likelihood that a large impact may have occurred during an ongoing flood basalt event sometime during the Phanerozoic, with increased environmental consequences, is not small—perhaps on the order of one chance in ~10 (that is, 20–30 Ma/2–3 Ma). However, if, as the evidence presented herein suggests, the Chicxulub impact occurred within only ~100,000 yr or so of the Wai Subgroup outburst of Deccan volcanism, then the odds of this occurring by chance would be an order of magnitude smaller, i.e., one chance in ~100. Such small odds provide a new impetus to explore plausible causal links.

The possibility that an impact at Cretaceous-Paleogene time caused Deccan volcanism has been investigated since the discovery of the iridium anomaly at Cretaceous-Paleogene boundary (Alvarez et al., 1980; Smit and Hertogen, 1980), with an emphasis on antipodal focusing of seismic energy. However, the Deccan continental flood basalts were not antipodal to the 66 Ma Chicxulub crater at the time of the impact, but instead separated by an epicentral distance of ~130° (Williams et al., 2012; Chatterjee et al., 2013).

Also, a Chicxulub-size impact does not in any case appear capable of generating a large mantle melting event (Ivanov and Melosh, 2003). Thus, impact-induced partial melting could not have caused the initiation of Deccan volcanism, consistent with the occurrence of Deccan volcanism well before Cretaceous-Paleogene/Chicxulub time.

Instead, Deccan volcanism is widely thought to represent the initial outburst of a new mantle plume “head” at the beginning of the Réunion hotspot track (Morgan, 1981; Duncan and Pyle, 1988; Richards et al., 1989; Campbell and Griffiths, 1990). As for the Deccan Traps, studies of other continental flood basalt provinces reveal two distinct time scales: a brief ~0.5–2.0 Ma “main phase” during which typically >80% of the total basalt volume is erupted, which itself occurs within a much longer ~10 Ma time scale representing both pre– and post–main-phase eruptions (Courtillot and Renne, 2003). Mantle plume models do not satisfactorily explain the existence of these two distinct time scales (e.g., Farnetani and Richards, 1994; Leitch and Davies, 2001; Karlstrom and Richards, 2011), but acceleration of decompression melting may occur due to lithospheric thinning (White and McKenzie, 1989) or instability of the lower lithosphere (Elkins-Tanton and Hager, 2000; Sobolev et al., 2011). In any case, a region of partially molten mantle of diameter ~1000 km or larger (Richards et al., 1989; White and McKenzie, 1989; Campbell and Griffiths, 1990) is presumed to have already been active beneath the Deccan region when the Chicxulub impact occurred.

The hypothesis we explore in this paper is that the Chicxulub impact may have triggered an anomalously large outburst of Deccan volcanism, as represented by the Wai Subgroup formations. We begin with a critical examination of the eruption history of the “main-phase” Deccan basalts. We then explore the possibility that the Chicxulub impact might have triggered eruptions from the Deccan plume head. Finally, we examine a variety of other geological and geochemical evidences that appear to be consistent with our hypothesis, and we suggest further tests.

DECCAN VOLCANIC STRATIGRAPHY AND VOLUME ESTIMATES

The volcanic stratigraphy of the Western Ghats Province of the Deccan Traps has been established through a variety of field studies (Cox and Hawkesworth, 1985; Beane et al., 1986; Lightfoot et al., 1990; Peng et al., 1994; Subbarao et al., 1994), resulting in the widely used nomenclature portrayed in Figure 1A, which is based in large part upon the analysis of characteristic major- and minor-element signatures. Within this scheme, the Deccan Traps Flood Basalt Group is subdivided into the Kalsubai, Lonavala, and Wai Subgroups, which are further subdivided into individual formations as shown in Figure 1A. Earlier eruptions include the Narmada/Rajpipla basalts, as well as other formations located generally northward from the main Western Ghats formations. (For plots illustrating the distinct compositional character of the various Deccan formations, the reader is referred to figure 5 *in*Beane et al. [1986].)

The most intensive studies of the Deccan Traps have been carried out in the Western Ghats region (in the area of Mumbai, Pune, and Mahabaleshwar), in large part due to the excellent exposures afforded by the characteristic step-like mountains ("ghats") of this region, which are made up almost entirely of Deccan lava flows. However, this region represents only a small fraction of the total areal exposure of Deccan lavas. Following previous workers and estimating formation volumes by using only mapped formation thicknesses while assuming that they all have the same areal extent, as in Figure 1B, column ii, the estimated volumes of the Kalsubai, Lonavala, and Wai Subgroups appear comparable to each other, suggesting perhaps relatively uniform rates of eruption throughout the stratigraphic column. However, a more careful estimate of the volumes using both formation thicknesses and mapped areal extents, as in Figure 1B, column iii, suggests that the Wai Subgroup flows were much larger by comparison. Indeed, the Ambenali and Mahabaleshwar Formations are the most extensive mapped lava flows on Earth, with single-eruption volumes approaching ~10,000 km^3, and with flows >1000 km long crossing the entire Indian subcontinent from the Western Ghats to the present-day Bay of Bengal (Self et al., 2008a).

Figure 2 compares the Ambenali/Poladpur (Wai Subgroup) mapped flow area with the restricted known areal occurrence of the Thakurvadi Formation of the Kalsubai Subgroup. Although it is possible that the Kalsubai and Lonavala Subgroup formations are buried beneath some areas covered by the Wai Subgroup flows and have escaped exposure, the Wai Subgroup flows clearly occur over a vastly greater area. Indeed, without the Wai Subgroup lava flows, the Deccan province might not be considered a major flood basalt event at all compared to other large igneous provinces such as the Siberian, Karoo, and Paraná-Etendeka flood basalts, with volumes of at least 1–3 million km^3.

GEOCHRONOLOGIC AND STRATIGRAPHIC CONSTRAINTS

Existing $^{40}Ar/^{39}Ar$ and Paleomagnetic Timing Constraints

Existing radioisotopic age constraints on Deccan volcanism (see Table 1) are, unfortunately, not very precise. Figure 1A, column ii, includes $^{40}Ar/^{39}Ar$ plateau ages for lavas in the Western Ghats that are assigned to specific formations, published since the year 1990, along with 2σ error bars (Venkatesan et al., 1993; Hofmann et al., 2000; Knight et al., 2003; Pande et al., 2004; Hooper et al., 2010). All ages have been recalculated per the calibration of Renne et al. (2011). As a whole, these data show that the Deccan Basalt Group formations were erupted at approximately Cretaceous-Paleogene boundary time, or ca. 66 Ma, but possibly over a time range as large as 69–64 Ma. Within these ages, if we discount results of Venkatesan et al. (1993) for the Poladpur and Ambenali Formations, which appear to be biased by ^{39}Ar recoil artifacts, in favor of the more reliable ages from Knight et al. (2003) and Hooper et al. (2010), the Wai Subgroup ages are consistent with eruption at or just following Cretaceous-Paleogene boundary time. The

Lonavala Subgroup formations have not been dated yet, and the Kalsubai Formations are only poorly constrained, but presumably were erupted before Cretaceous-Paleogene time.

Extensive paleomagnetic work has been done on the Deccan Traps, with an emphasis on determining how much time may have elapsed during the main-phase basalt eruptions (e.g., Courtillot et al., 1988, 2000; Chenet et al., 2008, 2009), independent of radioisotopic age constraints. Remarkably, nearly all samples from the Neral Formation up through the lower part of the Mahabaleshwar Formation show reversed polarity. The lower Mahabaleshwar Formation records a change to normal polarity, which is ascribed to the 29N/29R reversal, consistent with both radioisotopic and other stratigraphic constraints (see following). This has led some researchers to conclude that the bulk of Deccan lavas were erupted during magnetic polarity chron 29R, although other researchers have reported a few normal polarity samples within these same formations (Pande, 2002). If we follow Chenet et al. (2008, 2009) in ascribing everything between the Neral and Mahabaleshwar Formations to chron 29R, then this enormous stack of lavas may have occurred in as little as the ~350,000–710,000 yr duration of chron 29R (Sprain et al., 2014; Ogg, 2012). Indeed, Chenet et al. (2008, 2009) concluded that a time interval of as little as a few tens of thousands to perhaps only ~100,000 yr may have elapsed while these formations were actively erupted, based in large part upon an episodic lack of secular variation in paleomagnetic samples throughout the stack. This conclusion would be inconsistent, however, with the relatively well-determined $^{40}Ar/^{39}Ar$ ages for the Thakurvadi Formation from Venkatesan et al. (1993), which might place this formation in chron 31R, given its reversed polarity (Ogg, 2012).

Biostratigraphic Constraints

In addition to radioisotopic dating and timing evidence from magnetic polarity stratigraphy, there are also potential age constraints on the Deccan volcanic formations from fossils in sedimentary rocks associated with the volcanic units. In any given locality, Cretaceous sedimentary deposits under the lowest basalt are referred to in the literature as “infratrappeans,” sedimentary layers between basalt flows are called “intertrappeans,” and deposits above the highest basalt are called “supratrappeans.” The intertrappeans are potentially the most useful for this purpose.

Evidence for the presence of dinosaurs is common in the area of the Deccan Traps, with dinosaur eggs of particular interest, since these often occur in nests or egg clutches that must be in situ and not transported or reworked. Assuming that the nonavian dinosaurs became extinct at the time of the Chicxulub impact, and if dinosaur egg clutches were found in intertrappeans within the Poladpur, Ambenali, or Mahabaleshwar Formation, this might argue against our hypothesis. However, if both the impact and Deccan volcanism combined to cause the Cretaceous-Paleogene extinctions, it is possible that the extinction of some species continued during the Wai Subgroup eruptions following the impact.

Most dinosaur eggs come from the infratrappeans such as the Lameta beds and therefore simply support the conclusion that the Cretaceous-Paleogene boundary occurred within the span of Deccan volcanism. One possible exception occurs in the Mandla Lobe of the northeast Deccan, at Ranipur, 28 km east of Jabalpur, where there is a report of a “large, well preserved fossil of a dinosaur pelvic girdle” in a 2-m-thick sedimentary layer between two basalt units (Kumar et al., 1999). This part of the Deccan Traps has been identified geochemically as composed of Poladpur overlain by Ambenali Formation (Peng et al., 1998; Vanderkluysen et al., 2011). It will be important to determine whether this fossil is articulated and therefore not reworked, and whether it is in the Poladpur or the Ambenali Formation, or some other formation. Another possible exception is at Anjar, Gujarat, where dinosaur eggshells and bones have been reported from intertrappeans (Bajpai and Prasae, 2000); the age of the basalts from this area has not yet been determined precisely. Other dinosaur fossils within the intertrappeans must be carefully sought and evaluated.

More definitive paleontological evidence may come from planktic foraminifera in places where intertrappean sediments were deposited in marine or near-marine conditions. In a remarkable set of recent papers, Keller and colleagues have been able to find planktic foraminifera of late Maastrichtian and Danian age closely associated with the Deccan Traps, although the sedimentological evidence for environments of deposition needs to be carefully studied. At Jhimili in the Mandla Lobe, 150 km southwest of Jabalpur, Keller et al. (2009) found earliest Danian (P1a zone) planktic foraminifera in a 14 m sedimentary unit between two lava flows. If these foraminifera are not reworked, and if the underlying basalt belongs to the Ambenali Formation, probably present throughout the Mandla lobe (Peng et al., 1998; Vanderkluysen et al., 2011), this discovery of zone P1a foraminifera would suggest that the Cretaceous-Paleogene boundary and the Ambenali flows were nearly time-coincident; geochemical identification of the basalt and environmental analysis of the sediments will thus be critical.

In wells penetrating the subsurface part of the Rajahmundry Traps of southeastern India, now recognized as isolated, preserved distal Deccan flows (Self et al., 2008b), Keller et al. (2012) found Danian zone P1a planktic foraminifera in an intertrappean layer between the two exposed basalt units of the Rajahmundry Traps, which have been identified chemically as Ambenali below and upper Mahabaleshwar Formation above. This at first looks like the same situation as at Jhimili, but Keller et al. (2012) also reported Upper Maastrichtian foraminifera from sedimentary layers intercalated in the lower basalt flows. They considered species seen as single specimens to be reworked, but species with multiple specimens to be in place, and thus put the Cretaceous-Paleogene boundary at the very top of the lower (Ambenali) basalt unit. However, if all the Maastrichtian foraminifera between these basalt flows are reworked, the Cretaceous-Paleogene boundary would lie immediately below the Ambenali (since the Poladpur is apparently not present there), and in agreement with the calcareous nanoplankton report of Saxena and Misra (1994). Clearly, it is critical to test these alternative interpretations, in particular, by careful analysis of the environment of deposition of the sediments within and below the Rajahmundry basalt units, and to determine whether the foraminifera in sedimentary layers in the lower basalt unit are reworked or in place.

Evidence for a Disconformity at the Poladpur-Bushe Formation Contact

The extensive geochemical database used by Beane et al. (1986) to define the classical formations of the Western Ghats shows that the most conspicuous change in overall geochemical characteristics within the Deccan Traps occurs at the Poladpur-Bushe Formation contact. Figure 5 of Beane et al. (1986) shows major changes in the SiO_2, TiO_2, P_2O_5, and various minor- and trace-element signatures. All evidence, including strontium and neodymium isotopic data, suggests that the early Wai Subgroup magmas had less contamination from interaction with the crust than the Lonavala Subgroup (e.g., Mahoney et al., 1982; Lightfoot and Hawkesworth, 1988; Peng et al., 1994; Vanderkluysen et al., 2011), and also much less geochemical variation overall within the Poladpur, Ambenali, and Mahabaleshwar Formations than in lower formations.

Here, we add to this evidence for a stratigraphic time break, or disconformity, at the Poladpur-Bushe contact from some new field observations. During March 2014, a group of the coauthors (Self, Vanderkluysen, Renne, Sprain, Richards), while collecting a new suite of samples for $^{40}Ar/^{39}Ar$ dating, noticed that in the region of the cities of Pune and Mahabaleshwar (Maharashtra Province, India), the Poladpur-Bushe contact defines a conspicuous and extensive set of topographic terraces. Figure 3A shows a photograph of the Poladpur-Bushe contact in a road cut, where we observed an unusually thick weathered zone separating lava flows inferred to be from the Bushe (underlying) and Poladpur (overlying) Formations. Using Google Earth's capability for viewing landscapes in tilted perspective employing topographic data, and integrating this with previous mapping studies (e.g., Beane et al., 1986), as well as recent field observations of a pervasive terrace level associated with the Poladpur-Bushe contact, Figure 3B shows a tilted regional view from a location between the towns of Mahabaleshwar and Mahad, identifying these terraces as "P/B," which are seen to dominate the landscape, along with the steep cliffs of the overlying Ambenali Formation in the background. We note also that the underlying Bushe exposures are permeated by a system of approximately NNW-trending fractures of some type, likely jointing, a regional feature within the Bushe that rarely penetrates into the overlying Poladpur and Ambenali Formations.

Figure 3C, also from Google Earth, shows an even more conspicuous pattern of fractures, again with NNW orientation, to the west of Pune, which are seen to not penetrate into the overlying Poladpur Formation. Furthermore, Figure 3D shows a close-up image of the fault in the upper left-hand corner of Figure 3C, which is a prominent feature within the Bushe Formation, but which clearly does not penetrate the overlying Poladpur lava flows. The fault in particular is classical field evidence for a disconformity, although further field work will be needed to be certain that the fracture array is not somehow lithologically controlled.

Taken together with the geochemical stratigraphic evidence from Beane et al. (1986), this new evidence for extensive jointing and a prominent fault that are so conspicuously truncated makes a strong case for a major disconformity, or volcanic hiatus, at the Poladpur-Bushe contact. Although these investigations are preliminary, they strongly suggest that the onset of the huge Wai Subgroup flows of the Poladpur, Ambenali, and Mahabaleshwar Formations may have occurred somewhat suddenly after a hiatus following the eruption of the Bushe Formation. We know of no similar evidence—geomorphologic, tectonic, or geochemical—for “missing time” anywhere between the Jawhar and Bushe Formations.

Reassessment of the Cretaceous-Paleogene–Chicxulub–Deccan “Coincidence”

These new inferences regarding the timing of the various formations within the Deccan Basalt Group lead naturally to a critical re-examination of the probabilistic argument for the Cretaceous-Paleogene–Chicxulub–Deccan “coincidence” outlined at the end of the introduction to this paper. If there was a time gap between the eruption of the Lonavala and Wai Subgroups (Poladpur-Bushe contact), and if we follow previous workers in inferring that the onset of the Wai Subgroup coincided closely with Cretaceous-Paleogene time, then it is apparent that this troubling “coincidence” is much more severe than previously thought. A sudden outburst of Deccan eruptions occurring within ~100,000 yr or less of Cretaceous-Paleogene time and accounting for >70% of the Deccan main-phase eruptions would seem to have a miniscule chance of occurring at random, or one in ~100, as was noted in the Introduction. It is therefore reasonable to ask: Might this most voluminous phase of Deccan eruptions have been triggered by the Chicxulub impact?

VOLCANIC TRIGGERING BY THE CHICXULUB IMPACT

Observations of Triggering of Volcanic Eruptions by Earthquakes

Triggering of volcanic eruptions by earthquakes is now well documented (Linde and Sacks, 1998). About 0.4% of explosive volcanic eruptions occur within a few days of distant earthquakes, an eruption frequency that is ~10 times greater than background rates (Manga and Brodsky, 2006). Other geofluid systems also respond to earthquakes: Examples include geysers, mud volcanoes, water levels in wells, and discharge into streams (Manga et al., 2012). Figure 4 summarizes evidence for earthquake triggering of both magmatic and mud volcanoes as a function of earthquake moment magnitude M*w* and epicentral distance. We include mud volcanoes in this compilation because they are more numerous and there are more documented examples of triggered mud eruptions. Figure 4 shows that larger earthquakes trigger eruptions over greater distances.

The mechanisms by which earthquakes influence magmatic volcanoes and hydrologic systems are unknown, and debated, but owing to the large distances between the earthquakes and the sites where the responses occur, the dynamic stresses produced by passing seismic waves are usually implicated in the triggering mechanisms (Manga and Brodsky, 2006). The amount of seismic energy dissipated by seismic waves has thus been proposed as a measure that may be correlated with triggered eruptions (Wang and Manga, 2010):

$$\log_{10} H = 0.48 M_w - 0.33 \log_{10} E - 1.4, \qquad (1)$$

where E is the seismic energy density in J/m^3, H is epicentral distance in km, and Mw is the moment magnitude of the earthquake. There appears to be a threshold energy density of ~10^{-2}–10^{-1} J/m^3 for triggering the hydrologic (mud volcano) events, and a somewhat larger threshold of ~10^{-1}–10^{0} J/m^3 for triggering magmatic volcanoes.

Scaling the Effects of the Chicxulub Impact

The kinetic energy of the Chicxulub impact has been estimated at ~3×10^{23} Joules (Boslough et al., 1996). Estimates for the efficiency of conversion of impact energy into seismic waves range from 10^{-2} to 10^{-5} (Schultz and Gault, 1974; Shishkin, 2007), implying effective moment magnitudes for Chicxulub in the range M*w* ~9–11, although such extrapolations depend upon the seismic frequency. At the ~130° epicentral distance (~13,000 km) of the Deccan Traps from Chicxulub at 66 Ma, and at uppermost mantle depths, seismic motions would have been dominated by long-period Rayleigh (surface) waves. Calculations of seismic radiation from the Chicxulub impact (Meschede et al., 2011) suggest that peak stresses and strains of order 2–4 bars and 0.1–0.2 µstrains, respectively, would have occurred globally in the crust and upper mantle, assuming a seismic conversion efficiency of 10^{-4}, a commonly adopted value (Schultz and Gault, 1974). This implies energy densities of order 10^{-1}–10^{0} J/m^3, as shown in Figure 4, which, according to historical evidence, appear to be large enough to trigger volcanic eruptions worldwide. Evidence for strong seismic motions at great distances from Chicxulub also includes distal continental margin collapse events (liquefaction; Bralower et al., 1998; Klaus et al., 2000), also shown in Figure 4. These latter events are important because the largest tectonic earthquakes are known to cause liquefaction effects only up to ~500 km from the earthquake source, so that these margin-collapse events imply that Chicxulub dynamic stresses must have resulted from the equivalent of a M*w* > 10 event (Day and Maslin, 2005).

Physical Mechanisms of Triggering and Magmatic Response Time Scales

How seismic waves trigger volcanic eruptions is not well understood, but it must involve either increasing the hydraulic head driving upward magma flow, or increasing the effective permeability of the magmatic system, and hence the rate of magma flow. Proposed mechanisms include exsolution/growth/advection of gas bubbles (e.g., H_2O, CO_2, which are supercritical fluids at lower lithosphere pressures), overturn of magma chambers, failure/fracture of rock, unclogging of fluid pathways, or disruption of grain-grain contact within zones of partial melt (reviewed in Manga and Brodsky, 2006), each possibly increasing the permeability and hence the rate of magma flow beneath volcanic centers. In hydrological systems at distances more than a few lengths of the ruptured fault, earthquake-induced permeability changes exceeding two orders of magnitude have been inferred at the scale of sedimentary basins (Wang et al., 2013), although factors of a few or more are more typical (Elkhoury et al., 2006).

Owing to observation bias, the majority of documented responses to earthquakes are phenomena that occur in the shallowest crust, though the initiation of magmatic and mud volcanoes may occur at depths of many kilometers, sometimes >10 km. The effect of overburden pressure on processes leading to eruption depends upon the mechanism. Liquefaction is usually viewed by geotechnical engineers as a process limited to depths less than 10 m, because at greater depths the difference between hydrostatic and lithostatic pressures is too large to be overcome. However, in the deeper crust, compaction, dewatering, and sealing can all lead to fluid pressures being close to lithostatic, in which case only small changes in pore pressure can mobilize granular materials.

Mechanisms that involve changes in permeability should not be sensitive to ambient pressure. Stresses from the passage of seismic waves are probably too small to produce new fractures at depth. Oscillatory flows induced by time-varying strains can mobilize solid particles (e.g., Candela et al., 2014) or bubbles (e.g., Beresnev et al., 2011) blocking pores, in either case increasing the mobility of fluids. Indeed, permeability changes induced by earthquakes have been measured or inferred at depths up to many kilometers (e.g., Wang et al., 2013). It has also been documented experimentally that the temporary lowering of pressure during the passage of P and Rayleigh waves can nucleate CO_2 bubbles, leading to a net increase in pore pressure after the final passage of the waves (Crews and Cooper, 2014).

Long-period motions appear to be more effective than short-period waves in exciting hydrologic systems (Rudolph and Manga, 2012). However, it must be emphasized that the physical mechanisms involved in the triggering of volcanic eruptions by dynamic stresses from earthquakes have not yet been determined even for shallow magmatic systems, and therefore we simply do not know what mechanisms may apply at deep crustal and sublithospheric depths in the case of the Deccan Traps. Nevertheless, as we show later herein (and in more detail in the Appendix), some inferences may be made as to how the overall system might respond to a sudden increase in permeability, regardless of the physical cause(s).

We start by considering how Chicxulub may have affected the various regimes of the Deccan magmatic plumbing system (Fig. 5): Flood basalts at the surface (regime 1) are fed by crustal dike and sill systems (regime 2), which are in turn fed by partial melting of the sublithospheric mantle within the hot, rising mantle plume head (regime 5), with porous melt migration coalescing into channelized flow through the overlying intact lithosphere (regime 4). Petrologic considerations (Cox, 1980), seismic imaging (Ridley and Richards, 2010; Richards et al., 2013), and thermodynamic models for melt equilibrium (Farnetani et al., 1996; Karlstrom and Richards, 2011; Richards et al., 2013) suggest a mediating zone (regime 3) at the crust-mantle boundary (Moho). Here, ultramafic mantle-derived melts collect in a density trap as laterally extensive magma chambers, undergoing crystal fractionation of Fe- and Mg-rich minerals (olivine and pyroxene) while lower-density, eruptible basaltic liquid is evolved.

Subsurface regimes 2–5 may be characterized by effective permeabilities, with the smallest permeabilities and longest magma transport times expected in the deeper regimes 4 and 5, where porous flow and compaction-dominated channelization of relatively low melt fractions (perhaps 0.2%–3.0%) occur within a porous matrix of hot mantle rock. Magma traverses regimes 1–3 relatively quickly, on time scales on the order of one year to a several thousand years. By contrast, flow within regime 5 would likely respond on time scales of thousands to hundreds of thousands of years.

Figure 6 illustrates qualitatively the response of a continental flood basalt magmatic system to impact-generated seismic stresses. (See Appendix for more detailed physics-based models.) The effective permeability within regime 5 increases almost instantaneously, with flow responding on a time scale (τt) set by the spatial distribution and magnitude of permeability change. This flow response is followed by a recovery period (τp) for permeability, and the two time scales may be similar if permeability recovery tracks magma transport. It is also possible that the onset of viscoelastic relaxation of stresses as newly injected magma heats the lower crust could impose an eruption shutoff time scale on the order of $\sim 10^5$–10^6 yr (Karlstrom and Richards, 2011), independent of the time scale for mantle permeability recovery.

We estimate τt by considering a generalized diffusive process of melt migration, wherein the relevant hydraulic diffusivity *D* may be written as

$$D = \kappa / (\mu * \beta_{eff}), \qquad (2)$$

where κ is a characteristic permeability, μ is the viscosity of ultramafic magma, and β_{eff} is the effective compressibility of the partially molten rock. Reasonable values for μ and β_{eff} are 0.1–1.0 Pa-s and 10^{-10} Pa^{-1}, respectively. Values for κ are less well constrained, but estimates range from 10^{-15} το 10^{-10} m^2 for partially molten mantle peridotite, depending upon the local melt fraction (Connolly et al., 2009; Sundberg et al., 2010). The largest volume of the plume head is likely characterized by melt fractions less than 3%, which would imply permeability values of order 10^{-11}–10^{-14} m^2.

The relevant length scale *L* for magma transport in regime 5 may be limited by matrix compaction, in which case a compaction length scale (see Appendix) of order 10 km or less may apply. By contrast, a practical estimate for the required length scale for magma extraction can be obtained by calculating the mantle source volume required to produce the ~500,000 km^3 volume of the Wai Subgroup lava flows. Assuming an average 2%–3% partial melt content (e.g., Villagomez et al., 2014) in the plume head (Fig. 3), we obtain a length scale of order *L* = 300–500 km, comparable to the size of the plume head itself. Estimating the magma transport time in response to a sudden permeability increase to be τ*t* ~ *L2/D*, we obtain a large range of time scales for magma flow adjustment in regime 5 of τ*t* ~ 10^3–10^8 yr, with 10^4–10^6 yr representing the likely midrange of values. Within these bounds, anomalously large Deccan eruptions, triggered by perturbation of the deep magmatic plumbing system due to seismic shaking from Chicxulub, could plausibly have occurred over a time scale on the order of ~100,000 yr, consistent with the time scale inferred stratigraphically for the huge Poladpur, Ambenali, and Mahabaleshwar flows. A more detailed mathematical model for these processes is given in the Appendix.

INDEPENDENT TECTONIC AND GEOCHEMICAL CONSTRAINTS

In the previous sections, we presented data and observations that suggest that the huge Wai Subgroup eruptions, which account for most of the Deccan flood basalt volume, may have occurred immediately after the Chicxulub impact and the Cretaceous-Paleogene extinction boundary. In the preceding section and in the Appendix, we developed a plausibility argument and generalized model for triggering of these large eruptions by the Chicxulub impact. In this section, we review three additional lines of evidence that were not part of the original motivation for this hypothesis, and therefore serve as independent tests for the volcanic triggering hypothesis.

Tectonic Conditions Inferred from Dike Orientations

In a recent study, Vanderkluysen et al. (2011) correlated mapped dike swarms with the various Deccan formations largely on the basis of the geochemical fingerprints established from previous work (Beane et al., 1986). In this work, it was possible to identify groups of Deccan formations for which presumed feeder dikes had distinct orientation characteristics, which can in turn be related to differing tectonic stress regimes associated with their emplacement. These are important observations, since the role of lithospheric extension in the generation of continental flood basalts is a subject of considerable debate (White and McKenzie, 1989; Richards et al., 1989; Campbell and Griffiths, 1990; Hooper, 1990, 1999), with possible bearing on the cause of the Wai Subgroup outburst.

In Figure 1A, column v, we summarize the main results from Vanderkluysen et al. (2011). The E-W–oriented Narmada-Tapi dike swarm (indicative of dominant N-S extensional stresses), to the north of the main Western Ghats Province, may have fed some of the lowermost Narmada and Rajpipla Formations, as well as some flows of the Kalsubai and Lonavala Subgroups in the lower and middle parts of the Deccan stratigraphy. The Kalsubai Subgroup formations are also associated with the largely N-S dike swarms in the Nasik-Pune region, indicating a shift to E-W extensional stresses. In striking contrast, the Wai Subgroup flows are associated with Nasik-Pune dike swarms that lack any consistent directional orientation, suggesting that imposed regional tectonic extensional stresses played little role in their genesis. This latter observation appears to negate any role for accelerated lithospheric extension in the increased production of magma from the Deccan plume head. Instead, we suggest that a surge in magma influx, leading to increased magma overpressure and buoyancy, may have driven the large Wai Subgroup eruptions, consistent with sudden mobilization of magma from the underlying mantle (regimes 4 and 5) as hypothesized in the previous section.

Mantle versus Crustal Signatures from Isotope Geochemistry of Deccan Basalts

A second test of our hypothesis comes from existing neodymium, strontium, and lead isotopic data (as well as other geochemical data), which show that the Wai Subgroup flows have a stronger mantle signature than the underlying Lonavala and Kalsubai flows, which show significant degrees of crustal contamination (Mahoney et al., 1982; Cox and Hawkesworth, 1985; Lightfoot and Hawkesworth, 1988; Lightfoot et al., 1990; Peng et al., 1994). Figure 1A, column iv, shows the ranges of ε_{Nd} values reported for the Deccan Traps, as summarized by Vanderkluysen et al. (2011). The Poladpur, Ambenali, and Mahabaleshwar Formations are characterized by higher (positive) ε_{Nd} values, indicating relatively uncontaminated mantle signatures, whereas the underlying formations (Bushe and below) generally exhibit lower ε_{Nd} values, indicating high degrees of crustal contamination/assimilation. (We note that the huge Ambenali Formation flows yield the most consistent mantle-like isotopic signatures.) These results are also consistent with high-TiO_2 values and other major- and trace-element signatures of the Wai Subgroup relative to underlying units (Beane et al., 1986), indicating that relatively uncontaminated mantle melts reached the surface in this period of time. The most straightforward interpretation of these data as a whole is that the Kalsubai and Lonavala Subgroups were formed by relatively small batches of magma that interacted strongly with the crust prior to eruption. By contrast, the Wai Subgroup flows must have been fed by enormous magma chambers in which relatively little crustal assimilation occurred before eruption. These observations suggest that the Wai Subgroup magmas were mobilized rapidly out of the mantle with little time for crustal interaction, consistent with the seismic triggering model, and also consistent with the stress regime implied by the associated randomly oriented Nasik-Pune dike swarms, as discussed earlier herein.

Rare Earth Element Constraints on Lithospheric Thinning

The amount of lithospheric extension that accompanies the eruption of flood basalts may be constrained by their rare earth element ratios (White and McKenzie, 1995). The most diagnostic tracers are the middle and heavy rare earth elements (MREEs and HREEs) because they are strongly influenced by the presence of garnet, which at temperatures appropriate to a mantle plume head is only stable at pressures >2.5 GPa (Klemme and O'Neill, 2000), equivalent to a depth of ~80 km. Unlike other mantle phases, garnet has a particularly strong affinity for the HREEs, and adiabatic decompression melting at depths >~80 km generates melts with relatively high MREE/HREE ratios (e.g., Sm/Yb). At shallower depths in the mantle, garnet is replaced by spinel, which has similar partition coefficients for all of the REEs, exerting a negligible effect on MREE/HREE ratios. Mantle melts generated beneath continental lithosphere of decreasing thickness are therefore characterized by progressively lower MREE/HREE ratios.

Figure 7A shows representative chondrite-normalized REE patterns for basalts from various formations in the Deccan continental flood basalt pile (see GSA Data Repository material[1]). Most notably, lavas from the base of the Deccan, i.e., from the lower part of the Kalsubai Subgroup at Narmada, have MREE/HREE ratios that are slightly elevated compared to those of basalts from higher up the succession. The lack of variation in MREE/HREE ratios of basalts from the upper part of the Kalsubai, Lonavala, and Wai Subgroups is striking. We used the REE inversion modeling procedure of McKenzie and O'Nions (1991; see GSA Data Repository material [see footnote 1]) to constrain how the degree of melting in the plume head would have varied with depth during the generation of Deccan basalts, and we used this information to estimate the shallowest depth at the top of the melting regime, i.e., the base of the rigid lithosphere. Our findings show that the base of the lithospheric lid resided at a similar depth (~60 km) immediately before, during, and after the eruption of the Ambenali Formation, and therefore that the Wai Subgroup lava pulse was not associated with a major lithospheric extension or delamination event. This contrasts with findings from other continental flood basalt provinces, such as the North Atlantic (Kerr, 1994), Paraná-Etendeka (Thompson et al., 2001), and Siberia (Sobolev et al., 2011; Elkins-Tanton and Hager, 2000). As for the dike orientation and isotopic data, these results on MREE/HREE abundances are consistent with the impact triggering hypothesis for the Wai Subgroup outburst, since this mechanism requires no lithospheric thinning, and they also seem to preclude alternative mechanisms involving a sudden pulse of tectonic extension or lithospheric instability/delamination.

DISCUSSION

The observations and data we have summarized here suggest that >70% of the main-stage of the Deccan flood basalt volume, contained in the Poladpur, Ambenali, and Mahabaleshwar Formations of the Wai Subgroup, was erupted during a relatively brief time interval on the order of one to several hundred thousand years at approximately the time of the Cretaceous-Paleogene mass extinction and the Chicxulub impact. We also suggest that a hiatus may have elapsed between the eruption of the Bushe Formation and the overlying Wai Subgroup formations, signaling a sudden pulse of enormous magma production from the Deccan plume head beneath the Indian subcontinent. We have further hypothesized that the Chicxulub impact may have led to this pulse by causing a transient increase in the effective permeability of the mantle magmatic system within the plume head.

The additional evidence we have reviewed from feeder-dike orientations and isotope and trace-element geochemistry supports this hypothesis and appears to exclude alternative mechanisms that invoke lithospheric thinning due to either tectonic extension or convective instability of the lithosphere. The only other plausible mechanism might be a sudden flow pulse within the Deccan mantle plume itself, which, like lithospheric thinning mechanisms, would require an appeal to extraordinary coincidence between such an event and the Cretaceous-Paleogene and Chicxulub events. Our hypothesis for impact triggering, by contrast, requires no such coincidence, but rather relies upon calculations that show that the dynamic stresses from Chicxulub were probably sufficient to cause volcanic eruptions worldwide.

Regarding the latter point, we have not yet carried out a global study of the possibility that other volcanic systems were triggered at Cretaceous-Paleogene time, although there are some reports that suggest this possibility (e.g., Shipboard Scientific Party, 1989; Busby et al., 2002). In most cases, it would be very difficult to determine whether ordinary volcanoes were triggered right at Cretaceous-Paleogene time, since their eruption repeat intervals are typically fairly short. Also, the occurrence of well-studied Cretaceous-Paleogene boundary sections in nontectonic sedimentary sections is rare (Claeys et al., 2002), so it would be difficult to discern normal-size eruptions occurring at exactly Cretaceous-Paleogene time. Nevertheless, pursuing this question is clearly worthwhile, perhaps particularly in regard to the possibility of global eruptions along the mid-ocean-ridge system, which might generate some signal in ocean trace-element or isotope chemistry.

Another question, perhaps relevant as well to the onset of main-phase flood basalt events, is whether dynamic stresses from a “local” great earthquake might also trigger anomalously large eruptive outbursts. Since the largest tectonic earthquakes occur at subduction zones and are not thought to exceed M*w* ~9.5, the fact that the subduction zone nearest the Deccan Traps at ca. 66 Ma was at least several thousand kilometers distant to the north (Chatterjee et al., 2013) suggests that the Deccan system may not have been severely perturbed by such events. However, the Columbia River Basalts were erupted much closer to the proto–Cascadia subduction zone, and hence one might wonder whether the enormous Grande Ronde flows, which account for ~80% of these flood basalts (Self et al., 2014), might have been induced by a nearby great subduction zone event. There are also reports linking the Central Atlantic magmatic province, the Triassic-Jurassic boundary, a possible impact event, and extensive "seismite" (earthquake-induced liquefaction) deposits (e.g., Simms, 2006; Lindstrom et al., 2015), suggesting possible linkages similar to those discussed here. We therefore suggest that the triggering of individual large eruptive events within flood basalt eruptions might be due to dynamic stresses generated by either impact events or large local earthquakes.

The greatest immediate need for future work is for more extensive high-precision radioisotopic dating of the Deccan formations. Such data could address not only the hypothesis we have put forward, but could also help to resolve other important questions regarding the Deccan Traps eruptions and their possible relation to climate change and mass extinction. We have recently undertaken new field sampling in the Western Ghats region to this end and hope thus to further clarify these issues.

If future geochronologic data confirm that the huge Wai Subgroup eruptions coincided closely in time with the Chicxulub impact, additional investigation of possible causal links and environmental consequences (including the possible consequences of triggered volcanic eruptions worldwide) would be warranted. On the other hand, if the Wai Subgroup eruptions were shown to have occurred well before or after Chicxulub/Cretaceous-Paleogene time, then the Deccan continental flood basalts may have played relatively little role in the extinctions. Thus, more precise radioisotopic dating and better biostratigraphic constraints on the largest main-stage flows within the Deccan Traps could reveal first-order answers to one of the great geological quandaries of our time.

CONCLUSIONS

The Cretaceous-Paleogene mass extinction, the Chicxulub impact, and the enormous Wai Subgroup lava flows of the Deccan Traps continental flood basalts appear to have occurred very close together in time, perhaps following a hiatus of uncertain duration in the main-stage Deccan eruptions. Evidence from the feeder dike systems and from isotope and trace-element geochemistry suggests that the Wai Subgroup flows were fed by a pulse of mantle plume-head–derived magma that interacted little with overlying crust, and also that lithospheric thinning did not cause this singular event.

Following model calculations suggesting that dynamic stresses (Rayleigh waves) from the Chicxulub impact may have been sufficiently large to have triggered volcanic eruptions worldwide, we hypothesize that the Wai Subgroup eruptions within the Deccan Traps may have been triggered by the impact. Although the physical mechanisms for how earthquakes trigger volcanic eruptions are not known, a transient permeability increase in the partially molten mantle underlying the Deccan province could explain the timing of the Wai Subgroup eruptions. Precise dating of the Deccan Traps formations offers the means by which to test the timing relationships implied by the impact triggering hypothesis. Better biostratigraphic constraints on the timing of the Wai Subgroup eruptions, as well as investigation of the possibility that the Chicxulub impact may have triggered volcanic eruptions worldwide, could also lead to a more refined understanding of the relative roles of the impact and volcanism in the Cretaceous-Paleogene mass extinction event.

Erratum

Triggering of the largest Deccan eruptions by the Chicxulub impact

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[1]GSA Data Repository item 2015164, rare-earth element (REE) data used, and a description of the procedures followed, in the REE inversion results summarized in Figure 7 of the paper, is available at http://www.geosociety.org/pubs/ft2015.htm or by request to editing@geosociety.org.

References

Alvarez, W., 2003, Comparing the evidence relevant to impact and flood basalt at times of major mass extinctions: Astrobiology, v. 3, p. 153–161, doi:10.1089 /153110703321632480.

Alvarez, L.W., Alvarez, W., Asaro, F., and Michel, H.V., 1980, Extraterrestrial cause for the Cretaceous-Tertiary extinction: Science, v. 208, p. 1095–1108, doi:10.1126 /science.208.4448.1095.

Bajpai, S., and Prasae, G.V.R., 2000, Cretaceous age for Ir-rich Deccan intertrappean deposits: Palaeontological evidence from Anjar, western India: Journal of the Geological Society of London, v. 157, p. 257–260, doi: 10.1144/ jgs.157.2.257.

Baksi, A.J., Byerly, G.R., Chan, L.-H., and Farrar, E., 1994, Intracanyon flows in the Deccan province, India? Case history of the Rajamundry Traps: Geology, v. 22, p. 605–608.

Barrera, E., and Savin, S.M., 1999, Evolution of late Campanian–Maastrichtian climates and oceans, in Barrera, E., and Johnson, C.C., eds., Evolution of the Cretaceous Ocean-Climate System: Geological Society of America Special Paper 332, p. 245–282, doi:10.1130/0-8137-2332 -9.245.

Basu, A.R., Renne, P.R., DasGupta, D., Teichmann, F., and Poreda, R.J., 1993, Early and late alkali igneous pulses and a high-3 He origin for the Deccan flood basalts: Science, v. 261, p. 902–906, doi:10.1126/science.261 .5123.902.

Beane, J.E., Turner, C.A., Hooper, P.R., Subbarao, K.V., and Walsh, J.N., 1986, Stratigraphy, composition and form of the Deccan basalts, Western Ghats, India: Bulletin of Volcanology, v. 48, p. 61–83, doi:10.1007 /BF01073513.

Beresnev, I.A., Gaul, W., and Vigil, R.D., 2011, Direct pore-level observation of permeability increase in twophase flow by shaking: Geophysical Research Letters, v. 38, L20302, doi:10.1029/2011GL048840.

Black, B.A., Lamarque, J.-F., Shields, C.A., Elkins-Tanton, L.T., and Kiehl, J.T., 2014, Acid rain and ozone depletion from pulsed Siberian Traps magmatism: Geology, v. 42, p. 67–70, doi:10.1130/G34875.1.

Boslough, M.B., Chael, E.P., Trucano, T.G., Crawford, D.A., and Campbell, D.L., 1996, Axial focusing of impact energy in the Earth's interior: A possible link to flood basalts and hotspots, in Ryder, G., Fastovsky, D.E., and Gartner, S., eds., The Cretaceous-Tertiary Event and Other Catastrophes in Earth History: Geological Society of America Special Paper 307, p. 817–840.

Bralower, T.J., Paull, C.K., and Leckie, R.M., 1998, The Cretaceous-Tertiary boundary cocktail: Chicxulub impact triggers margin collapse and extensive sediment gravity flows: Geology, v. 26, p. 331–334, doi:10.1130 /0091-7613(1998)0262.3.CO;2.

Busby, C.J., Yip, G., Blikra, L., and Renne, P., 2002, Coastal landsliding and catastrophic sedimentation triggered by Cretaceous-Tertiary bolide impact: A Pacific margin example?: Geology, v. 30, p. 687–690, doi:10.1130/0091-7613(2002)0302.0.CO;2.

Campbell, I., and Griffiths, R., 1990, Implication of mantle plume structure for the evolution of flood basalts: Earth and Planetary Science Letters, v. 99, p. 79–93, doi: 10.1016/0012-821X(90)90072-6.

Candela, T., Brodsky, E.E., Marone, C.J., and Elsworth, D., 2014, Laboratory evidence for particle mobilization as a mechanism for permeability enhancement via dynamic stressing: Earth and Planetary Science Letters, v. 392, p. 279–291, doi:10.1016/j.epsl.2014.02.025.

Chatterjee, S., Goswami, A., and Scotese, C.R., 2013, The longest voyage: Tectonic, magmatic, and paleoclimatic evolution of the Indian plate during its northward flight from Gondwana to Asia: Gondwana Research, v. 23, p. 238–267, doi:10.1016/j.gr.2012.07.001.

Chenet, A.L., Quidelleur, X., Fluteau, F., and Courtillot, V., 2007, 40K-40Ar dating of the main Deccan large igneous province: Further evidence of KTB age and short duration: Earth and Planetary Science Letters, v. 263, p. 1–15, doi:10.1016/j.epsl.2007.07.011.

Chenet, A.L., Fluteau, F., Courtillot, V., Gerard, M., and Subbarao, K.V., 2008, Determination of rapid Deccan eruptions across the KTB using paleomagnetic secular variation: II. Results from a 1200-m-thick section in the Mahabaleshwar escarpment: Journal of Geophysical Research, v. 113, B04101, 27 p.

Chenet, A.L., Courtillot, V., Fluteau, F., Gerard, M., Quidelleur, X., Khadri, S.F.R., Subbarao, K.V., and Thordarson, T., 2009, Determination of rapid Deccan eruptions across the Cretaceous-Tertiary boundary using paleomagnetic secular variation: II. Constraints from analysis of eight new sections and synthesis for a 3500-m-thick composite section: Journal of Geophysical Research, v. 114, p. B06103, doi:10.1029/2008JB005644.

Claeys, P., Kiessling, W., and Alvarez, W., 2002, Distribution of the ejecta at the Cretaceous-Tertiary boundary, in Koeberl, C., and MacLeod, K.G., eds., Catastrophic Events and Mass Extinction: Impacts and Beyond: Geological Society of America Special Paper 356, p. 55–69.

Connolly, J.A.D., Schmidt, M.W., Solferino, G., and Bagdassarov, N., 2009, Permeability of asthenospheric mantle and melt extraction rates at mid-ocean ridges: Nature, v. 462, p. 209–212, doi:10.1038/nature08517.

Courtillot, V.E., and Renne, P.R., 2003, On the ages of flood basalt events: Comptes Rendus de l'Académie des Sciences, v. 335, p. 113–140.

Courtillot, V.E., Feraud, G., Maluski, H., Vandamme, D., Moreau, M.G., and Besse, J., 1988, Deccan flood basalts and the Cretaceous/Tertiary boundary: Nature, v. 333, p. 843–846, doi:10.1038/333843a0.

Courtillot, V., Gallet, Y., Rocchia, R., Feraud, G., Robin, E., Hofmann, C., Bhandari, N., and Ghevariaya, Z.G., 2000, Cosmic markers, 40Ar/39Ar dating and paleomagnetism of the KT sections in the Anjar area of the Deccan large igneous province: Earth and Planetary Science Letters, v. 182, p. 137–156, doi:10.1016/S0012-821X(00)00238-7.

Cox, K., 1980, A model for flood basalt volcanism: Journal of Petrology, v. 21, p. 629–650, doi:10.1093/petrology /21.4.629.

Cox, K., and Hawkesworth, C.J., 1985, Chemical stratigraphy of the Deccan Traps at Mahabaleshwar, Western Ghats, India, with implications for open system magmatic processes: Journal of Petrology, v. 26, p. 355– 377, doi:10.1093/petrology/26.2.355.

Crews, J.B., and Cooper, C.A., 2014, Experimental evidence for seismically initiated gas bubble nucleation and growth in groundwater as a mechanism for co-seismic borehole water level rise and remotely triggered seismicity: Journal of Geophysical Research–Solid Earth, v. 119, p. 7079–7091, doi:10.1002/2014JB011398.

Day, S., and Maslin, M., 2005, Linking large impacts, gas hydrates, and carbon isotope excursions through widespread sediment liquefaction and continental slope failure: The example of the K-T boundary event, in Kenkmann, T., Hor, F., and Deutsch, A., eds., Large Meteorite Impacts III: Geological Society of America Special Paper 384, p. 239–258.

Devey, C.W., and Lightfoot, P.W., 1986, Volcanological and tectonic control of stratigraphy and structure in the western Deccan Traps: Bulletin of Volcanology, v. 48, p. 195–207, doi:10.1007/BF01087674.

Duncan, R.A., and Pyle, D.G., 1988, Rapid eruption of the Deccan flood basalts at the Cretaceous/Tertiary boundary: Nature, v. 333, p. 841–843, doi:10.1038/333841a0.

Elkhoury, J.E., Brodsky, E.E., and Agnew, D.C., 2006, Seismic waves increase permeability: Nature, v. 441, p. 1135–1138, doi:10.1038/nature04798.

Elkins-Tanton, L.T., and Hager, B.H., 2000, Melt intrusion as a trigger for lithospheric foundering and the eruption of the Siberian flood basalts: Geophysical Research Letters, v. 27, p. 3937–3940, doi:10.1029/2000GL011751.

Farnetani, C., and Richards, M., 1994, Numerical investigations of the mantle plume initiation model for flood basalt events: Journal of Geophysical Research, v. 99, p. 13,813–13,833, doi:10.1029/94JB00649.

Farnetani, C., Richards, M.A., and Ghiorso, M.S., 1996, Petrological models of magma evolution and deep crustal structure beneath hotspots and flood basalt provinces: Earth and Planetary Science Letters, v. 143, p. 81–94, doi:10.1016/0012-821X(96)00138-0.

Gibson, S.A., 2000, Ferropicrites: Geochemical evidence for Fe-rich streaks in upwelling mantle plumes: Earth and Planetary Science Letters, v. 174, p. 355–374, doi: 10.1016/S0012-821X(99)00274-5.

Hildebrand, A.R., Penfield, G.T., Kring, D.A., Pilkington, M., Camargo, A., Jacobsen, S.B., and Boynton, W.V., 1991, Chicxulub crater: A possible Cretaceous/Tertiary boundary impact crater on the Yucatán Peninsula, Mexico: Geology, v. 19, p. 867–871, doi:10.1130/0091 -7613(1991)0192.3.CO;2.

Hofmann, C., Feraud, G., and Courtillot, V., 2000, 40Ar/39Ar dating of mineral separates and whole rocks from the Western Ghats lava pile: Further constraints on duration and age of the Deccan Traps: Earth and Planetary Science Letters, v. 180, p. 13–27, doi:10.1016/S0012 -821X(00)00159-X.

Hooper, P.R., 1990, The timing of crustal extension and the eruption of continental flood basalts: Nature, v. 345, p. 246–249, doi:10.1038/345246a0.

Hooper, P.R., 1999, The Deccan Traps, in Subbarao, K.V., ed., Deccan Volcanic Province: Geological Society of India Memoir 43, p. 153–165.

Hooper, P., Widdowson, M., and Kelly, S., 2010, Tectonic setting and timing of the final Deccan flood basalt eruptions: Geology, v. 38, p. 839–842.

Ivanov, B.A., and Melosh, H.J., 2003, Impacts do not initiate volcanic eruptions: Eruptions near the crater: Geology, v. 31, p. 869–872.

Jay, A.E., and Widdowson, M., 2008, Stratigraphy, structure, and volcanology of the SE Deccan continental flood basalt province: Implications for eruptive extent and volumes: Journal of the Geological Society of London, v. 165, p. 177–188, doi:10.1144/0016-76492006-062.

Karlstrom, L., and Richards, M.A., 2011, On the evolution of large ultramafic magma chambers and timescales for flood basalt eruptions: Journal of Geophysical Research, v. 116, B08216, doi:10.1029/2010JB008159.

Keller, G., Adatte, T., Bajpai, S., Mahabey, D.M., Widdowson, M., Khosla, A., Sharma, R., Khosla, S.C., Gertsch, B., Fleitmann, D., and Sahni, A., 2009, K-T transition in Deccan Traps and intertrappean beds in central India mark major marine seaway across India: Earth and Planetary Science Letters, v. 282, p. 10–23, doi:10.1016/j.epsl.2009.02.016.

Keller, G., Adatte, T., Bhowmick, P.K., Upadhyay, H., Dave, A., Reddy, A.N., and Jaiprakash, B.C., 2012, Nature and timing of extinctions in Cretaceous-Tertiary planktic foraminifera preserved in Deccan intertrappean sediments of the Krishna-Godavari Basin, India: Earth and Planetary Science Letters, v. 341–344, p. 211–221, doi:10.1016/j.epsl.2012.06.021.

Kerr, A.C., 1994, Lithospheric thinning during the evolution of continental large igneous provinces: A case study from the North Atlantic Tertiary province: Geology, v. 22, p. 1027–1030, doi:10.1130/0091-7613(1994)022 2.3.CO;2.

Klaus, A., Norris, R.D., Kroon, D., and Smit, J., 2000, Impact-induced mass wasting at the K-T boundary: Blake Nose, western North Atlantic: Geology, v. 28, p. 319– 322, doi:10.1130/0091-7613(2000)282.0.CO;2.

Klemme, S., and O'Neill, H.S.C., 2000, The near-solidus transition from garnet lherzolite to spinel lherzolite: Contributions to Mineralogy and Petrology, v. 138, p. 237–248, doi:10.1007/s004100050560.

Knight, K.B., Renne, P.R., Halkett, A., and White, N., 2003, 40Ar/39Ar dating of the Rajamundry Traps, eastern India, and their relationship to the Deccan Traps: Earth and Planetary Science Letters, v. 208, p. 85–99, doi:10.1016/S0012-821X(02)01154-8.

Kumar, S., Singh, M.P., and Mohabey, D.M., 1999, Lameta and Bagh Beds: Uttar Pradesh, India, Paleontological Society of India, Field Guide, 48 p.
Leitch, A.M., and Davies, G.F., 2001, Mantle plumes and flood basalts: Enhanced melting from plume ascent and an eclogite component: Journal of Geophysical Research, v. 106, p. 2047–2059, doi:10.1029/2000JB900307.

Lemarchand, N., and Grasso, J.-R., 2007, Interactions between earthquakes and volcano activity: Geophysical Research Letters, v. 34, L24303, doi:10.1029/2007GL031438.

Li, L., and Keller, G., 1998, Maastrichtian climate, productivity and faunal turnovers in planktic foraminifera in South Atlantic DSDP Sites 525A and 21: Marine Micropaleontology, v. 33, p. 55–86.

Lightfoot, P.C., and Hawkesworth, C.J., 1988, Origin of Deccan Trap lavas: Evidence from combined trace element and Sr-, Nd- and Pb-isotope studies: Earth and Planetary Science Letters, v. 91, p. 89–104, doi:10.1016/0012-821X (88)90153-7.

Lightfoot, P.C., Hawkesworth, C.J., Devey, C.W., Rogers, N.W., and Van Calsteren, P.W.C., 1990, Source and differentiation of Deccan Trap lavas: Implications of geochemical and mineral chemical variations: Journal of Petrology, v. 31, p. 1165–1200, doi:10.1093/petrology/31.5.1165.

Linde, A.T., and Sacks, I.S., 1998, Triggering of volcanic eruptions: Nature, v. 395, p. 888–890, doi:10.1038/27650.

Lindstrom, S., Pedersen, G.K., van d Schootbrugge, B., Hansen, K.H., Kuhlmann, N., Thein, J., Johansson, L., Petersen, H.I., Alwmark, C., Dybkjaer, K., Weibel, R., Erlstrom, M., Hielsen, L.H., Oschmann, W., and Tegner, C., 2015, Intense and widespread seismicity during the end-Triassic mass extinction due to emplacement of a large igneous province: Geology, doi:10.1130/G36444.1.

Mahoney, J.J., Macdougall, J.D., Lugmair, J.W., Murali, A.V., Sankar Das, M., and Gopalan, K., 1982, Origin of the Deccan Trap flows at Mahabaleshwar inferred from Nd and Sr isotopic and chemical evidence: Earth and Planetary Science Letters, v. 60, p. 47–60, doi:10.1016 /0012-821X(82)90019-X.

Manga, M., and Bonini, M., 2012, Large historical eruptions at subaerial mud volcanoes, Italy: Natural Hazards and Earth System Sciences, v. 12, p. 3377–3386, doi: 10.5194/nhess-12-3377-2012.

Manga, M., and Brodsky, E., 2006, Seismic triggering of eruptions in the far field: Volcanoes and geysers: Annual Review of Earth and Planetary Sciences, v. 34, p. 263– 291, doi:10.1146/annurev.earth.34.031405.125125.

Manga, M., Brumm, M., and Rudolph, M.L., 2009, Earthquake triggering of mud volcanos: A review: Marine and Petroleum Geology, v. 26, p. 1785–1798, doi:10.1016/j .marpetgeo.2009.01.019.

Manga, M., Beresnev, I., Brodsky, E.E., Elkhoury, J.E., Elsworth, D., Ingegritsen, S.E., Mays, D.C., and Wang, C.-Y., 2012, Changes in permeability caused by transient stresses: Field observations, experiments, and mechanisms: Reviews of Geophysics, v. 50, doi:10.1029 /2011RG000382.

McDonough, W.F., and Sun, S.S., 1995, The composition of the Earth: Chemical Geology, v. 120, p. 223–253, doi: 10.1016/0009-2541(94)00140-4.

McKenzie, D., and O'Nions, R.K., 1991, Partial melt distributions from inversion of rare earth element concentrations: Journal of Petrology, v. 32, p. 1021–1091, doi: 10.1093/petrology/32.5.1021.

Meschede, M.A., Myhrvold, C.L., and Tromp, J., 2011, Antipodal focusing of seismic waves due to large meteorite impacts on Earth: Geophysical Journal International, v. 187, p. 529–537, doi:10.1111/j.1365-246X.2011.05170.x.

Morgan, W.J., 1981, Hotspot tracks and the opening of the Atlantic and Indian Oceans, in Emiliani, C., ed., The Sea, Volume 7: New York, Wiley Interscience, p. 443–487.

Ogg, J.G., 2012, The geomagnetic polarity time scale, in Gradstein, F.M., et al., eds., The Geomagnetic Time Scale 2012: Oxford, Elsevier, 85 p.

Pande, K., 2002, Age and duration of the Deccan Traps, India: A review of radiometric and paleomagnetic constraints: Proceedings of the Indian Academy of Sciences, Earth and Planetary Sciences, v. 111, p. 115–123.

Pande, K., Pattanayak, S.K., Subbarao, K.V., Navaneethakrishnan, P., and Venkatesan, T.R., 2004, 40Ar-39Ar age of a lava flow from the Bhimashankar Formation, Giravali Ghat, Deccan Traps: Proceedings of the Indian Academy of Sciences, Earth and Planetary Sciences, v. 113, p. 755–758.

Peng, Z.X., Mahoney, J.J., Hooper, P., Harris, C., and Beane, J., 1994, A role for lower continental crust in flood basalt genesis? Isotopic and incompatible element study of the lower six formations of the Western Ghats Traps: Geochimica et Cosmochimica Acta, v. 58, p. 267–288, doi:10.1016/0016-7037(94)90464-2.

Peng, Z.X., Mahoney, J.J., Hooper, P.R., Macdougall, J.D., and Krishnamurthy, P., 1998, Basalts of the northeastern Deccan Traps, India: Isotopic and elemental geochemistry and relation to southwestern Deccan stratigraphy: Journal of Geophysical Research, v. 103, no. B12, p. 29,843–29,865, doi:10.1029/98JB01514.

Renne, P.R., Balco, G., Ludwig, K.R., Mundil, R., and Min, K., 2011, Response to the comment by W.H. Schwarz et al. on “Joint determination of 40K decay constants and 40Ar*/40K for the Fish Canyon sanidine standard, and improved accuracy for 40Ar/39Ar geochronology” by P.R. Renne et al. (2010): Geochimica et Cosmochimica Acta, v. 75: p. 5097–5100.

Renne, P.R., Deino, A.L., Hilgen, F.J., Kuiper, K.F., Mark, D.F., Mitchell, W.S., III, Morgan, L.E., Mundil, R., and Smit, J., 2013, Time scales of critical events around the Cretaceous-Paleogene boundary: Science, v. 339, p. 684–687, doi:10.1126/science.1230492.

Richards, M.A., Duncan, R.A., and Courtillot, V.E., 1989, Flood basalts and hotspot tracks: Plume heads and tails: Science, v. 246, p. 103–107, doi:10.1126/science .246.4926.103.

Richards, M.A., Contreras-Reyes, E., Lithgow-Bertelloni, C., Ghiorso, M., and Stixrude, L., 2013, Petrological interpretation of deep crustal intrusive bodies beneath oceanic hotspot provinces: Geochemistry Geophysics Geosystems, v. 14, p. 604–619, doi:10.1029/2012GC004448.

Ridley, V., and Richards, M.A., 2010, Deep crustal structure beneath large igneous provinces and the petrologic evolution of flood basalts: Geochemistry Geophysics Geosystems, v. 11, GC004448, doi:10.1029/2009GC002935.

Rudolph, M.I., and Manga, M., 2010, Mud volcano response to the 4 April 2010 El Mayor–Cucapah earthquake: Journal of Geophysical Research, v. 115, B12211, doi: 10.1029/2010JB007737.

Rudolph, M.I., and Manga, M., 2012, Frequency dependence of mud volcano response to earthquakes: Geophysical Research Letters, v. 39, L14303, doi:10.1029 /2012GL052383.

Saxena, R.K., and Misra, C.M., 1994, Time and duration of Deccan volcanism in the Razole area, Krishna-Godavari Basin, India: Current Science, v. 66, p. 73–76.

Schulte, P., and 40 others, 2010, The Chicxulub asteroid impact and mass extinction at the Cretaceous-Paleogene boundary: Science, v. 327, p. 1214–1218, doi:10.1126 /science.1177265.

Schultz, P.H., and Gault, D.E., 1974, Seismic effects from major basin formations on the Moon and Mercury: Earth, Moon, and Planets, v. 12, p. 159–177.

Self, S., Widdowson, M., Thordarson, T., and Jay, A.E., 2006, Volatile fluxes during flood basalt eruptions and potential effects on global environment: A Deccan perspective: Earth and Planetary Science Letters, v. 248, p. 518–532, doi:10.1016/j.epsl.2006.05.041.

Self, S., Jay, A.E., Widdowson, M., and Keszhelyi, L.P., 2008a, Correlation of the Deccan and Rajahmundry Trap lavas: Are these the longest and largest lava flows on Earth?: Journal of Volcanology and Geothermal Research, v. 172, p. 3–19, doi:10.1016/j.jvolgeores.2006.11.012.

Self, S., Blake, S., Sharma, K., Widdowson, M., and Sephton, S., 2008b, Sulphur and chlorine in Late Cretaceous Deccan magmas and eruptive gas release: Science, v. 319, p. 1654–1657.

Self, S., Schmidt, A., and Mather, T.A., 2014, Emplacement characteristics, timescales, and volatile release rates of continental flood basalt eruptions on Earth, in Keller, G., and Kerr, A., eds., Volcanism, Impacts, Mass Extinctions: Geological Society of America Special Paper 505, p. 319–337, doi:10.1130/2014.2505(16).

Shipboard Scientific Party, 1989, Ninetyeast Ridge summary, in Pierce, J., et al., Proceedings of the Ocean Drilling Project, Initial Reports, Volume 121: College Station, Texas, Ocean Drilling Program, p. 517–537.

Shishkin, N., 2007, Seismic efficiency of a contact explosion and a high-velocity impact: Journal of Applied Mechanics and Technical Physics, v. 48, p. 145–152, doi:10.1007/s10808-007-0019-6.

Simms, M.J., 2006, Uniquely extensive soft-sediment deformation in the Rhaetian of the UK: Evidence for earthquake or impact?: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 244, p. 407–423, doi:10.1016/j.palaeo.2006.06.037.

Smit, J., and Hertogen, J., 1980, An extraterrestrial event at the Cretaceous-Tertiary boundary: Nature, v. 285, p. 198–200, doi:10.1038/285198a0.

Sobolev, S.V., Sobolev, A.V., Kuzmin, D.V., Krivolutskaya, N.A., Petrunin, A.G., Arndt, N.T., Radko, V.A., and Vasiliev, Y.R., 2011, Linking mantle plumes, large igneous provinces and environmental catastrophes: Nature, v. 477, p. 312–316, doi:10.1038/nature10385.

Spiegelman, M., 1993, Flow in deformable porous media. Part 1: Simple analysis: Journal of Fluid Mechanics, v. 247, p. 17–38, doi:10.1017/S0022112093000369.

Sprain, C.J., Renne, P.R., Wilson, G.P., and Clemens, W.A., 2014, High-resolution chronostratigraphy of the terrestrial Cretaceous-Paleogene transition and recovery interval in the Hell Creek region, Montana: Geological Society of America Bulletin, v. 127, p. 393–409, doi: 10.1130/B31076.1.

Subbarao, K.V., Chandrasekharam, D., Navaneethakrishnan, P., and Hooper, P.R., 1994, Stratigraphy and structure of parts of the central Deccan basalt province: Eruptive models, in Subbarao, K.V., ed., Volcanism: New Delhi, Wiley Easter, p. 321–332.

Sundberg, M., Hirth, G., and Kelemen, P.B., 2010, Trapped melt in the Josephine Peridotite: Implications for the permeability and melt extraction in the upper mantle: Journal of Petrology, v. 51, p. 185–200, doi:10.1093/petrology/egp089.

Thompson, R.N., Gibson, S.A., Dickin, A.P., and Smith, P.M., 2001, Early Cretaceous basalt and picrite dykes of the southern Etendeka region, NW Namibia: Windows into the role of the Tristan mantle plume in Paraná–Etendeka magmatism: Journal of Petrology, v. 42, p. 2049–2081, doi:10.1093/ petrology/42.11.2049.

Vanderkluysen, L., Mahoney, J.J., Hooper, P.R., Sheth, H.C., and Ray, R., 2011, The feeder system of the Deccan Traps (India): Insights from dike geochemistry: Journal of Petrology, v. 52, p. 315–343, doi:10.1093 /petrology/egq082.

Venkatesan, T.R., Pande, K., and Gopalan, K., 1993, Did Deccan volcanism predate the Cretaceous/Tertiary transition?: Earth and Planetary Science Letters, v. 119, p. 181–189, doi:10.1016/0012-821X(93)90015-2.

Villagomez, D.R., Toomey, D.R., Geist, D.J., Hooft, E.E., and Solomon, S.C., 2014, Mantle flow and multistage melting beneath the Galápagos hotspot revealed by seismic imaging: Nature Geoscience, v. 7, p. 151–156, doi:10.1038/ngeo2062.

Wang, C.-Y., and Manga, M., 2010, Hydrologic responses to earthquakes: Geofluids, v. 10, p. 206–216.

Wang, C.-Y., Wang, L.-P., Manga, M., Wang, C.-H., and Chen, C.H., 2013, Basin-scale transport of heat and fluid induced by earthquakes: Geophysical Research Letters, v. 40, p. 3893–3897, doi:10.1002/grl.50738.

White, R.S., and McKenzie, D.P., 1989, Magmatism at rift zones: The generation of volcanic continental margins and flood basalts: Journal of Geophysical Research, v. 94, p. 7685–7729, doi:10.1029/JB094iB06p07685.

White, R.S., and McKenzie, D., 1995, Mantle plumes and flood basalts: Journal of Geophysical Research–Solid Earth, v.100, p. 17,543–17,585, doi:10.1029/95JB01585.

Wilf, P., Johnson, K.R., and Huber, B.T., 2003, Correlated terrestrial and marine evidence for global climate changes before mass extinction at the Cretaceous-Paleogene boundary: Proceedings of the National Academy of Sciences of the United States of America, v. 100, p. 599– 604, doi:10.1073/pnas.0234701100.

Williams, S.E., Muller, D., Landrebe, C.W., and Whittaker, J.M., 2012, An open-source software environment for visualizing and refining plate tectonic reconstructions using high-resolution geological and geophysical data sets: GSA Today, v. 22, no. 4, p. 4–9.

Wilson, G.P., 2005, Mammalian faunal dynamics during the last 1.8 million years of the Cretaceous in Garfield County, Montana: Journal of Mammalian Evolution, v. 12, p. 53–76, doi:10.1007/s10914-005-6943-4.

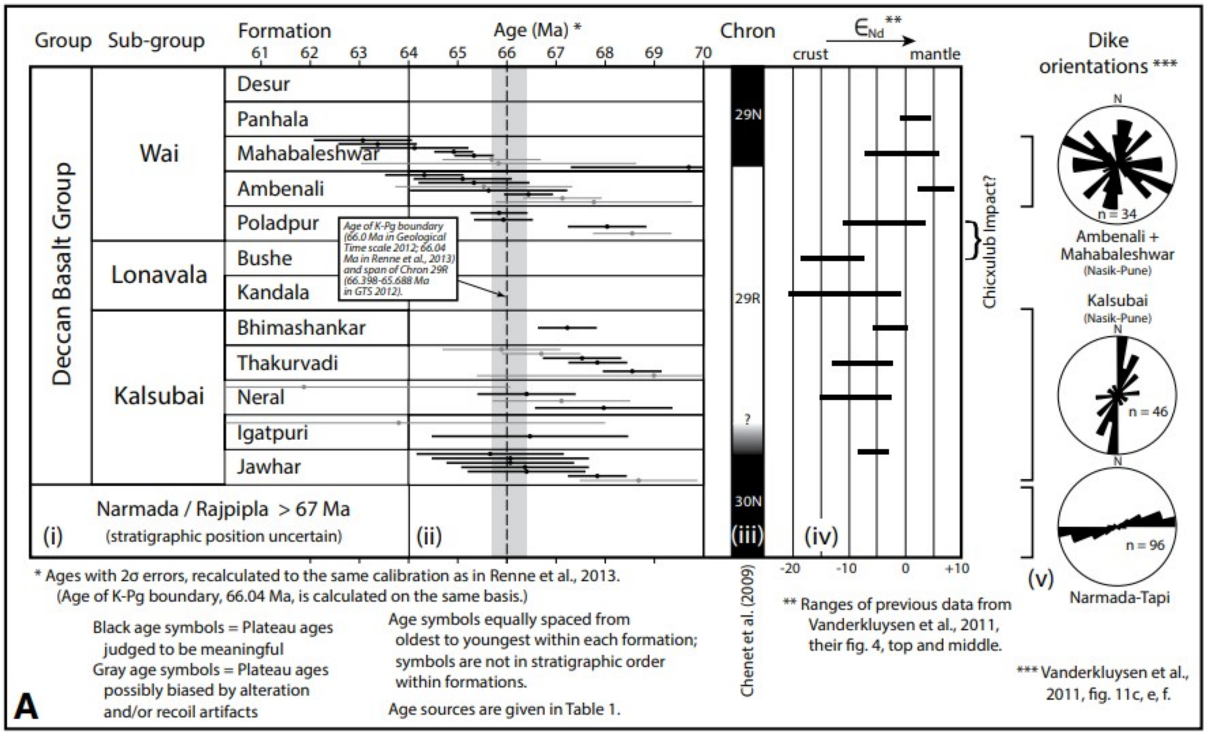


Figure 1. (A) (i) Deccan subgroups and formations (Chenet et al., 2008). Additional information includes (ii) radioisotopic ages (see Table 1 for sources) with 2σ uncertainties indicated by error bars, (iii) geomagnetic polarity assuming all reversed-polarity formations are within chron 29R, (iv) ϵ_{Nd} ranges (Vanderkluysen et al., 2011), and (v) orientations of associated dikes (Vanderkluysen et al., 2011). Relatively minor-volume Narmada/Rajpipla basalts occurred during earlier phases of Deccan volcanism along the east-west Narmada Rift, and these clearly predate the Chicxulub impact. Gray area in column (ii) indicates approximate time range of chron 29R. (*Continued on following page.*)

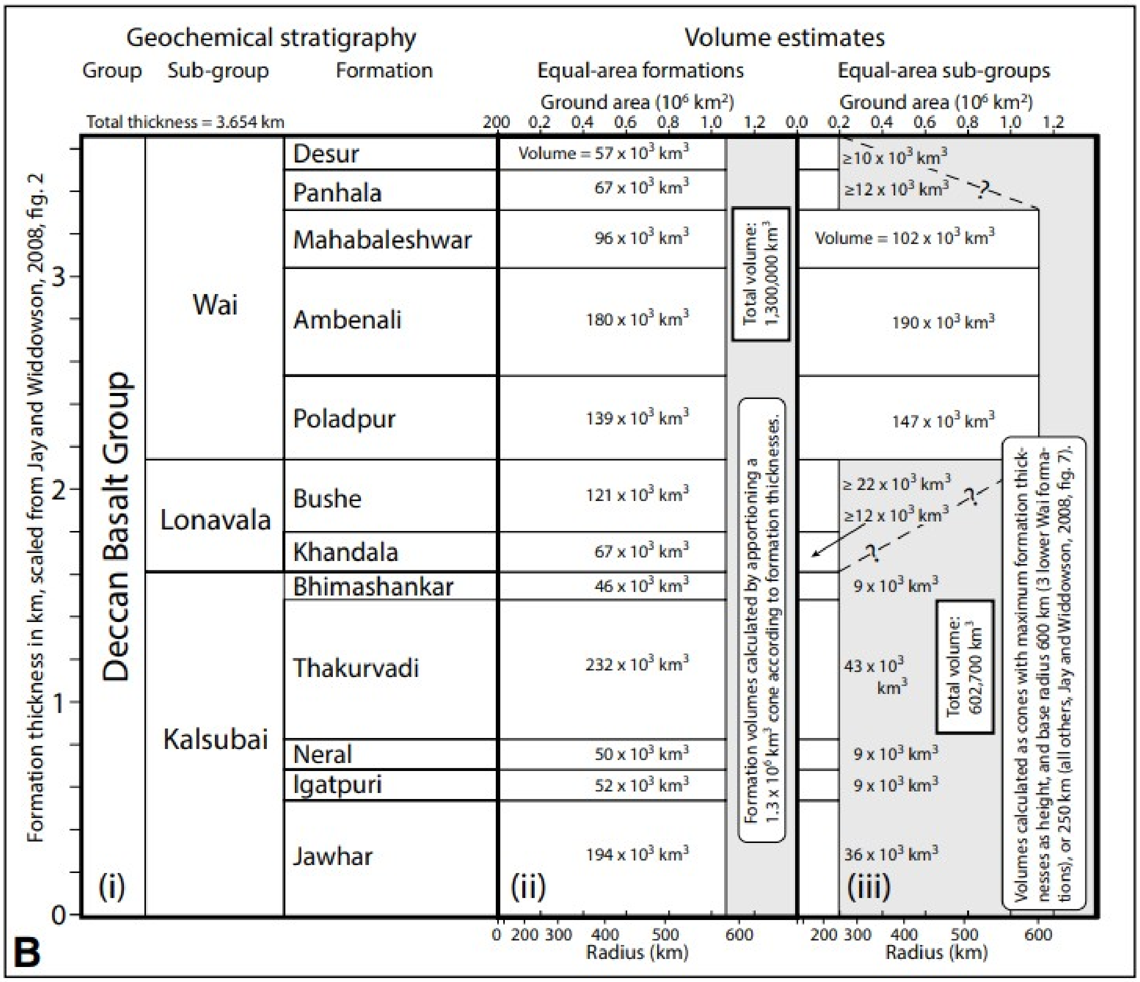


Figure 1 (*continued*). (B) (i) Maximum formation thicknesses. (ii) Formation volumes based on maximum formation thicknesses, assuming total volume is 1.3×10^6 km^3 (Jay and Widdowson, 2008) and that each formation extends over the entire footprint area of the Deccan Traps (Self et al., 2006). (iii) Volumes based on the assumption that the Poladpur, Ambenali, and Mahabaleshwar Formations cover circular footprint areas with radius 600 km, and that all the other formations have circular footprint areas with radius 250 km, as suggested for the Kalsubai subgroup (Jay and Widdowson, 2008). Total volume is less than half the earlier estimate of 1.3×10^6 km^3. Dashed lines represent uncertainty about the pre– and post–Poladpur/Ambenali/Mahabaleshwar Formations. Erupted volumes are proportional to the areas of the bars representing each formation.

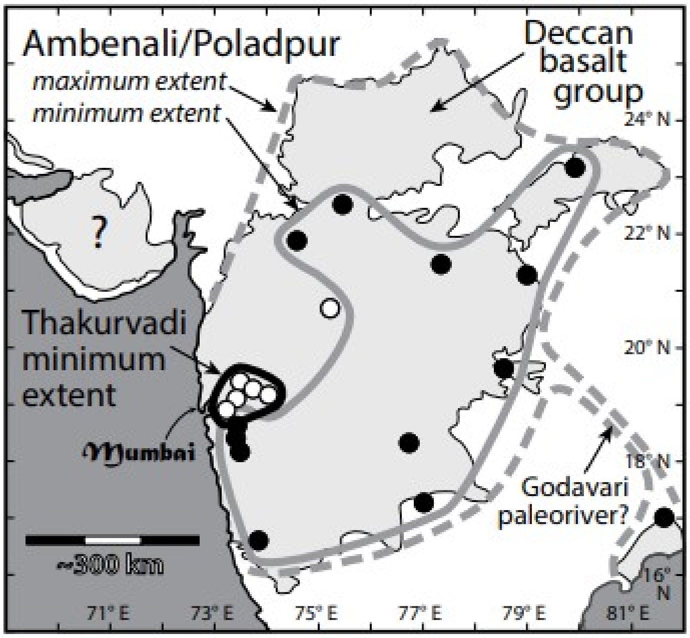


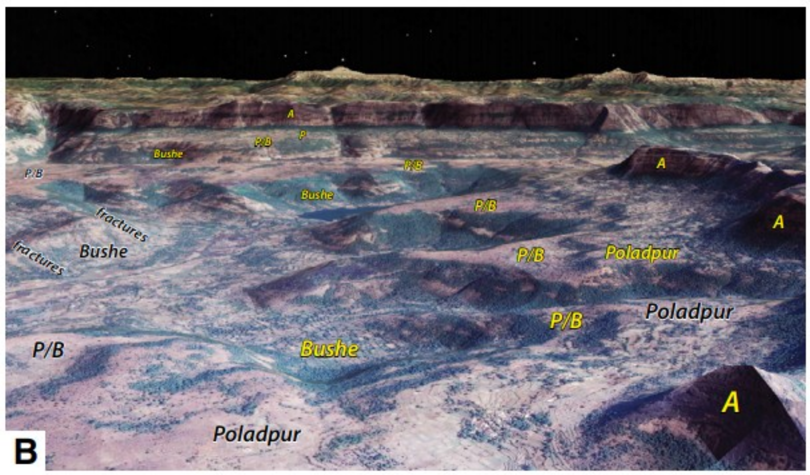
Figure 2. Map of the minimum (mapped) and maximum (inferred) areal extent of the Ambenali and/or Poladpur Formations. Solid circles indicate field locations (Vanderkluysen et al., 2011) in which Ambenali and/or Poladpur have been found; open circles are locations where these formations have not been found. The limited mapped occurrences of thick Thakurvadi Formation are enclosed by thick black line. Gray stippled area gives the areal extent of the Deccan Flood Basalt Group.

TABLE 1. PUBLISHED $^{40}Ar/^{39}Ar$ AGES ON THE DECCAN VOLCANIC PROVINCE

Sample	Formation	Quality	Age (Ma)	±2σ (Ma)	Author
AM83-7	Mahabaleshwar	1	63.1	1.0	Venkatesan et al. (1993)
MB81-24	Mahabaleshwar	1	63.4	0.8	Venkatesan et al. (1993)
RA99.06	Mahabaleshwar	1	64.1	1.1	Knight et al. (2003)
RA99.14	Mahabaleshwar	1	64.9	0.4	Knight et al. (2003)
RA99.23	Mahabaleshwar	1	65.3	0.4	Knight et al. (2003)
MAP-052MP	Mahabaleshwar	0	65.7	1.0	Baksi et al. (1994)
MAP-056	Mahabaleshwar	0	65.8	2.8	Duncan and Pyle (1988)
MAP-057	Mahabaleshwar	1	69.7	2.4	Duncan and Pyle (1988)
RA99.11	Ambenali	1	64.3	0.8	Knight et al. (2003)
MB81-10	Ambenali	1	65.1	1.0	Venkatesan et al. (1993)
RA99.12	Ambenali	1	65.33	1.13	Knight et al. (2003)
CAT-034	Ambenali	0	65.5	1.8	Duncan and Pyle (1988)
RA99.1B	Ambenali	1	65.6	1.6	Knight et al. (2003)
RA99.02	Ambenali	1	66.4	0.5	Knight et al. (2003)
MB81-4	Ambenali	0	67.1	0.8	Venkatesan et al. (1993)
CAT-021	Ambenali	0	67.8	2.0	Duncan and Pyle (1988)
Mur 2	Poladpur	1	65.84	0.58	Hooper et al. (2010)
Ma 1	Poladpur	1	65.93	0.60	Hooper et al. (2010)
MB81-3/A	Poladpur	1	68.04	0.8	Venkatesan et al. (1993)
MB81-3/B	Poladpur	0	68.6	0.8	Venkatesan et al. (1993)
JEB127	Bhimashankar	1	67.23	0.6	Pande (2002)
JEB-013MP	Thakurvadi	0	65.9	1.2	Baksi et al. (1994)
JEB-013QT	Thakurvadi	0	66.7	0.8	Baksi et al. (1994)
IG82-27	Thakurvadi	1	67.5	0.8	Venkatesan et al. (1993)
IG82-39	Thakurvadi	1	67.8	0.6	Venkatesan et al. (1993)
IG82-34	Thakurvadi	1	68.6	0.6	Venkatesan et al. (1993)
JEB-023	Thakurvadi	0	69.0	3.6	Duncan and Pyle (1988)
JEB-311	Neral	0	61.9	4.2	Duncan and Pyle (1988)
JEB-339Q	Neral	1	66.4	1.0	Baksi et al. (1994)
JEB-339MP	Neral	0	67.1	1.4	Baksi et al. (1994)
TEM-004	Neral	1	68.0	1.4	Duncan and Pyle (1988)
JEB-334B	Igatpuri	0	63.8	4.2	Duncan and Pyle (1988)
JW7	Igatpuri	1	66.5	2.0	Hofmann et al. (2000)
JW4	Jawhar	1	65.7	1.5	Hofmann et al. (2000)
JW2	Jawhar	1	66.1	1.3	Hofmann et al. (2000)
JW5	Jawhar	1	66.1	1.6	Hofmann et al. (2000)
JW6	Jawhar	1	66.4	1.3	Hofmann et al. (2000)
IGA-009Q	Jawhar	1	66.4	1.2	Baksi et al. (1994)
IG82-4	Jawhar	1	67.8	0.6	Venkatesan et al. (1993)
IGA-004	Jawhar	0	68.7	1.2	Duncan and Pyle (1988)

Note: Ages are ordered first by formation, in stratigraphic sequence (youngest at the top), and then by age within each formation. Ages and 2σ errors have been recalculated using the same calibration as in Renne et al. (2013). Quality 1 signifies a plateau age that we judge to be meaningful; quality 0 signifies a plateau age possibly biased by alteration and/or recoil artifacts.

Figure 3. The Poladpur-Bushe contact as seen in the field (A) and in satellite images using Google Earth (B–D). Abbreviations, from bottom to top stratigraphically: Bh—Bhimishankar Formation; K—Kandala Formation; B—Bushe Formation; P/B—Poladpur/Bushe terrace; P—Poladpur Formation; A—Ambenali Formation. (A) Photograph of Poladpur-Bushe contact exposed in road cut east of Mahad. Head of rock hammer is at contact between top of an extremely weathered lava flow (Bushe?) and an ~10–15-cm-thick laminated depositional layer of unknown origin. This layer is overlain by a relatively intact lava flow believed to be Poladpur Formation. (B–D) Satellite images of Western Ghats (Sahyadri) escarpment from the regions west of Pune and Mahabaleshwar, showing the prominent topographic terrace at contact between Bushe and Poladpur Formations (and between Lonavala and Wai Subgroups), and showing that formations below this contact are cut by numerous fractures and at least one fault, while formations above the contact do not display these deformational features. This difference suggests that the Poladpur-Bushe contact may be a hiatus representing a considerable interval of time. Formations were identified on the basis of maps of Beane et al. (1986) and Devey and Lightfoot (1986) and Deccan field experience of authors Stephen Self and Loÿc Vanderkluysen. The Poladpur-Bushe terrace was recognized during field work in 2014. (B) Low-angle oblique northward view from Google Earth, showing the prominent terrace eroded on the Poladpur-Bushe contact. Center of the image is at 18.138°N, 73.590°E; village in the foreground is at 18.1106°N, 73.5812°E. (*Continued on following page*.)



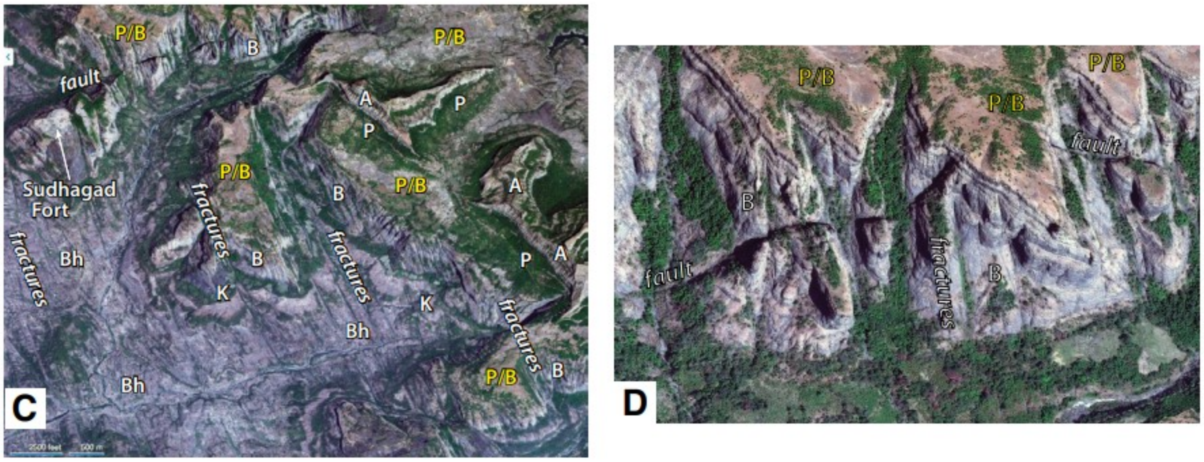


Figure 3 (*continued*). (C) High-angle oblique "bird's-eye" northward view of a high-resolution image from Bing Maps, to show the widespread, dense array of NNW-ESE fractures and local NE-SW fractures (just east of the center of the image), as well as the WSW-ENE fault in the upper-left corner, all of which affect the Bhimishankar, Kandala, and Bushe Formations but do not extend upward into the Poladpur-Bushe terrace or the Poladpur or Ambenali Formations. Center of the image is at 18.522°N, 73.357°E. The NNW-ESE fracture set is also marked in B. (D) High-angle oblique "bird's-eye" northward view of a high-resolution image from Bing Maps, showing a detail of the top center of C. This image makes it clear that both the fault and the NNW-SSE fractures cut the Bushe Formation but do not affect the terrace at the Poladpur-Bushe contact. It also shows that the Poladpur-Bushe terrace is underlain by a thin set of beds, not yet identified, so this is not simply an erosional surface. Center of the image is at 18.556°N, 73.345°E. (Be aware, when examining this and nearby areas on Google Earth, that the digital elevation model [DEM] used in that software contains errors in steep topography, so that flat surfaces like the Poladpur-Bushe terrace may appear to slope down into valleys.)

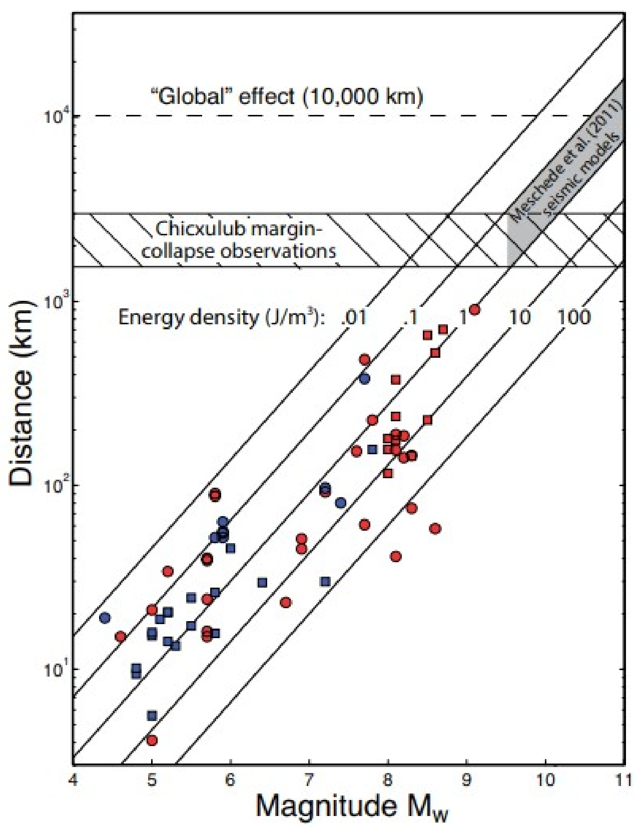


Figure 4. Magmatic and mud volcanoes triggered by earthquakes as a function of moment magnitude (M_w) and epicentral distance. Squares are magmatic volcanoes: red from Manga and Brodsky (2006), and blue from Lemarchand and Grasso (2007). Circles are mud volcanoes: red from the compilation of Manga et al. (2009), and blue representing additional events since that publication identified in Manga and Bonini (2012) and Rudolph and Manga (2010, 2012), plus a recent M_w = 7.7 event in Pakistan on 24 September 2013. Contours are seismic energy density from Equation 1 in the text. Gray stippled area indicates maximum energy densities reached globally for Chicxulub inferred from Meschede et al. (2011). The horizontal cross-hatched bar indicates distal continental margin collapse events (Bralower et al., 1998; Klaus et al., 2000).

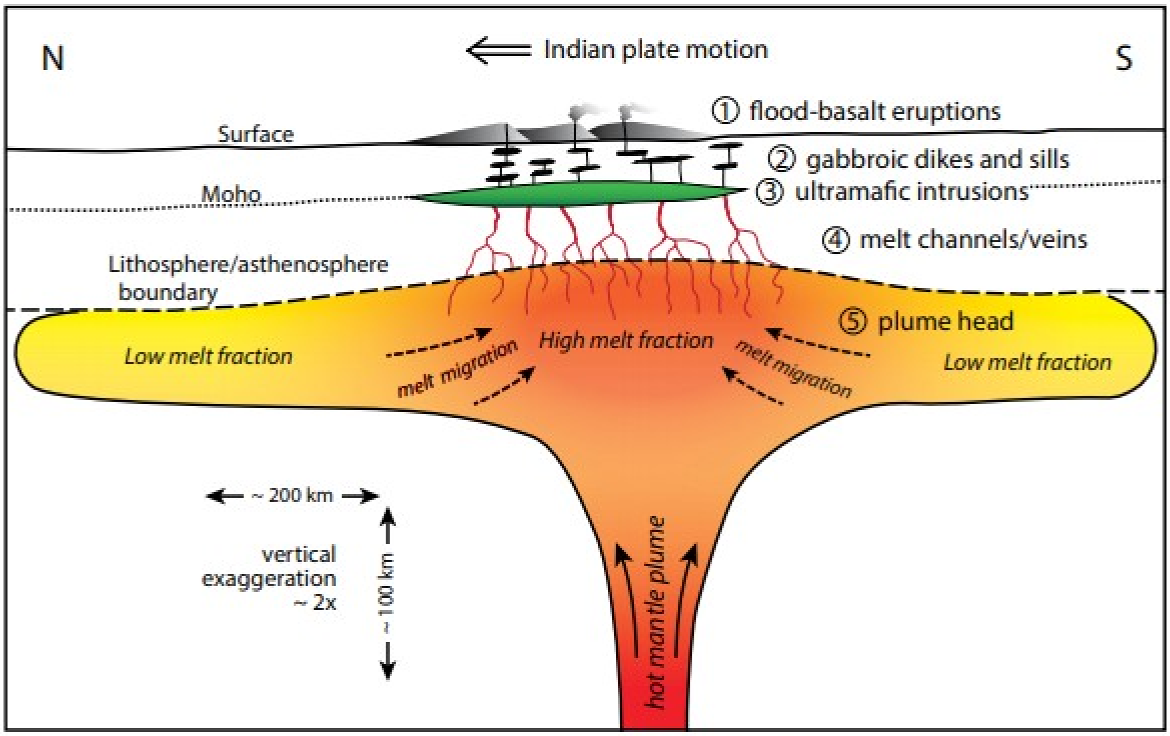


Figure 5. Cross-sectional diagram of the Deccan plume head melting beneath the Indian subcontinent. Magma transport regimes 1–5 described in text.

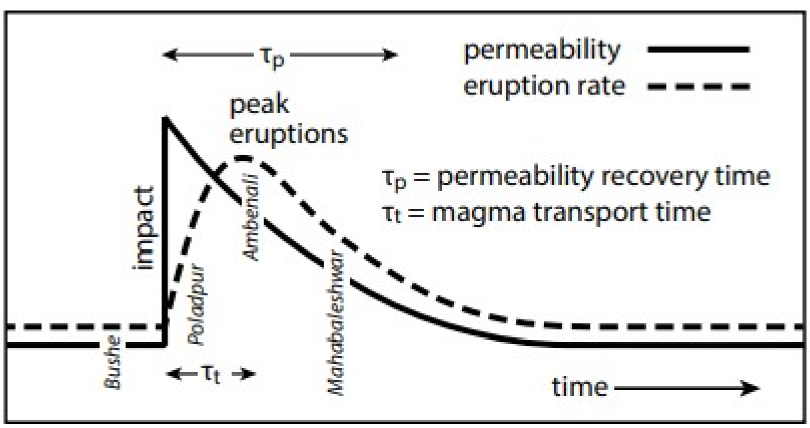


Figure 6. Time line for the response of the Deccan magmatic system to the seismic disturbance due to Chicxulub.

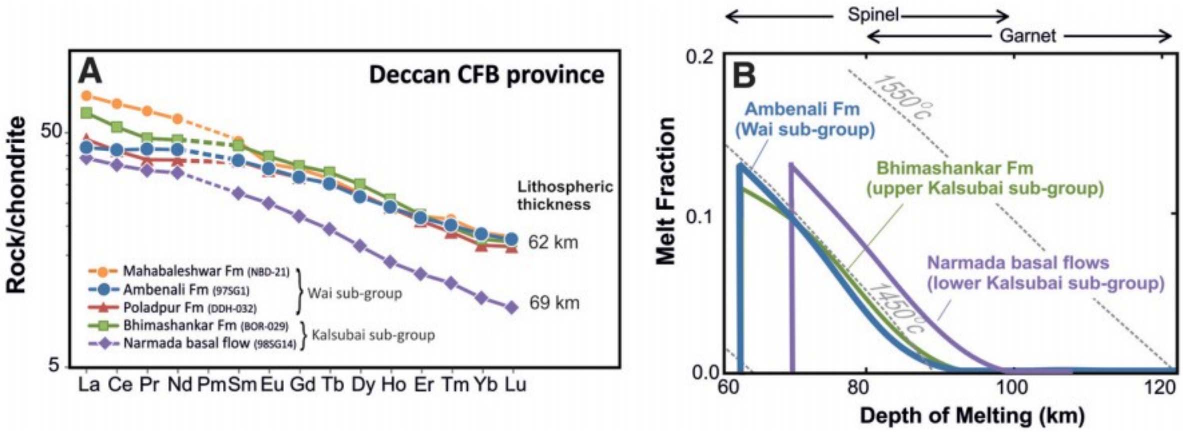


Figure 7. (A) Conventional chondrite-normalized (McDonough and Sun, 1995) rare earth element (REE) plot illustrating analyses of individual basalts from various Deccan formations. Where possible, we show data for basalts that have undergone minimum lithospheric contamination, as evident from ε_{Nd} values and concentrations of strongly incompatible trace elements. Nevertheless, there are clearly variations in the light to middle rare earth element (LREE/MREE) ratios (e.g., La/Sm) of the different basalt formations; these most likely reflect variable amounts of contamination during ascent through the lithosphere, but this crustal processing does not appear to significantly affect MREE/HREE ratios, which are important tracers of lithospheric thinning. Estimates of lithospheric thickness are from REE inversion modeling (see GSA Data Repository material [see text footnote 1]). Whole-rock data are from Vanderkluysen et al. (2011), Gibson (2000), and this work. (B) Melt fraction vs. depth curves from REE inversion modeling. CFB—continental flood basalt.