# The magmatic architecture of continental flood basalts I : Observations from the Deccan Traps

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**Abstract**

Flood basalts are some of the largest magmatic events in Earth history, with intrusion and eruption of millions of km$ˆ3$ of basaltic magma over a short time period ($\sim$ 1-5 Ma). A typical continental flood basalt (CFB) is emplaced in hundreds of individual eruptive episodes lasting decades to centuries with lava flow volumes of 10$ˆ3$- 10$ˆ4$ km$ˆ3$. These large volumes have logically led to CFB models invoking large magma reservoirs ($>$ 10$ˆ5$-10$ˆ6$ km$ˆ3$) within the crust or at Moho depth. Since there are currently no active CFB provinces, we must rely on observations of past CFBs with varying degrees of surface exposure to develop and test models. In the last few decades, significant improvements in geochronological, geochemical, paleomagnetic, volcanological, and paleo-proxy measurements have provided high-resolution constraints on CFB eruptive tempo - the volume, duration, and frequency of individual eruptive episodes. Using the well-studied Deccan Traps as an archetype for CFB systems, we compile multiple lines of evidence - geochronology, eruption tempo, dike spatial distribution, intrusive-extrusive ratio, geochemical variations, and volcanological observations - to assess the viability of previous models. We find that the presence of just a few large crustal magma reservoirs is inconsistent with these constraints. Although observations from the Deccan Traps primarily motivate our model, we discuss constraints from other CFBs to illustrate that this conclusion may be broadly applicable, with important implications for interpreting CFB geochemical datasets as well as the timing and volumes of climate-altering volatile emissions associated with CFBs.

# The magmatic architecture of continental flood basalts I : Observations from the Deccan Traps

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**Key Points:**

- We compile diverse observations from geochronology, geochemistry, volcanology for the Deccan Traps to constrain magmatic architecture.
- These different datasets consistently suggest large, frequent eruptions for Deccan Traps (and potentially other CFBs).
- Constraints from Deccan eruptive tempo, geophysics, and geochemistry are inconsistent with the large crustal magma reservoir model.

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**Abstract**

Flood basalts are some of the largest magmatic events in Earth history, with intrusion and eruption of millions of km$^3$ of basaltic magma over a short time period ($\sim$ 1-5 Ma). A typical continental flood basalt (CFB) is emplaced in hundreds of individual eruptive episodes lasting decades to centuries with lava flow volumes of $10^3$- $10^4$ km$^3$. These large volumes have logically led to CFB models invoking large magma reservoirs ($> 10^5$-$10^6$ km$^3$) within the crust or at Moho depth. Since there are currently no active CFB provinces, we must rely on observations of past CFBs with varying degrees of surface exposure to develop and test models. In the last few decades, significant improvements in geochronological, geochemical, paleomagnetic, volcanological, and paleo-proxy measurements have provided high-resolution constraints on CFB eruptive tempo - the volume, duration, and frequency of individual eruptive episodes. Using the well-studied Deccan Traps as an archetype for CFB systems, we compile multiple lines of evidence - geochronology, eruption tempo, dike spatial distribution, intrusive-extrusive ratio, geochemical variations, and volcanological observations - to assess the viability of previous models. We find that the presence of just a few large crustal magma reservoirs is inconsistent with these constraints. Although observations from the Deccan Traps primarily motivate our model, we discuss constraints from other CFBs to illustrate that this conclusion may be broadly applicable, with important implications for interpreting CFB geochemical datasets as well as the timing and volumes of climate-altering volatile emissions associated with CFBs.

## 1 Introduction

Continental flood basalt provinces (CFBs) are cataclysmic magmatic events, whose brief “main-phases” (durations $\sim$ 1 Ma; V. E. Courtillot & Renne, 2003; Bryan et al., 2010; V. Courtillot & Fluteau, 2014; R. E. Ernst & Youbi, 2017; H. Svensen et al., 2018) are associated with eruption of millions of km$^3$ of dominantly pāhoehoe basaltic lava flows over vast areas (e.g., Self et al., 1998; J. J. Mahoney & Coffin, 1997; Bryan & Ferrari, 2013; R. E. Ernst, 2014, and references therein). CFBs are commonly associated with a large degree of mantle melting due to the arrival of a deep mantle plume head at the base of the lithosphere (e.g., M. A. Richards et al., 1989; Campbell & Griffiths, 1990; Farnetani & Richards, 1994; R. E. Ernst, 2014; R. E. Ernst et al., 2019). Typically, full CFB sequences have an overall duration of about 5-15 Ma (V. E. Courtillot & Renne, 2003; V. Courtillot & Fluteau, 2014; H. Svensen et al., 2018), with the much briefer main-phase eruptions accounting for the majority of the erupted volume (e.g., $>$ 60% for the Deccan Traps, M. A. Richards et al. (2015), $\approx$ 87 % for the Columbia River Basalt, Kasbohm and Schoene (2018)). CFBs are also important events for solid earth-climate interaction, since they are frequently temporally correlated with significant environmental perturbations on a global scale, including major mass extinctions and rapid climate change (e.g., Wignall, 2001; M. T. Jones et al., 2016; R. E. Ernst & Youbi, 2017; Clapham & Renne, 2019; Torsvik, 2020).

The main phase of each CFB is composed of hundreds of individual eruptive episodes, each representing almost continuous eruptions from a single or set of connected vents (Self et al., 2014). In the field, each eruptive episode comprises a flow-field built up of one or several lava flows (See Thordarson & Self, 1998; Self et al., 1998; Jay et al., 2009, for more discussion of the terminology). Analysis of typical CFB flow fields, especially for the well-studied Columbia River Basalt province, suggest that they were emplaced over at least a decade, and likely over multiple centuries (Vye-Brown, Self, & Barry, 2013). Individual flow fields in CFBs have lava volumes ranging from $10^3$- $10^4$ km$^3$ (Self et al., 2008; Bryan et al., 2010; Self et al., 2014).

By comparison with present-day basaltic volcanism (e.g., Kilauea, Laki, Piton de la Fournaise), individual CFB eruptive episodes represent a separate class of basaltic volcanism in terms of both volume (1000s of $km^3$ vs. 19.6 $km^3$ for Eldgjá 939 eruption, the largest known lava eruption in human history) and areal extent ($10^5$ $km^2$ vs. 780 $km^2$ for Eldgjá 939 eruption) (Thordarson et al., 2001; Oppenheimer et al., 2018). Hence, the term "flood" basalt. *These huge erupted volumes, combined with relatively homogeneous magma compositions during eruptive episodes, pose significant challenges to understanding crustal magma transport, underscoring two fundamental questions: What are the geophysical conditions, with respect to melt generation and transport, that are required for CFBs ? What is the crustal plumbing system of these CFBs that permits large, repeated individual eruptive events? This magmatic architecture is related to a key question: why do flood basalts erupt persistently in such large eruptive episodes?*

In this set of papers, we explore the hypothesis that the unique character of "flood" basalt eruptions is a consequence of extraordinarily large mantle melt flux distributed over a broad region at the base of the lithosphere. This magma flux in turn establishes a distinct crustal magmatic architecture (compared to modern Ocean Island Basalts), which acts as the transfer function between the quasi-continuous mantle melts and the spatially and temporally localized individual eruptive episodes. Melt transport through the mantle lithosphere may act as an additional control on this behavior of melt focusing. However, once the melt pathways (dikes or two-phase channelization instabilities) are established, the timescale for magma transport through this system is expected to be relatively fast, continuous, and long-lived (Lister & Kerr, 1991; Rubin, 1995; Fowler, 2011).

The CFB magmatic system must consist of an interconnected network of sills or magma reservoirs (layered mafic intrusions) throughout the crust, Moho depth (magmatic underplating), and within the mantle lithosphere (R. E. Ernst et al., 2019, and references therein). This magmatic plumbing network transports millions of $km^3$ of magma from the underlying mantle plume sources and distributes it as intrusions, and onto the Earth's surface as flood basalts (e.g., R. E. Ernst, 2014). However, there is significant uncertainty on how large these components are and how they combine into a crustal magma system geometry that enables efficient transfer of large magma volumes from a mantle plume source to the surface (Jerram & Widdowson, 2005; R. E. Ernst, 2014; Cruden & Weinberg, 2018; Coetzee & Kisters, 2018; Magee et al., 2018; Magee, Ernst, et al., 2019). The only quantitative models for CFB magmatic architecture (Karlstrom & Richards, 2011; Black & Manga, 2017, see more description in Section 2.2) assume that the large eruptive episodes are fed from correspondingly large magma reservoirs ($> 10^4$-$10^6$ $km^3$) within the crust and/or at Moho depth through dikes. As we show in later sections of this paper, this assumption is not well-founded in terms of available constraints.

In order to test possible models, we must rely upon observations of erupted CFBs since there is no modern-day active CFB province where the magmatic architecture can be inferred from direct geophysical methods. However, one of the primary difficulties with this approach is that we seldom have exposure of both the sub-crustal magmatic system and the overlying lava flow system in the same setting. For example, layered mafic intrusions (e.g., the Bushveld complex) and giant radiating dike swarms (e.g., McKenzie dike swarm) are interpreted to represent the crustal magma system and the feeders for massive overlying volcanic provinces (R. Ernst et al., 2010; Buchan & Ernst, 2019; R. E. Ernst et al., 2019). However, in most of these cases, there are either no remaining lava flows or geochronologic constraints on the eruptive fluxes or volumes of lava flows (e.g., Bushveld complex Lenhardt & Eriksson, 2012). Additionally, there is the additional complexity that we are observing the final integrated system after solidification, and it is difficult to infer the temporal history vis-a-vis surface eruptions.

Conversely, for most Phanerozoic flood basalts with good temporal constraints on eruption rate for individual flow units, there are limited direct observations (exposures) of the trans-crustal magmatic system. In a few CFBs such as the Siberian

Traps, North Atlantic Magmatic Province, and Karoo-Ferrar Traps (all with a thick sedimentary basin overlying the basement and underlying the lavas), multiple studies have analyzed the numerous interconnected stacked sill-dike complexes to provide constraints on very shallow (< 3 km) CFB plumbing systems (e.g., H. Svensen et al., 2012; S. M. Jones et al., 2019; Hoggett, 2019; Magee, Ernst, et al., 2019, and references therein). Analogously, extensive geophysical and geochemical analysis of the dike swarms feeding surface lava flows has helped develop our understanding of the shallow CFB plumbing system (e.g., Vanderkluysen et al., 2011; Rivalta et al., 2015; Kavanagh, 2018; Magee, Ernst, et al., 2019; Buchan & Ernst, 2019). However, the magma transport relationship between dike swarms, sills, mid-crustal, and Moho depth magma reservoirs, and the mantle melt generation region remains unclear due to limited observational constraints (e.g., D. H. Elliot & Fleming, 2018, and references therein for the Karoo CFB). The primary direct observational support for the large magma reservoir model comes from the presence of thick ($\sim$ 5-15 km) high-velocity features above the Moho location in multiple CFBs (Emeishan, Deccan, Siberian, and Columbia River flood basalt provinces K. G. Cox, 1993; Ridley & Richards, 2010) as expected from the presence of solidified olivine- and clinopyroxene-rich cumulates in deep magma reservoirs (Farnetani et al., 1996). However, it is difficult to distinguish lower velocity crustal magma bodies from seismic data alone. Similar to layered mafic intrusions, geophysically observed mafic-ultramafic bodies represent the final time-integrated igneous evolution of mantle-generated melts with no information whether the features represent a single magma reservoir that was molten at the same time or was constructed incrementally (R. B. Larsen et al., 2018; Robb & Mungall, 2020).

An alternative method to test the validity of models for CFB magmatic architecture is to use the tempo of eruptive episodes. Specifically, this requires constraints on three key variables: the duration, volume, and frequency of individual eruptive episodes. *The primary process that controls the eruptive tempo must be the magmatic architecture of CFBs*. For example, the analysis by Black and Manga (2017) shows that in a magmatic architecture consisting of two large magma reservoirs (Moho and shallow-crustal levels), buoyancy overpressure is the primary mechanism for triggering CFB eruptions. This buoyancy overpressure is, in turn, controlled by the volatile content of the primary magmas. Due to the incompatible nature of magmatic volatiles ($CO_2$ and $H_2O$) during mantle melting, the initial low degree partial melts will be volatile-rich and are expected to erupt in frequent, low volume eruptive episodes. By contrast, subsequent higher degrees of partial melts will accumulate in a large magma reservoir and become eruptible by volatile buoyancy over $10^5$ - $10^6$ years. Thus, the expected eruptive tempo during the main volumetric phase of CFBs is a few brief ($\sim 10^4$ yr) eruptive episodes with long ($10^5$ - $10^6$ year) hiatuses in surface eruptions in-between them, again incompatible with observations, as discussed in detail below.

In this study, we use a compilation of recent datasets from the Deccan Traps CFB to estimate the eruptive tempo of eruptive episodes, in order to test the models for CFB magmatic architecture. The Deccan Traps (DT) are a continental flood basalt province (primarily tholeiitic) occupying an area of over 500,000 sq. km. in Western India, whose main-phase eruptions spanned $\sim$1 Ma beginning at about 66.3 Ma (e.g., Sprain et al., 2019; Schoene et al., 2019). The primary source of erupted magma was likely decompression melting in a large plume head (hundreds of km across) marking the arrival of the Réunion plume (M. A. Richards et al., 1989; Campbell & Griffiths, 1990). Due to extensive research over the past few decades, there are high-resolution constraints on eruption rate based on geochronology (Renne et al., 2015; Schoene et al., 2015; Sprain et al., 2019; Schoene et al., 2019), paleomagnetic directional groups (Chenet et al., 2008, 2009), as well as Hg chemostratigraphy (Percival et al., 2018; I. M. Fendley, Mittal, Sprain, et al., 2019) along with extensive geochemical characterization of the lava flows (e.g., K. G. Cox & Hawkesworth, 1984; Beane et al., 1986; Lightfoot & Hawkesworth, 1988; J. J. Mahoney, 1988; Z. Peng et al., 1994, 1998; Sano et al., 2001). Additionally, the many geophysical datasets from gravity, seismic, and magnetotelluric observations provide constraints on the sub-surface magmatic structure feeding the lava flows (e.g., Patro et al., 2018). The Deccan Traps have a clear geophysical, as well as geochemical (and isotopic) signature associated with melt production from a mantle plume head, which also provides some constraints on the rate of melt input into the crustal column.

In aggregate, the available data sets constraining eruptive tempo for the Deccan Traps are unique among all CFBs. The closest CFB with comparable data sets is the Miocene Columbia River Basalts (CRB) with extensive geochemical and volcanological studies (P. Hooper, 1988a; P. Hooper & Hawkesworth, 1993; Barry et al., 2013; Wolff & Ramos, 2013). However, the total erupted volume of the CRB ($\geq$ 210,000 km$^3$; S. P. Reidel et al., 2013) is much smaller than for other CFBs. Additionally, the onset of the largest phase of Deccan volcanism, responsible for about 2/3 of the total erupted known DT volume, is coincident, within Ar-Ar age precision, with the Cretaceous-Paleogene boundary (M. A. Richards et al., 2015; Sprain et al., 2019). This temporal coincidence makes understanding the magmatic architecture of the Deccan Traps of special interest, given the recent hypothesis that this most voluminous phase of Deccan eruptions were accelerated ("triggered") by the strong ground motion from the Mw$\sim$11 Chicxulub impact (M. A. Richards et al., 2015).

In Section 2, we review published conceptual and physical models for CFBs to illustrate the predictions from previous models. In Section 3, we provide a summary of the various observational constraints from the Deccan Traps (along with some additional data sets from other well studied flood basalt provinces such as the Columbia River Basalts and Siberian Traps) for eruptive tempo constraints. This is followed by a discussion of the stratigraphic geochemical variations (Section 4), exposed magmatic architecture (dikes and sills, Section 5), and geophysical observations (Section 6) as constraints on the magmatic plumbing system of CFBs. Finally, in Section 7, we summarize the results of our observational analysis and comparison with previous models.

## 2 Proposed Models for CFB magmatic architecture

Upon reviewing the literature on CFBs it becomes apparent that there have been remarkably few quantitative, much less well-accepted, models proposed to explain the uniquely large total erupted volumes (typically $> 10^6$ km$^3$), or the large volumes of individual eruptive episodes, compared to contemporary basaltic volcanism. To first order, it has been proposed that CFBs are just a consequence of the larger mantle melt flux from a mantle plume head with this melt erupting to the surface through some process. Nevertheless, we feel that it is important here to review the basics of the models that have been put forward previously in order to provide a fuller context for understanding our focus on the question: How are the processes of melt generation and lithosphere-crustal melt transport different for CFBs compared to other, much less voluminous, forms of basaltic volcanism?

Regarding melt generation, the conclusion based on a large number of studies analyzing extensive geochemical data sets for major CFBs, seismic imaging, plate reconstruction models, and geodynamic models is that CFB melts are derived from three potential sources: partial melting of hot mantle plume heads (e.g., M. A. Richards et al., 1989; J. J. Mahoney, 1988; R. White & McKenzie, 1995; Lassiter & DePaolo, 1997; S. A. Gibson, 2002; Sobolev et al., 2011; G. Sen & Chandrasekharam, 2011; Jennings et al., 2019, and references therein), melting of the overlying metasomatized lithospheric mantle material due to the plume thermal anomaly (Arndt & Christensen, 1992; J. S. Marsh, 1987; Lightfoot & Hawkesworth, 1988; Mckenzie & Bickle, 1988; Lightfoot et al., 1993; Turner & Hawkesworth, 1995; Natali et al., 2017; Black &

Gibson, 2019; S. A. Gibson et al., 2020, and references therein), and other crustal-asthenospheric melting processes (e.g., Arndt, 1989; J. S. Marsh, 1989; Lassiter & DePaolo, 1997; Kempton & Harmon, 1992; H. C. Sheth, 2005). Of these, the plume head model is by far the most widely accepted. Nevertheless, irrespective of the specific source, the large volume of buoyant partial melts must migrate through the lithosphere via a combination of dikes and/or two-phase channelization instabilities to feed the crustal CFB magmatic system (Rabinowicz & Ceuleneer, 2005; Schiemenz et al., 2011; Solano et al., 2012; T. Keller et al., 2013; Madrigal et al., 2015; Yarushina et al., 2015; Schmeling et al., 2018, 2019).

Rising CFB melts encounter a rheological contrast at or above the Moho since, depending on the geotherm and the fluid saturation state, the lower crust is either weaker than the upper crust, and lithospheric mantle or is stronger than the underlying lithospheric mantle (e.g., Bürgmann & Dresen, 2008; K. G. Cox, 1993; Fyfe, 1992). In both cases, the CFB magmas are expected to accumulate at Moho depth by magmatic underplating. Additionally, depending on the volatile content of the melt, the Moho may also be a density barrier leading to melt accumulation at these depths. Thus there are physical reasons to expect the formation of magmatic reservoirs at Moho depths beneath flood basalt provinces. These conclusions are further supported by various geophysical observations for crustal underplating in flood basalts (K. G. Cox, 1993; Bryan et al., 2010; Ridley & Richards, 2010; M. A. Richards et al., 2015; Farnetani et al., 1996, See Section 6.1 for more discussion). Finally, melt accumulated in these Moho-depth magma reservoirs, as well as additional shallow crustal magma bodies, feed large, dike-fed fissure eruptions (e.g., R. Ernst et al., 2001; R. E. Ernst et al., 2019; Magee et al., 2018, and references therein).

Besides this general framework, it remains unclear how the large individual magma reservoirs are if they also exist at shallower depths, and how the magmatic system supplies vast volumes of magma to large individual eruptions (e.g., Morrison et al., 1985). So far, only a few models have been advanced for crustal magmatic processes specific to CFBs despite decades of geochemical studies on these systems. These models are primarily conceptual in nature, and largely motivated by petrological observations. There are only two published (to the best of the authors' knowledge) quantitative or physical models for crustal magmatic processes specific to CFBs, which seems curious for the largest eruptions of igneous rock on our Earth.

In this study, our focus is on the crustal magma transport system for CFBs. Hence we are largely agnostic about the actual source of primary melting as well as the processes of melt transport through the lithosphere. *Additionally, we use the term "magma reservoir" to refer to a well-mixed magma body which can be represented by a volume-averaged temperature, melt and volatile composition.* We choose this definition not to imply that individual magma bodies of any size must necessarily be compositionally, thermally, and rheologically well mixed (see discussion in B. D. Marsh (2013) about the relevant processes), but rather for model consistency. Typically, most petrological models and magma physics models (including the new models in this Paper II), assume that the magma bodies have a single representative temperature, pressure, and melt composition, such that a single very large magma chamber with significant spatial variations in thermal, chemical, and rheological properties (e.g., mush zones) will be represented as multiple magma reservoirs with a high degree of connectivity in between them.

In the following, we first describe the various conceptual CFB models followed by a description of the two published quantitative models. We specifically highlight the magmatic architecture proposed by these models, especially the size of individual magma reservoirs, along with the observational motivation.

## 2.1 Conceptual Models

The first conceptual magmatic models for continental flood basalts was proposed by K. G. Cox (1980). He envisioned that the CFB magmatic system consists of a large (multi-km thick) crustal sill complex at/or close to Moho depth as the primary melt accumulation location, along with an elaborate network of dikes and sills for feeding surface lava flows. In the K. G. Cox (1980) model, the parental picritic magma (MgO >15 wt.%) undergoes extensive fractional crystallization in the lower crustal sill complex with progressive fractionation of an ultramafic cumulate phase, and with a residual lower density evolved basaltic magma (MgO contents <8 wt.%, also see K. G. Cox & Hawkesworth, 1984). If the density of this magma were sufficiently lower than the overlying crustal column, the consequent buoyancy overpressure was hypothesized to lead to surface eruptions with some short-duration storage in an upper sill complex. A key feature of this model is that the relatively homogeneous composition of flood basalts is explained as a consequence of compositional buffering due to the crystallization of olivine, clinopyroxene, and plagioclase. In addition,K. G. Cox (1980) proposed that the observed small scale variation within this relatively homogeneous basaltic composition can be explained by small contributions of trans-crustal polybaric fractionation during magma ascent through the dike-sill network (See also K. Cox and Hawkesworth (1985), and H. Sheth (2016) for variants of this model).

More recently, R. E. Ernst et al. (2019) proposed a similar structural model for a CFB magmatic system with an extensive primary underplated magma reservoir(s) spanning hundreds of km laterally and up to 20 km in thickness. The magma is then envisioned to ascend both laterally outward and vertically through the crust via a large number of dikes (radiating, linear, or circumferential). During this transport, the magma can periodically accumulate in mafic-ultramafic intrusions of various shapes (e.g., multiple-km wide dikes: 2580 Ma Great dike of Zimbabwe; multiple km thick upper-to-mid crustal sills: 2060 Ma Bushveld and 2710 Ma Stillwater complexes; funnel-shaped intrusions: Rum and Skye igneous complexes associated with 56 Ma North Atlantic Magmatic province), and in sill complexes in sedimentary basins.

Another class of related models has been proposed by Heinonen et al. (2019a), and Neumann et al. (2011) (and references cited therein) for CFB magma transport, specifically focused upon explaining the range of lava major and trace element compositions and mineralogy. These studies use a combination of Assimilation Fractional Crystallization (AFC) and energy-constrained AFC (EC-AFC) models (Spera & Bohrson, 2001; Bohrson & Spera, 2001; Spera & Bohrson, 2002; Bohrson & Spera, 2007) to assess what P-T conditions can match measured compositions. They suggest that the flood basalt magmatic plumbing system consists of a two-stage network wherein the magma undergoes AFC first at deep crustal depths ($\sim$ 10-30 km), followed by Fractional Crystallization (FC) in a shallow magma reservoir (3-5 km depth). A vital feature of this model is that of order 10-20 % assimilation of crustal material into the parental magma only occurs at a deeper depth. The Heinonen et al. (2019b) study proposes that the lack of shallow crustal assimilation is a consequence of sufficiently limited melt transport in individual dikes emplaced in a cold upper crust so as to prevent significant wall-rock melt back as required by observations of CFB dikes (H. L. Petcovic & Dufek, 2005; H. Petcovic & Grunder, 2003; H. C. Sheth & Cañón-Tapia, 2015). Additionally, if the same dike is used again for vertical melt transport, the chilled margins in the dike would provide a chemical barrier to reduce mixing between deep crustal melts and country-rock. Finally, the overall magmatic system is hypothesized to be in a quasi-steady state through active RTF processes (periodically Replenished and Tapped, continuously Fractionating O'Hara & Mathews, 1981). This model is essentially equivalent to REFC - Recharge FC model, (DePaolo, 1981; C.-T. A. Lee et al., 2014) in individual magma reservoirs. This provides a mechanism for buffering magma compositions to explain the relative homogeneity of flood basalt lava compositions (e.g., K. Cox, 1988; Luttinen & Furnes, 2000).

Yu et al. (2015) proposed a very similar conceptual CFB magmatic architecture model to explain the variations in lava compositions in continuous stratigraphic sequences (Also see Potter et al., 2018, for a similar model). They propose that since the earliest magmas of a magma reservoir are emplaced in the colder crust, they will rapidly cool and form a compositional boundary plating layer (a high crystallinity solidification front B. D. Marsh, 2013), which progressively helps reduce crustal assimilation for subsequent magmas emplaced into the reservoir. Similar to other models, Yu et al. (2015) find that REAFC (REFC with crustal Assimilation C.-T. A. Lee et al., 2014) processes are necessary to explain the range and pattern of lava flow compositions (See discussion in Section 4).

A final class of conceptual models for LIP systems describe the long-term evolution of the CFB magmatic system from the initiation of the flood basalt province to its termination. As a representative example, we briefly describe a model presented by S. R. Krans et al. (2018) based on the petrological analysis of continuous 1635 m-thick lava flow sequence of the Ethiopian Traps. The principal observational basis for this model is relative changes in phase assemblages (specifically olivine, clinopyroxene, ortho-pyroxene, and plagioclase) as a function of depth (e.g., Morse, 1980; Albarede, 1992; M. Richards et al., 2013). S. R. Krans et al. (2018) propose that initially, the parental mantle melts accumulate in the Moho depth magma reservoirs that periodically feed ol-phyric surface flows with minimal intermediate storage. Over time, a shallower magmatic plumbing system is established, leading to magma stalling and differentiation, as evidenced by a higher fraction of plag-phyric lava flows. This evolution in the magmatic plumbing network naturally leads to polybaric crystal fractionation (as argued by other conceptual models discussed above). Eventually, decreasing partial melt flux from the lithosphere leads to freezing of an increasing fraction of crustal magma reservoirs and increasing fractionation of the magmas that do erupt. Finally, the termination of the flood basalt activity in Ethiopian Traps is followed by a transition to much lower volume, localized shield building (Kieffer et al., 2004). S. R. Krans et al. (2018) posited that atop this general evolution of the magmatic system, there are variations in magmatic flux into the shallow crustal system. These include multiple pulses of magma recharge and brief hiatuses (as indicated by weathering horizons). Since there are no evident systematic mineralogical or volcanological variations on a flow-by-flow scale to these variations, they propose that the complex response is indicative of a complex magma plumbing system with multiple reservoirs. Depending on their pathway through the magmatic system, different melt batches will have varying storage times before eruption and hence experience different amounts of magma mixing and fractionation.

As additional evidence for this inference, S. R. Krans et al. (2018) interpret the occurrence of the few plagioclase-megacrystic basalt flows as indicative of a well established shallow magmatic mush zone. Based on the presence of complex zoning patterns and high An-content in plagioclase megacrysts, they propose that frequent recharge of deep melt into the magma reservoirs is required to explain these flows, with some of the larger influxes leading to large scale re-mobilizing the mush zone and eruption. A critical feature of this class of conceptual models is the presence of a complex magmatic plumbing network and the presence of multiple magma reservoirs and eruptive pathways in the system.

In summary, the various conceptual models all propose the presence of a deep crustal/Moho depth magma reservoir as well as some significant upper crustal plumbing network, which is continuously replenished in order to explain both the overall geochemical homogeneity as well as inter-flow variations in a stratigraphic section. Although these models do not explicitly discuss the size of individual magma reservoirs, it appears to be generally assumed that they are large enough to feed individual eruptive episodes (a size concomitant with volumes in excess of 1000 $km^3$ per eruption). The one exception to this is H. C. Sheth and Cañón-Tapia (2015) who explained geochemical variations within composite dikes in Deccan Traps with a model wherein individual CFB eruptive episodes are fed by multiple interacting magma reservoirs.

## 2.2 Quantitative Models

### *2.2.1 Recharge-crystallization model*

A first quantitative model for CFB magmatic systems was proposed by Karlstrom and Richards (2011). Akin to the conceptual models described above, they assumed that the parental ultramafic melt forms ellipsoidal intrusions at Moho depth or in the lower crust, since they are denser than the overlying crust. With subsequent fractional crystallization (up to $\sim$ 30% crystallinity), $CO_2$ exsolution from the melt makes the melt - volatile mixture buoyant and eruptible. Using thermo-mechanical models, they showed that during the early phases of a CFB, melt in a magma reservoir of volume comparable to the size of a single eruptive episode ($10^3$ - $10^4$ $km^3$, $\sim$ 10-30 km semi-major axis horizontal length, 1-5 km vertical length) cools sufficiently to become buoyant on a timescale shorter than the viscous stress relaxation timescale. Consequently, buoyant melt from each of these reservoirs feeds an individual surface eruption through dikes initiated by elastic stresses in the lower crust caused by the growth of the magma reservoirs (recharge). Continued magma flux will progressively heat the lower crust, leading to faster viscous stress relaxation and dominantly intrusive magma bodies instead of surface lava flow eruptions. For representative melt fluxes and crustal rheology, Karlstrom and Richards (2011) found that this transition occurs on a timescale of order 1 Ma, which is comparable to the duration of the main phase of CFB emplacement.

The Karlstrom and Richards (2011) model envisions the formation of a dense network of intrusions in the lower crust. Due to melt buoyancy and background tectonic stresses, these magma reservoirs are expected to progress upward into colder, more elastic crust. The model also proposes that the maximum size of individual reservoirs is similar to the depth from the surface. This restriction is a consequence of free-surface shear stress and dike initiation at the edges of reservoirs for laterally extensive CFB magma bodies. Based on analog experiments and seismic datasets for sill complexes in Karoo Basin and North Atlantic magmatic province, the typical total diameter of sills is 1.5- 15 times the depth of the emplacement (e.g., Hoggett, 2019, and references therein), with a median value of 2.5 consistent with theoretical expectations (Malthe-Sørenssen et al., 2004; Manga & Michaut, 2017; Galland et al., 2009). We would however note that there are some very large sills (e.g., Basement Sill with $\sim$ 10,000 $km^2$ area, 400 m thickness) exposed in a 3-4 km deep surface section of the magmatic plumbing structure of the Karoo-Ferrar CFB in McMurdo Dry Valleys, Antarctica (B. D. Marsh, 2004). Thus, in practice, the concentration of free-surface shear stress may not be a strong limit on the spatial size of magma bodies due to visco-plastic deformation processes (Schofield et al., 2012; Galland et al., 2019). Finally, the eruption frequency is controlled by the melt recharge timescale, which determines the rate of deviatoric stress buildup in the lower crust and cooling timescale for buoyancy production by $CO_2$ exsolution. Since the lower crustal temperature will rapidly increase during the course of a CFB event, the first-order prediction from this model is that the time period between individual eruptive episodes will increase during generation of a CFB province, potentially with larger eruptive volumes later (e.g., Deccan Traps) due to expanding magma reservoirs.

#### *2.2.2 Buoyancy overpressure model*

Another quantitative model for CFBs was proposed by Black and Manga (2017) with a focus on explaining the available geochronological constraints and paleo-proxy inferred environmental changes for the Siberian Traps. They model the CFB plumbing architecture as composed of two magma chambers, one at Moho-depth and another at upper crustal depth. The reservoir sizes dynamically evolve in the model with melt influx from the underlying magma reservoir (mantle) and outflux to surface eruptions (overlying chamber) for the crustal-depth (Moho-depth) reservoir. The critical feature of this model is that the brittle tensional failure of large magma reservoirs is hypothesized to be a consequence of the buoyancy of the magma-volatile mixture. This buoyancy with respect to the surrounding country rock ($\delta_{rho}$) leads to a buoyancy overpressure ($\Delta P_{buoy} = \delta_{rho} g h$; h is the magma reservoir thickness). This buoyancy overpressure cannot be relaxed by the viscous stress relaxation. Hence, $\Delta P_{buoy}$ can slowly accumulate over time without dissipation except by fluid flow into the surrounding crust.

Black and Manga (2017) used a 1D thermo-chemical model of the two magma reservoirs with thermal evolution, volatile exsolution, crustal assimilation of carbon-rich crust, bubble coalescence, and diffusive volatile escape into the country rock, and found that buoyancy overpressure alone is sufficient for the failure of large magma reservoirs. With regard to the eruptive tempo of CFBs, they find three primary eruptive regimes. At the start of a flood basalt sequence, the low-degree melts are volatile-rich and hence highly buoyant due to volatile exsolution during decompression. Consequently, the melt rises rapidly from the Moho-depth magma reservoir to surface eruptions with minimal storage in the crustal reservoirs. Additionally, the size of individual magma reservoirs is small ($\sim$ 1-2 km in height) since most of the mantle melt erupts rather than accumulating. With time, volatile content in the higher-degree mantle melts decreases, reducing magma buoyancy. As buoyancy slowly builds up slowly with volatile-exsolution as well as thicker/larger magma chambers (since $\Delta P_{buoy} \propto h$, the typical thickness is 8-15 km), the eruption frequency decreases along with a larger volume of individual eruptions. The eruptions are further inhibited by volatile loss to the surrounding medium during this accumulation phase. As a result, crustal permeability around the magma reservoir needs to be reduced sufficiently by thermal annealing of fractures, and crustal compaction before overpressure can build up and lead to crustal failure. These processes are repeated in the upper-crustal reservoir, which has an additional source of volatiles from the assimilation of carbon-rich crust (typically organic sediments in sedimentary basins, e.g., Karoo sills and Siberian Trap Sills H. H. Svensen et al., 2018), but also higher crustal permeability (Ingebritsen & Manning, 2010; Mittal & Richards, 2019). Finally, in the third stage, the decrease in parental flux prevents the buildup of sufficient $\Delta P_{buoy}$ in the magma reservoirs, and the reservoirs remain molten with some overpressure for $10^5$ - $10^6$ yr before eventual solidification.

During the second eruptive regime, Black and Manga (2017) found that the eruptive tempo of surface volcanism is very pulsed with perhaps just 2-4 eruptive events. Each eruptive pulse is expected to last $10^3$ - $5\text{x}10^4$ yrs with long hiatus of $2\text{x}10^5$ - $5\text{x}10^5$ yrs. Since each failure event at the Moho depth transfers a large volume of magma to the upper crustal system, the size of the upper crustal magma reservoirs feeding each eruptive episode is large (typical thickness is 2-4 km). Extrapolating from the 1D calculations, Black and Manga (2017) propose that a full continental flood basalt sequence is fed by order 1-10 pairs of Moho and upper crustal magma reservoirs each with a melt source in the mantle of order 100-300 km (compaction lengthscale). Thus, on a province scale, the hiatuses between eruptive pulse may be shorter of order $10^5$ - $5\text{x}10^5$ yrs, whereas individual stratigraphic sections would have longer hiatus since the erupted lava is expected to have been sourced from the same localized eruptive plumbing system in the crust.

#### *2.2.3 Difficulties associated with large magma reservoirs*

Although assuming that each large eruptive episode is fed from a correspondingly large magma reservoir makes logical sense, at least as a starting point, the description of the two models described above (Black & Manga, 2017; Karlstrom & Richards, 2011) illustrate that both the assembly and eruption of large magma reservoirs is challenging. In addition, the results illustrate that the presence of larger magma reservoirs naturally leads to more prolonged hiatus between eruptive episodes.

Among the two primary mechanisms for triggering eruptions, the efficiency of recharge associated overpressure decreases for a large reservoir for the following reasons. First, since the recharge associated overpressure is inversely proportional to the volume of the magma reservoir, it becomes progressively harder to erupt large magma bodies (Jellinek & DePaolo, 2003; Karlstrom et al., 2010; Townsend et al., 2019). Second, with the progressive assembly of lower crustal magma bodies, the increasing crustal temperature leads to faster visco-elastic stress relaxation of any accumulated recharge associated overpressure (Jellinek & DePaolo, 2003; Karlstrom et al., 2017). Additionally, in order to assemble the large magma bodies in the first place, the magma recharge rate needed to be low enough to prevent an eruption. Thus, using recharge-induced over-pressurization for the assembled larger magma bodies requires the special, likely unreasonable, co-incidence of significant changes in melt influx into the magma reservoir. Although there may be some physical feedbacks for melt focusing (e.g., Karlstrom et al., 2009, 2015), it would seem unlikely that these would counteract the negative feedbacks described above.

As a consequence, recharge associated elastic stresses alone become inefficient as triggers for the eruption of large magma bodies, so that additional mechanisms such as buoyancy overpressure (e.g., Caricchi et al., 2014; Black & Manga, 2017) or roof failure (e.g., de Silva & Gregg, 2014; Gudmundsson, 2016) are required. However, if buoyancy overpressure is the eruption trigger, a long eruption hiatus is a natural consequence due to the timescale for vertically assembly of the magma body (which would be slowed by lateral viscous flow), the timescale for crustal permeability reduction, and the timescale for sufficient solidification for volatile exsolution (Karlstrom et al., 2010; Black & Manga, 2017; Mittal & Richards, 2019). Thus, we contend that if individual eruptive episodes are indeed fed by large magma reservoirs, the eruptive tempo is naturally expected to be very pulsatory with hiatus of order $10^5$ - $10^6$ yrs, completely contrary to observations. Additionally, a large magma reservoir will lead to large melt fluxes into the upper crustal reservoir when crustal failure does happen, and hence naturally leads to very large crustal magma reservoirs. Finally, since crustal assimilation of carbon rich country rock is an important source of volatiles for the upper-crustal magma reservoir, the eruptive tempo will be expected to be different for CFBs emplaced into a granitic upper crust instead of a thick sedimentary basin (e.g., Siberian Traps). Specifically, one would expect that the hiatus time between eruptive episodes is longer for CFBs with granitic upper crust (e.g.,Deccan Traps) along with overall reduced likelihood of eruption.

## 3 Observational constraints on magmatic architecture

In order to test the predictions from the previous models for CFB architecture, we use the Deccan Traps flood basalt as a representative example. Since the 1980s, the DT has been extensively studied and consequently has the most extensive geochemical, geochronological, and volcanological datasets among the large ($>$ 1 Million km$^3$ erupted volume) CFBs. In the following, we first give a brief geological overview of the Deccan Traps with a special emphasis on the chemo-stratigraphic framework and contribution of different crustal, lithospheric, and mantle sources to the erupted lava flows. We then summarize the direct constraints on the eruptive tempo of individual eruptive episodes for the DT main phase using geochronology, paleomagnetic secular variation, Mercury proxy records, and lava flow morphology. Although our primary focus for this study is the DT, we include some complementary observations from other CFBs, layered mafic intrusions, and modern flood basalt analogs (e.g., the 1783 Laki eruption) to assess whether our DT observations are representative of CFBs in general.

### 3.1 Geological Background - Deccan Traps

The Deccan Traps is a late Cretaceous–Paleogene continental flood basalt province covering more than 500,000 km$^2$ of peninsular India (J. J. Mahoney, 1988; Verma & Khosla, 2019; Kale et al., 2020; Manu Prasanth et al., 2019; Krishnamurthy, 2020, and references therein). The Deccan Traps mark the beginning of the Réunion hotspot track and are associated with partial melting due to the arrival of a deep mantle plume under the Indian subcontinent (presently beneath Réunion) (e.g., M. A. Richards et al., 1989); see Peters et al. (2017) for discussion of isotopic evidence). The present-day subaerial volume of the DT lava flows is about 600,000 km$^3$ (Jay & Widdowson, 2008; M. A. Richards et al., 2015) along with a significant volume offshore in the Arabian Sea (Gombos Jr et al., 1995; Calvès et al., 2011; D. Pandey et al., 2011; P. Kumar & Chaubey, 2019; Fainstein et al., 2019), and some small-volume Deccan-related intrusions in Seychelles (Devey & Stephens, 1991; Ganerød et al., 2011; T. M. Owen-Smith et al., 2013; Shellnutt et al., 2017). Estimates of the total pre-erosional DT lava flow volume range from 1 to 2 x 10$^6$ km$^3$ (Sukheswala, 1981; G. Sen, 2001; Jay & Widdowson, 2008).

Based on observed structural discontinuities as well as different geochemical and isotopic compositions of the lava flows, the Deccan Traps CFB is typically subdivided into four separate subprovinces each with potentially different eruptive history and corresponding magmatic system (Figure 1). These four subprovinces are the Western Ghat-Central Indian Volcanic province (WVP), the Mandla Lobe province, the Malwa plateau province (including the volcanic sequences in the Narmada-Tapti rift zone which may be a separate sub-province), and the Saurashtra-Kutchh province (Z. X. Peng et al., 2014; Kale et al., 2020). Among these, the WVP hosts some of the thickest continuous basalt flow sections for DT along the Western Ghats Escarpment (WGE) along the Western Coast of India (J. J. Mahoney, 1988). With a $\sim$ 3.5 km composite section of 10-50 m thick basalt flows emplaced atop Neoarchean basement (e.g., 1251 m thick Koyna core Mishra et al., 2017) easily logistically accessible, WGE sections have been extensively studied geochemically with detailed analysis of major and trace elements and isotopic composition (Sr, Pb, and Nd).

#### *3.1.1 Western Ghats Geochemical Formations*

On the basis of these measurements and volcanological features, the Western Ghats (WG) section has been grouped into three subgroups which are further subdivided into 12 geochemical formations each with several hundred meter thicknesses (Deshmukh, 1977; J. Mahoney et al., 1982; K. Cox & Hawkesworth, 1985; Basu, Saha-Yannopoulos, & Chakrabarty, 2020; Devey & Lightfoot, 1986; Beane et al., 1986; K. V. Subbarao, 1988; K. Subbarao et al., 1988; Beane & Hooper, 1988; Bodas et al., 1988; Khadri et al., 1988; Lightfoot, Hawkesworth, et al., 1990; C. Mitchell & Widdowson, 1991; Z. Peng et al., 1994; Choudhary & Jadhav, 2014; Hegde et al., 2014). The three subgroups of the WG (Wai, Lonavala, and Kalsubai, top to bottom) together represent more than 70 % of the total erupted volume of the DT and hence

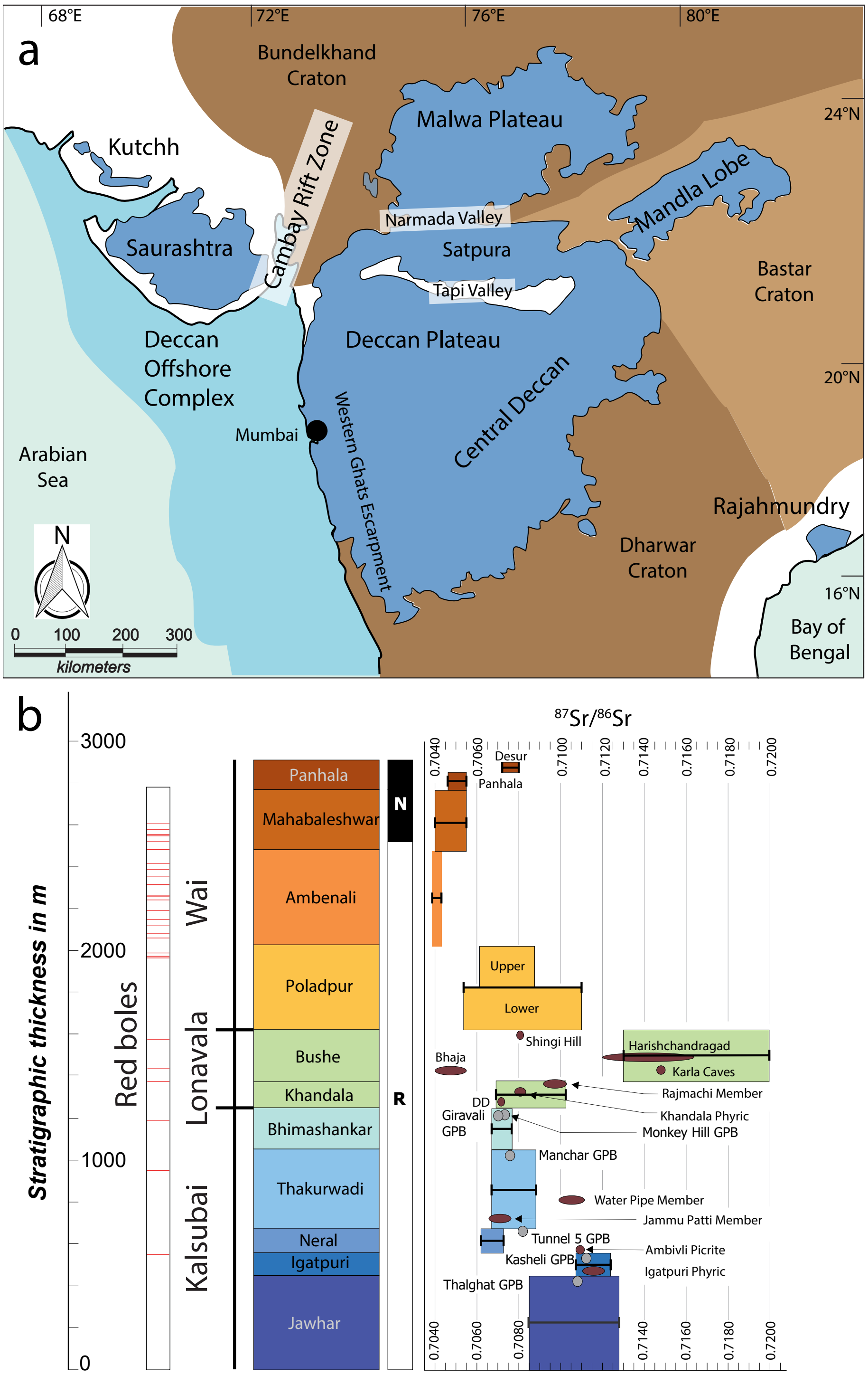
a
68°E
72°E
76°E
80°E
24°N
20°N
16°N
Bundelkhand Craton
Kutchh
Cambay Rift Zone
Malwa Plateau
Narmada Valley
Mandla Lobe
Saurashtra
Satpura
Tapi Valley
Bastar Craton
Deccan Offshore Complex
Deccan Plateau
Mumbai
Western Ghats Escarpment
Central Deccan
Arabian Sea
Rajahmundry
Dharwar Craton
Bay of Bengal
N
0 100 200 300
kilometers
b
Stratigraphic thickness in m
3000
2000
1000
0
Red boles
Wai
Lonavala
Kalsubai
Panhala
Mahabaleshwar
Ambenali
Poladpur
Bushe
Khandala
Bhimashankar
Thakurwadi
Neral
Igatpuri
Jawhar
N
R
87Sr/86Sr
0.7040
0.7060
0.7080
0.7100
0.7120
0.7140
0.7160
0.7180
0.7200
Desur
Panhala
Upper
Lower
Shingi Hill
Bhaja
Harishchandragad
Karla Caves
Rajmachi Member
DD
Giravali GPB
Khandala Phyric
Monkey Hill GPB
Manchar GPB
Water Pipe Member
Jammu Patti Member
Tunnel 5 GPB
Kasheli GPB
Ambivli Picrite
Igatpuri Phyric
Thalghat GPB

Figure 1 *(previous page)*: A) : Outline of the Deccan Traps Volcanic Province India including the offshore Deccan complex (based on data from Kale et al., (2019) and Jay & Widdowson (2008)). The whole subaerial Deccan province is subdivided into four main subprovinces : Western Ghat-Central Indian Volcanic province (WVP), the Mandla Lobe province, the Malwa plateau province (including the volcanic sequences in the Narmada-Tapti rift zone), and the Saurashtra-Kutchh province. The southern edge of the Malwa province is typically the Narmada Tapi zone though some of the lava flows in the Satpura region may have been sourced from the central Deccan region. The Saurashtra-Kutchh and the Malwa subprovinces are divided by the Cambay rift zone. B) : A composite stratigraphic sections for the Western Ghats region showing the chemostratigraphic formations, magnetostratigraphic with the chron C29N and C29R boundary, and the locations of red boles in the section (following Sprain et al., 2019; Chenet et al., 2008). We show the range of $^{87}Sr/^{86}Sr$ for each geochemical formation (using compilation from Vanderkluysen et al., 2011; Sheth et al., 2016). The gray ellipses show the values for some selected Giant Plagioclase basalts while the brown ellipses show the values of some other Central Deccan chemical types (Figure modified from Kale et al., 2019).

are the "main volumetric component" (M. A. Richards et al. (2015) and references therein).

The Kalsubai subgroup, consisting of the Jawhar, Igatpuri, Neral, Thakurwadi, and Bhimashanker formations, exhibits a large range of compositions ranging from picritic flows to evolved flows (Mg # <36, Beane et al. (1986)). The Lonavala subgroup, consisting of the Khandala and Bushe formations, includes flows in the Bushe formation with significant crustal assimilation. Finally, the Wai subgroup, comprising the Poladpur, Ambenali, Mahabaleshwar, Panhala, and Desur formations, are the most evolved lavas with regards to fractional crystallization. The Ambenali Formation records the lowest $^{87}Sr/^{86}Sr$ and the highest $\varepsilon$Nd values among the WG flows, indicative of minimal crustal assimilation (J. J. Mahoney, 1988; Lightfoot & Hawkesworth, 1988; Z. Peng et al., 1994). Geochemical analysis of lava flow sections in the Indian Deccan plateau has shown that the WG geochemical formations can potentially be extended 100s of km laterally into the central Deccan (e.g., the Khandala and Poladpur formations Z. Peng et al., 1998; Melluso et al., 2004), the south-eastern Deccan (e.g., the Poladpur, Ambenali, and Mahabaleshwar formations C. Mitchell & Widdowson, 1991; Jay & Widdowson, 2008; Kaotekwar et al., 2014), and the Rajamundry Traps (about 1000 km from the WG Baksi, 2001; Self et al., 2008; I. M. Fendley et al., 2020). These results illustrate that the DT was typically associated with individual lava flows 100s of km long, and hence large individual eruptive episodes of order 1000s of km$^3$.

In the other DT subprovinces, lava flow major and trace element compositions similar to the WG geochemical formations have been found. For example, Z. Peng et al. (1998) (also see J. J. Mahoney, 1988; J. Mahoney et al., 2000) found Poladpur formation type lava flows in a Malwa plateau province section about 1000 km from the WG Poladpur outcrops. However, these Poladpur-like lava flows have a very different Pb isotopic composition than the WG Poladpur geochemical formation. Thus, they could not be fed overland by the same magmatic plumbing system as the Poladpur flows. Instead, they must have a different crustal magmatic eruptive system interacting with different crustal assimilant (see discussion in the next section). There are similar examples of lava flows from the Saurashtra and Mandla lobe provinces, which have isotopic mismatches with WG formations despite similar major and trace element characteristics (e.g., Z. Peng et al., 2014; Vanderkluysen et al., 2011; H. Sheth et al., 2018). In addition to isotopic mismatches, there are also temporal differences between similar geochemical chemo-type lava flows across the different DT subprovinces (e.g., H. Sheth et al., 2018, and references therein). A particularly illustrative case is the presence of the Bushe chemo-type (both isotopic and major/trace element composition) in lava flows and dikes in Saurashtra (Melluso et al., 1995, 2006; H. C. Sheth et al., 2013), the Mandla Lobe (Shrivastava et al., 2014), the Malwa Plateau subprovinces (Z. Peng et al., 1998; P. Hooper & Subbarao, 1999; J. Mahoney et al., 2000; H. Sheth et al., 2004), and Seychelles Island dikes (Devey & Stephens, 1991). However, given the chronostratigraphic and volcanologic constraints, these flows can not represent the same Bushe Formation of the Western Ghats (See the cited references for details). As described in Section 3.2.2, these subprovinces erupt over a multiple Ma time-period with some provinces predating the WG sequence (e.g. Saurashtra, Malwa Plateau though some of the uppermost Malwa flows may be coveal with WG), while the other post-dating the WG Bushe flows (Mandla Lobe). In aggregate, these observations illustrate that the magmatic plumbing system for these subprovinces, at least the upper crustal transport network, is different.

Nevertheless, the repeated occurrence of lava chemically similar to the WG formations suggests a common sub-continental magma source and that specific petrologic processes occurred multiple times across the CFB magmatic system (Z. Peng et al., 1998; J. Mahoney et al., 2000; Haase et al., 2019). Because of the complexities in magmatic architecture across multiple DT subprovinces described above, in this study we principally focus on the most voluminous Western Ghat-Central Deccan province. There is a broad consensus that most of this province represents eruptions from a single magmatic plumbing system, barring some components of the Central Deccan Plateau in the Narmada-Tapi rift valley (See Fig.1, discussion in Kale et al., 2020).

#### *3.1.2 Magma composition and melt components*

The dominant lava flow composition in the Deccan Traps is tholeiitic ($> 95\%$ of the exposed area) with minor volumes of carbonatitic, felsic, and alkaline eruptive products and intrusions (e.g. , Krishnamurthy, 2020; Krishnamurthy & Cox, 1977, 1980; Devey & Cox, 1987; J. J. Mahoney, 1988; K. V. Subbarao, 1988; Z. Peng et al., 1994; Shrivastava & Pattanayak, 2002; Melluso et al., 2002; Jay & Widdowson, 2008; Talusani, 2010; A. Ray et al., 2010; Chandra et al., 2019; Prasanth et al., 2019; Sheikh et al., 2020, and references therein). Picritic lava flows are most abundant in the Saurashtra region and the northern part of the WG (Krishnamurthy & Cox, 1977; K. Cox & Hawkesworth, 1985; Beane & Hooper, 1988; Melluso et al., 1995; Sano et al., 2001). These olivine-phyric picritic lava flows are generally erupted in the early part of the lava flow sequence along with some alkali basalts (e.g., Krishnamurthy et al., 2000, and references therein). In the Western Ghats section, picritic basalts constitute less than 10% of the total volume and are mostly reported in the lower formations (Igatpuri, Neral, Thakurvadi, Bushe, and Poladpur Beane & Hooper, 1988; Krishnamurthy et al., 2000). The reduced occurrence of picritic flows in younger DT sections has been typically interpreted as a sign of a steady-state magmatic plumbing system with progressively more fractionation and ponding of picritic magmas (Krishnamurthy et al., 2000). The rhyolitic/silicic flows associated with Deccan are minor volume ($<$ a few percent of total Deccan volume) with limited present day exposures primarily in Saurashtra region (and adjoining parts of Narmada-Tapi rift zone and Barmer rift zone) (Sheikh et al., 2020; A. Pandey et al., 2017; Dolson et al., 2015).

The geochemical diversity of WG lava flows (and silic complexes) suggests a complex petrologic evolution, involving significant fractional crystallization of plagioclase, clinopyroxene, and olivine as well as crustal assimilation and magma mixing (J. Mahoney et al., 1982; J. J. Mahoney, 1984; K. G. Cox & Hawkesworth, 1984; Ganguly et al., 2014; K. Cox & Hawkesworth, 1985; J. J. Mahoney, 1988; Chatterjee & Bhattacharji, 2008; Basu, Chakrabarty, et al., 2020). The initial parental magma for the WG lava flows (the Ambenali Formation end-member) is a high-degree partial melt from the Réunion plume source along with some asthenospheric and lithospheric mantle contributions (Z. Peng et al., 1994; Z. Peng & Mahoney, 1995). K. G. Cox (1980) proposed that these high degree partial melts are initially picritic but undergo extensive fractional crystallization in the crustal system to form basalts with $< 8$ % MgO before surface eruptions (K. G. Cox & Hawkesworth, 1984; Herzberg & Gazel, 2009). Based on rare-earth element (REE) inversions (McKenzie & O'nions, 1991; S. Gibson & Geist, 2010), M. A. Richards et al. (2015) concluded that the top of the asthenospheric melt column was at about 60 km depth during the WG eruptive sequence.

Several different assimilates have been proposed for the DT lavas, including granitic upper crust (J. Mahoney et al., 1982; K. G. Cox & Hawkesworth, 1984; Lightfoot & Hawkesworth, 1988; Lightfoot, Hawkesworth, et al., 1990; Z. Peng et al., 1994; Dessai et al., 2008), lower granulitic continental crust (K. G. Cox & Hawkesworth, 1984; Z. Peng et al., 1994; Peters et al., 2017), and partial melting of subcontinental lithospheric mantle (Lightfoot & Hawkesworth, 1988) either individually or in sequence (Z. Peng et al., 1994). Some of the geochemical variations in lava flows have also been suggested to be indicative of partial melt from the lithospheric, asthenospheric, and deep plume components (e.g., J. J. Mahoney, 1988; K. V. Subbarao, 1988; Jennings et al., 2017; Hari, Swarnkar, & Prasanth, 2018; R. E. Ernst, 2014, and references therein). Furthermore, the crust beneath each subprovince is compositionally different due to the pre-Deccan structure of the Indian craton (See Figure 1). For instance, the WG region is underlain by continental crust similar to the Western Dharwar craton while the central DT is emplaced atop Archaen Eastern Dharwar Craton and the Bastar craton (Dessai & Vaselli, 1999; Y. B. Rao et al., 2017; Kale et al., 2017). In contrast, the Mandla subprovince is underlain by the Bastor craton and some Proterozoic mobile belts (J. S. Ray & Parthasarathy, 2019; Kale et al., 2020, and references therein) while the crust under the Malwa province includes the Bundelkhand craton, Proterozoic Vindhyan Supergroup (sandstone, shale, and carbonates), and a Late Archean-Proterozoic collisional mobile belt (R. Ray et al., 2008; Ramakrishnan & Vaidyanadhan, 2010; L. Ray et al., 2016). As a consequence, the isotopic and geochemical composition of the crustal contaminants in each of the DT subprovinces can vary significantly.

It is clear that the assimilates can vary rapidly between geochemical formations, despite their uncertain composition, or even their crustal vs. lithospheric–mantle nature (e.g., (J. J. Mahoney, 1988; Lightfoot, Hawkesworth, et al., 1990; Gallagher & Hawkesworth, 1992; Arndt et al., 1993; Hawkesworth & Gallagher, 1993; Z. Peng et al., 1994; Turner & Hawkesworth, 1995; Lassiter & DePaolo, 1997; H. Sheth & Chandrasekharam, 1997; Allegre et al., 1999; G. Sen, 2001; Chandrasekharam et al., 2000). For instance, the crustal assimilation component for the highly contaminated Bushe Formation is isotopically different from the assimilate for the Mahabaleshwar Formation (e.g., K. Cox & Hawkesworth, 1985; Beane et al., 1986; Lightfoot & Hawkesworth, 1988; Z. Peng et al., 1994; J. Mahoney et al., 2000; Gangopadhyay et al., 2003; Melluso et al., 2006). Similarly, the main WG geochemical formations show abrupt changes in $^{87}Sr/^{86}Sr$ ratios, sometimes across even a single sheet lobe. These changes are indicative of rapid changes in the source components for the erupted magma (Beane et al. (1986), Figure 1b modified from Kale et al. (2020), Haase et al. (2019) for Mandla Lobe). Hence, the magmatic plumbing system for the DT is complex, with multiple assimilates, but also dynamic, with the ability to rapidly switch magma compositions. The geochemical memory of the magmatic system is relatively small. This is difficult to reconcile with a magmatic plumbing system composed of only a few large well stirred magma chambers where each eruptive episode typically integrates magma compositions over multiple 100s of kyr (Black & Manga, 2017).

### 3.2 Deccan Traps Geochronology

The total DT eruptive sequence lasted about 10 Ma, with the oldest alkalic lava flows in Pakistan around 72-73 Ma (Khan et al., 1999; J. Mahoney et al., 2002), and the youngest lava flows around 60-61 Ma in the Mumbai sequence (H. C. Sheth et al., 2001a, 2001b; Pande, Yatheesh, & Sheth, 2017). However, the majority (> 90%) of volcanism occurred during a $\sim$ 1 Ma time-interval from 66.3 to 65.3 Ma (Renne et al., 2015; Schoene et al., 2015; Parisio et al., 2016; Sprain et al., 2019; Schoene et al., 2019). This interval includes almost all of the Western Ghats eruptive sequence and hence constitutes the "main phase" of DT volcanism (Sprain et al., 2019; Schoene et al., 2019).

In the following section, we first discuss the geochronological datasets from the Western Ghats, since they represent the primary volumetric component of the Deccan Traps and have the best constraints on eruptive tempo. We then briefly describe the available geochronological constraints for the DT subprovinces, with a focus on their implications for the overall magmatic activity of the DT flood basalt province.

#### *3.2.1 Western Ghats geochronology*

Our best direct constraints on the eruptive rates for the Deccan Traps comes from the high-precision $^{40}$Ar/$^{39}$Ar dating of lava flows and U-Pb dating of inter-flow horizons in the Western Ghats region (Renne et al., 2015; Schoene et al., 2015; Sprain et al., 2019; Schoene et al., 2019). Using their new ages with previously published high-precision dates, Sprain et al. (2019) concluded that the DT lava flows in the Western Ghats-Central Deccan province erupted continuously (within age precision) for a total duration of 0.991 Ma from $\sim$ 66.413 Ma (Jawahar Formation) to $\sim$ 65.422 Ma (Upper Mahabaleshwar Formation). This time period spans the Cretaceous-Paleogene Boundary (KPB; 66.052 Ma Sprain et al., 2018) with more than 75 % of the DT volume erupted within 650 kyr of the KPB. The zircon U-Pb dates in Schoene et al. (2019) are consistent with the $^{40}$ Ar/$^{39}$Ar results with respect to the overall duration of the WG although their exact location of the KPB within the WG section is slightly different.

However, these recent studies do differ with respect to eruptive tempo. Sprain et al. (2019) found no evidence for a long hiatus (> 150 kyr) between individual lava flows or geochemical formations, and the mean magma eruption rate for the Kalsubai and Lonavala (0.4 $\pm$ 0.1 km/year) subgroups is statistically similar to that of the Wai subgroup (0.6 $\pm$ 0.2 km/year). In contrast, Schoene et al. (2019) proposed multiple hiatuses within the WG stratigraphy: a $\sim$ 100 kyr hiatus between Poladpur and Ambenali, and a $\sim$ 250 kyr hiatus between Ambenali and Mahabaleshwar formations. Correspondingly, they conclude that the eruptive rates were also pulsed with the Ambenali and Poladpur Formations each having an eruptive pulse of 6-10 km$^3$/year with a $\sim$ 100 kyr duration.

Although ascertaining why the results of the two studies differ is an area of ongoing work, we can use volcanological features to assess if there is any evidence for the proposed long hiatuses. Since sub-continental India was located in the equatorial belt around 66 Ma with potentially high rainfall and weathering rates (e.g., Dessert et al., 2003; Johansson et al., 2018), a long hiatus would be expected to correspond to a thick weathering horizon in between basaltic flows. Indeed, multiple Western Ghats sections have red weathering horizons (red boles, typically $\leq$ 1 m thick) in between basaltic flows (e.g., stratigraphic sections in Jay, 2005, Steve Self personal comm.). However, there is no evidence of a stratigraphically continuous, extraordinarily thick red-bole between the Poladpur and Ambenali formations (Jay, 2005; Jay et al., 2009; Chenet et al., 2009). Furthermore, some of the Ambenali-Poladpur transition sections do not have any red boles (Jay et al., 2009; Chenet et al., 2009; Sprain et al., 2019, ;

also See Figure 1). Finally, red boles are frequently found between lava flows within the Ambenali formation, suggesting some hiatus time-period between individual eruptive episodes. Thus, we conclude that the eruptive tempo of the Western Ghats sequence does not show any evidence for multiple 100 kyr hiatuses between eruptive time-periods contrary to expectations of a large magma reservoir model (See Section 2.2.3).

#### *3.2.2 Constraints for other DT subprovinces*

Volcanism in the Saurashtra-Kutchh province (including some erupted products in Rajasthan) is generally considered to be the oldest in the Deccan stratigraphy. The largest silicic complexes in the DT - Barda Hills and Alech Hills in Saurashtra - and the overlying basalts in the region were emplaced between 69.5- 68.5 Ma based on the $^{40}Ar/^{39}Ar$ ages and field relationships (Dave, 1971; "Acid ring dykes and lava flows in Deccan trap basalt, Alech hills, Saurashtra, Gujarat", 1984; Shukla et al., 2001; A. Sen et al., 2012; Cucciniello et al., 2015, 2019). We would note that some recent U-Pb dates from the Alech, Barda, Girnar (also Ar-Ar), Rajula, and Phenai Mata intrusions suggest that these silicic-alkaline complexes were emplaced between 66.2 - 65.7 Ma, coincident with the main Deccan eruptive phase (Basu, Chakrabarty, et al., 2020; Sahoo et al., 2020). More work is needed to reconcile the various age constraints, and whether the different dating methods are sampling different parts of a long lived igneous complex (Basu, Chakrabarty, et al., 2020; Sahoo et al., 2020; Pande, Cucciniello, et al., 2017). Additionally, although $^{40}Ar/^{39}Ar$ dates on two alkaline complexes (Sarnu-Dandali and Mundwara) in the Cambay graben suggest a multi-phase emplacement spanning 90 to 60 Ma (H. Sheth et al., 2017; Pande, Cucciniello, et al., 2017), the oldest DT associated components have ages of $69.62 \pm 0.08$ Ma and $69.58 \pm 0.16$ Ma (Basu et al. (1993); recalculated to Renne et al. (2011) standards by Parisio et al. (2016)). Finally, there were subaerial/submarine eruptions on the Saurashtra Volcanic Platform between 75-68 Ma (Calvès et al., 2011; G. Bhattacharya & Yatheesh, 2015) and the Anjar Traps in Kutchh erupted between $67.47 \pm 0.30$ and $67.67 \pm 0.60$ (V. Courtillot et al. (2000); recalculated to Renne et al. (2011) standards by Parisio et al. (2016)). It is noteworthy that Central Saurashtra mafic dikes have $^{40}Ar/^{39}Ar$ ages spanning 66.6 Ma to 62.4 Ma (Cucciniello et al., 2015), younger than any published ages of the lava flows in the region. Additionally, Paul et al. (2008) used paleomagnetic data to conclude that magmatism in the Kutchh basin (dominantly tholeiitic, though with a significant alkaline component) occurred across chrons C30N, C29R, and C29N (i.e., across the KPB boundary) although the data do permit older ages. In summary, the available geochronological datasets suggest that the magmatic activity in the Saurashtra-Kutchh province is among the oldest in any DT province and may have had large time-gaps between periods of activity (G. Sen et al., 2009). Nevertheless, it may also have partially overlapped in time with the Western Ghat main phase volcanism.

For the Malwa plateau subprovince, Parisio et al. (2016) found that the alkaline and tholeiitic rocks were emplaced around the same time (within the age precision) with $^{40}Ar/^{39}Ar$ ages between $66.60 \pm 0.35$ to $65.25 \pm 0.29$ Ma for the Phenai Mata intrusive complex (also Basu et al. (1993) found similar ages) and $66.40 \pm 2.80$ to $64.90 \pm 0.80$ Ma for the Mount Pavagadh region. Based on paleomagnetic data as well as some $^{40}Ar/^{39}Ar$ dates (Schöbel et al., 2014, oldest date of $67.73 \pm 0.22$ Ma), the volcanism in the Malwa plateau may have been long-lived, spanning the chron C30N-C29R-C29N transitions.

In contrast to both the Saurashtra-Kutchh and Malwa Plateau provinces, the lava flows in the Mandla lobe are generally younger, with ages around $64.42 \pm 0.33$ Ma (Shrivastava et al., 2015; Pathak et al., 2017). However, some of the dikes in the region are older than published lava flow ages (e.g., $66.56 \pm 0.42$ Ma Shrivastava et al., 2017) suggesting the potential presence of older lava flows in the sequence. Some of the dikes in the eastern-most extension of the Mandla lobe (in the Damodar valley region) have Deccan associated mafic dikes with ages ranging from 70.5 to 65.5 Ma, potentially indicating a long duration magmatic activity in the Narmada-Tapi rift zone associated structures (Srivastava et al., 2020).

The youngest section in the DT is the Mumbai sequence in the WG (along with some intrusive dikes in Goa) with some subaqueous lavas (Duraiswami et al., 2019). The Mumbai sequence includes a much wider compositional diversity than the rest of the DT, containing tholeiitic basalts, rhyolites, trachytes, and pyroclastics (Melluso et al., 2002). The $^{40}Ar/^{39}Ar$ ages for this sequence are typically between 61.5 and 63 Ma (Widdowson et al., 2000; H. C. Sheth et al., 2014; Samant et al., 2019; Duraiswami et al., 2019) with some volcanism ranging from 64.55 Ma to 60.8 Ma (H. C. Sheth et al., 2001b, 2001a; P. Hooper et al., 2010; Basu, Chakrabarty, et al., 2020). The main phase of Laxmi Ridge-India-Seychelles continental rifting and voluminous offshore magmatism (Misra et al., 2014; Pande, Yatheesh, & Sheth, 2017; Samant et al., 2019) as well as intrusions in Seychelles (Shellnutt et al., 2017), also have an age around 62.5 Ma, suggesting a possible association.

In summary, there is a general progression of younger Deccan Trap ages moving southward, consistent with the northward motion of the Indian plate over a quasi-stationary mantle plume head. With a typical Indian plate motion of $\sim$ 15-20 cm/yr during Late Cretaceous-Early Paleogene (Cande & Patriat, 2015), the northward plate motion over 2 Ma is 300 - 400 km comparable to the distance between the Saurashtra region and the Mumbai-Pune region. However, the geochronology also suggests additional complexities, principally associated with the role of pre-existing tectonic features on the Indian sub-continent such as the Narmada-Tapi and the Cambay rift zone (See Figure 1). Petrologically, observations from each subprovince show that the silicic lavas and intrusions were typically emplaced after the local primary flood basalt sequence (e.g., Mount Pavagadh, Parisio et al. (2016); Mumbai : H. C. Sheth et al. (2014)). With a few exceptions, the alkaline rocks in the DT are generally emplaced both prior to and post the main phase of tholeiitic lava flows (e.g., Parisio et al., 2016), analogous to some other CFBs (e.g., S. Gibson et al., 2006; R. E. Ernst & Bell, 2010; R. B. Larsen et al., 2018). This is consistent with an initial lower degree melt that quickly traverses the crust, potentially with minimal assimilation (e.g., Malwa plateau, Haase et al., 2019). This is followed by an increasing degree of partial melting in the mantle plume head and more extensive fractionation (K. G. Cox & Mitchell, 1988). Eventually, the system reverts back to a lower degree, alkalic melt, along with some silicic melts, coincident with progressive plate motion away from the plume (M. A. Richards et al., 1989).

With regards to eruptive tempo, the subprovinces appear to typically have a much more extended period of volcanic activity compared to the Western Ghats sections. However, with the available geochronological constraints, it is unclear whether the eruptive activity is pulsed with long hiatuses in between or if the typical time between individual eruptive episodes is simply longer.

#### *3.2.3 Geochronological Constraints for other CFBs*

Most CFBs, including the DT, show evidence for a short duration $\sim$ 0.5-1 Ma long "main phase" of volcanic activity wherein the majority of the CFB volume is erupted (e.g., V. E. Courtillot & Renne, 2003; Bryan et al., 2010; R. E. Ernst & Youbi, 2017; Wilkinson et al., 2017; H. Svensen et al., 2018, and references therein). However, most of them lack a robust chemostratigraphic framework, or as abundant surface exposure of lava flows as the DT. Thus, at present, it is unclear based on geochronology alone whether the majority of eruptive episodes in other CFBs are further clustered into pulses within the main phase volcanism.

The two exceptions to this are the Columbia River Basalts (CRB) and the Siberian Traps. For the CRB, Kasbohm and Schoene (2018) used high precision zircon U-Pb geochronology on volcanic ash beds between basaltic flows to constrain the duration of the main volcanic phase ($> 95\%$ of CRB volume) to about 800 kyr (16.7 to 15.9 Ma). These results, in combination with the magnetostratigraphy, show no evidence for a $>$ 50-100 kyr hiatus between eruptions.

For the Siberian Traps, high-resolution U-Pb geochronology (S. D. Burgess & Bowring, 2015; S. Burgess et al., 2017) suggests that the volcanic sequence should be divided into three stages. More than two thirds of the extrusive volume was erupted in the first stage, which lasted 700 kyr (252.2 to 251.9 Ma). This was followed by a $\sim$ 420 kyr period of emplacement of mostly intrusives across the Siberian platform (also see Jerram et al., 2016; Augland et al., 2019), potentially due to the volcanic load of the overlying lava flows (S. Burgess et al., 2017). Finally, in stage 3, a combination of extrusive and intrusive volcanism occurred for another 300-500 kyr. As there are only four dates for the basaltic lavas (two in Stage 1, and one each in Stage 2 and 3), it is not possible to assess if there were any eruptive hiatuses within each stage of extrusive lava flows. The intrusions continued throughout Stage 2 without any resolvable hiatus ($>$ 100 kyr). It is important to note that the U-Pb dates are from intrusions emplaced within the lava flow stratigraphy or exposed on the periphery of lavas and volcaniclastic rock. This suggests that these intrusions were likely emplaced at shallow depths ($<$ 2 km, also see Tomshin et al. (2005, 2014) for petrographic data suggesting shallow emplacement). Consequently, we would argue that the Stage 1-Stage 2 transition from extrusive eruptions to shallow intrusive magmatic activity for $\sim$ 420 kyr does not represent a hiatus for the crustal magmatic system. In addition, the sill complex does not correspond to a single well-mixed large magma reservoir, as the intrusions have varying geochemical compositions and emplacement ages (e.g., N. A. Krivolutskaya et al., 2018, see Description of a magma reservoir in Section 2). Hence, the frequency of crustal diking will be controlled by the timescale of pressure buildup in the upper crustal magma reservoir, regardless of whether each dike feeds a shallow sill complex or eruptions at the surface.

In conclusion, there is no clear evidence of long ($>$ 100 kyr) eruptive (including shallow sill emplacement) hiatuses for either the Siberian Traps or Columbia River Basalts given the available datasets. These observations are inconsistent with the expectations of the large magma reservoir model (See Section 2.2.3).

### 3.3 Paleomagnetic constraints on eruptive tempo

Although there have been significant improvements in analytical techniques and consequent improvements in the precision of geochronological age precision to order 0.1 - 0.01%, the geochronological methods still have an absolute age uncertainity of order 10,000 kyr (for KPB age samples Sprain et al., 2019; Schoene et al., 2019). In order to obtain a higher resolution eruptive tempo, multiple studies have utilized the record of paleo-secular variation recorded in successive lava flows (directional groups, DGs) to constrain lava flow eruption rates at $\sim$ 100 yr resolution (Riisager et al., 2003; Knight et al., 2004; Chenet et al., 2008, 2009; V. Courtillot & Fluteau, 2014; Moulin et al., 2017; Pavlov et al., 2019). The Earth's magnetic field experiences secular variation, i.e. it naturally moves slightly over time. The paleomagnetic directions recorded in rocks accordingly display these variations in field position. Sets of lava flows with indistinguishable or very similar field directions (typically 5-10 $\circ$)are assigned to a directional group, which is commonly assumed to have been emplaced very rapidly (within $\sim$ 400 years) based on estimates of modern secular variation (Pavlov et al., 2019). The number and distribution of directional groups is thus an estimate of the temporal frequency of eruptions.

Based on a large study in the Western Ghats region, Chenet et al. (2009) concluded that the whole WG stratigraphy erupted as a combination of 30 major eruptive periods along with 41 additional individual lava units. Using the available geochemical stratigraphy, they estimated that each eruptive period had a volume of 1000 to 20,000 $km^3$ while individual lava flows were $\sim$ 1300 $km^3$ (based on analogy with Roza flow field in the Columbia River basalt (Thordarson & Self, 1998). Using similar secular variation timescale arguments as described in the previous section, they concluded that each of these 71 units had a duration of 10-100 years. In conclusion, Chenet et al. (2008, 2009) estimated that a very short active eruptive period of 1000 to 7000 years for the Western Ghats Deccan lava flows with the rest being hiatuses. If correct, this would suggest very pulsed eruption tempo consistent with the large magma reservoir model. However, the relationship between the correlation of paleomagnetic directions to eruptive timescales has been based on strongly simplifying assumptions about secular variation. In particular, the quasi-cyclic nature of paleosecular variation (with a time scale of order 10 kyr, (Panovska et al., 2019)) can introduce spurious correlations wherein lava flows separated by multiple secular variation cycles still have a small difference in paleomagnetic directions.

We addressed these challenges by developing a generalized forward modeling approach to compare synthetic eruptive histories with field datasets in a Bayesian inversion framework (Mittal et al., 2019). Since the real paleomagnetic field has a complex temporal structure with excursions and changing timescales for secular variation, we use a multi-million year long, low latitude deep sea sedimentary record of field variations for our forward model (Ohneiser et al., 2013). Additionally, we utilize the full statistical properties of the flow-by-flow records (e.g., fraction of lava flows in DGs and mean number of lava flows in DGs) instead of just the number of DGs to constrain permissible models. Using the same paleomagnetic dataset as Chenet et al. (2008, 2009), we find that the observations for the WG composite section (spanning Kalsubai, Lonavala, and Wai subgroups) is most consistent with continuous eruptions every 6,000-12,000 years (Figure 2). We find that the key characteristic of the "spurious" DGs is that each DGs only has 2-3 lava flow members. Additionally, a substantial fraction of lava flows from the full stratigraphy aren't part of a DG. When only considering DGs for a single physical stratigraphic section in the Western Ghats, the Deccan lavas satisfy both these characteristics with only $\sim$ 50% flows part of a unique DG in the Wai-Panchgani section spanning the Wai subgroup. Our new analysis results are further supported by the observation that many of the Deccan DGs as defined by Chenet et al. (2008, 2009) have weathering horizons (red boles) in between flows of a single DG from detailed stratigraphic logs (Jay, 2005). Since it is commonly assumed that 10cm - 1m red boles take at least a few 1000 years to form (Sheldon, 2003), a few hundred years time scale for emplacement of Deccan DG is unlikely. In conclusion, we find no evidence for a long ($>$ 50 kyr,Figure 2) eruptive hiatus in the Western Ghats region contrary to the large magma chamber model predictions.

### 3.4 Hg eruptive tempo estimates

Another indirect method to estimate the eruptive tempo of the DT at a 100-1000 yr resolution is the use of mercury (Hg) chemostratigraphy (e.g., Sial et al., 2013; Font et al., 2016; G. Keller et al., 2018; Percival et al., 2018; I. M. Fendley, Mittal, Sprain, et al., 2019). The primary source of Hg in the geological Hg cycle is emission from volcanism (See I. M. Fendley, Mittal, Sprain, et al. (2019) for more discussion). Hg, a highly [volatile??] species, is emitted as a vapor during individual LIP eruptions and potentially during passive degassing of LIP intrusives. Since Hg has a long atmospheric lifetime (order 6 month-2 years, Horowitz et al. (2017)) it can be globally distributed by atmospheric transport. Thus, large changes of the global Hg budget due to flood basalt eruptions will lead to increases in Hg concentration in sediments globally.

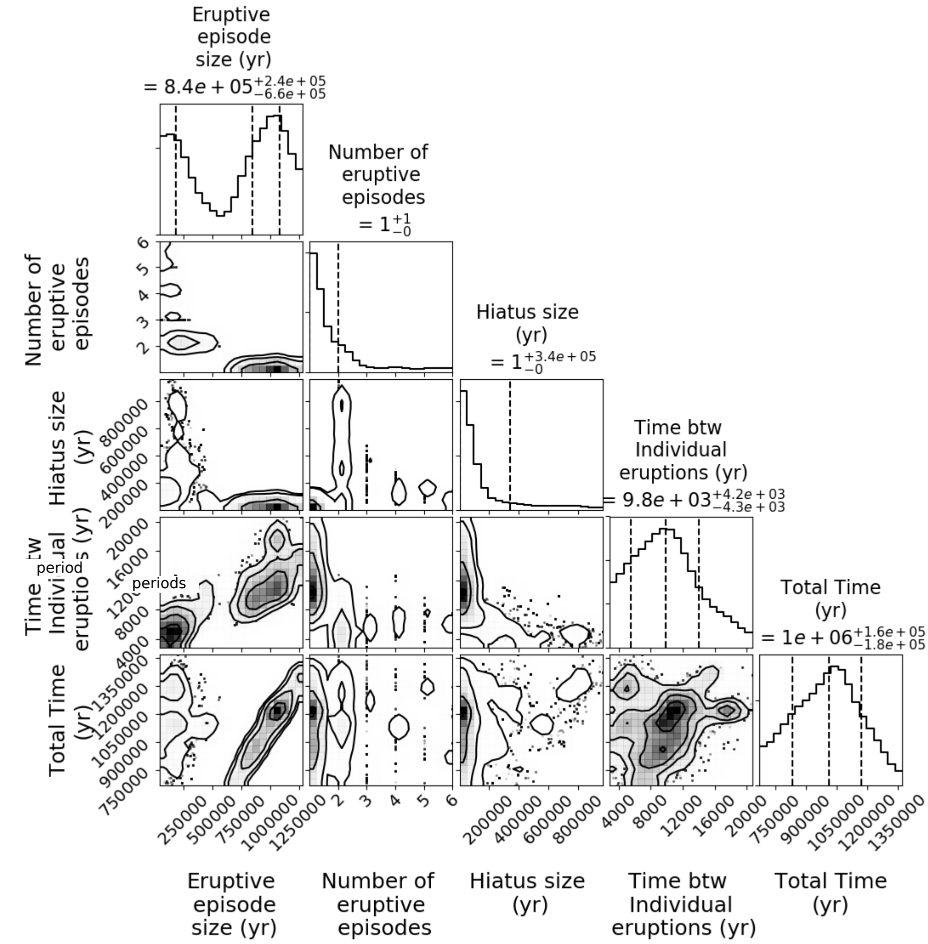


Figure 2: Bayesian parameter estimates for paleo-secular variation analysis for Eruptive period size (yr), Number of eruptive periods, Hiatus size (yr), Time between individual eruptive episodes (yr), and the Total time for the Composite Deccan Section. We use the flow-by-flow dataset from Chenet et al. (2008), and Chenet et al. (2009) for this analysis

However, to compare Hg concentration records from different locations and depositional environments, one must account for differences in sedimentation rate, sampling resolution, and sedimentary conditions such as organic carbon and sulfur concentration. I. M. Fendley, Mittal, Sprain, et al. (2019) used an environmental Hg cycle box model (Amos et al., 2013) to estimate eruptive rates and volumes by quantitatively evaluating the changes in the global mercury budget indicated by varying concentrations in sedimentary records. To do this analysis, they used a variety of eruptive rates and durations as inputs into the Hg cycle box model Amos et al. (2013). The inversion includes a time-smoothing step to account for sedimentation rate, bioturbation and diffusion, as well as the physical sample size. These synthetic model histories are then compared with the calculated Hg enrichment factors for many eruptive parameters with those from an actual sedimentary terrestrial Hg record (I. M. Fendley, Mittal, Renne, & Marvin-DiPasquale, 2019). For sedimentary Hg records from the time-period near the KPB boundary IRA-NV section ($\sim$ 40kyr before and $\sim$ 80kyr after the KPB boundary), I. M. Fendley, Mittal, Sprain, et al. (2019) find that the best-fit model parameters suggest eruptive episodes which lasted approximately 100 to 1000 years, occurred at least every 5-10 thousand years, and erupted around 40-240 cubic kilometers of basalt per year (I. M. Fendley, Mittal, Renne, & Marvin-DiPasquale, 2019). All of these estimates are averaged over the duration of the eruptive episode. Other longer sedimentary Hg records spanning the DT eruptive period also do not have pronounced hiatuses in peaks, after controlling for lithological effects and sampling resolution (Percival et al., 2018; G. Keller et al., 2018). Thus, to first order, Hg results do not find any support for a long ($>$ 50 kyr) hiatus throughout the "main-phase" DT volcanism, between $\sim$ 66.3 - 65.6 Ma.

We would emphasize that at present, the available Hg datasets do not have sufficiently uniform high resolution to reconstruct a time-varying eruptive history for the DT. Consequently, it is unclear if the IRA-NV section eruption rates are representative of the whole Wai subgroup, or just the Khandala-Bushe-Poladpur formations. Based on the available datasets (e.g., Percival et al., 2018; I. Fendley et al., 2019), individual eruptive episodes during the Ambenali formation had similar eruptions rates as the full DT main phase interval results, but with a longer eruption duration. However, more data are needed to confirm this preliminary conclusion. Nevertheless, we do not find any evidence for a significant increase in the time-period between eruptive intervals with the Hg dataset. This is in direct contrast to the model expectations for a larger magma reservoir wherein eruptions are expected to become less frequent during a CFB sequence.

It is also important to note that the Hg proxy integrates eruptions over the whole of the Deccan Traps (at any given time) instead of a single stratigraphic lava flow section. Thus, the results are not biased by local hiatus in a particular lava flow stratigraphic section. Based on modern CFB analogs, a $\sim$ 20 m thick lava flow may not cover the topography uniformly and completely (e.g., Thordarson & Self, 1993; Neal et al., 2019). Consequently, a single stratigraphic CFB section would probably not record all the lava flows in the region, especially for spatially extensive flows, e.g., Wai sub-group flows. We hence naturally expect that the eruptive tempo from paleo-secular variation measurements is slower than that from Hg. This process also helps explain the more frequent occurrence of weathering horizons (red boles) in the Wai subgroup flows vis-a-vis Kalsubai and Lonavala subgroups (Figure 1). In the field (WGE sections), most of the red boles are spatially heterogeneous and are, in most cases, not continuous across 10s of km. Our Hg eruptive rates are thus very compatible with results from geochronology (Section 3.2) and paleo-secular variation analysis (Section 3.3).

Our overall results are also consistent with the lava flow eruption rate estimates (10-200 $km^3$/year) based on flow morphology from Columbia River basalts (e.g., Vye-

Brown, Self, & Barry, 2013, Section 3.5). In conclusion, Hg chemostratigraphy does not provide any evidence for a prolonged province-wide eruptive hiatus, contrary to expectations from the large magma reservoir model (See Section 2.2.2).

### 3.5 Eruptive estimates based on lava flow morphology

Every CFB is comprised of 100-1000s of individual eruptive episodes, each of which is associated with a dominantly pahoehoe lava flow field (Self et al., 1998) with exceedingly rare fully developed a'a lava flows (Self et al., 1997; N. R. Bondre et al., 2004; Waichel et al., 2006; Sengupta & Ray, 2006; Passey & Bell, 2007; Duraiswami et al., 2008; Brown et al., 2011; El Hachimi et al., 2011; Vye-Brown, Self, & Barry, 2013; Braz Machado et al., 2015). Early studies of CFB flow fields proposed that individual flow fields were emplaced within a few days to weeks timescale with a correspondingly large eruption rate (Shaw & Swanson, 1970; Mangan et al., 1986; Tolan, 1989). However, extensive work on the lava flow fields in the Columbia River Basalt (Self et al., 1998; Vye-Brown, Self, & Barry, 2013, and references therein) has illustrated a strong similarity between CFB lava flows and modern Hawaiian and Icelandic lava flows despite the significantly different spatial scale. As a consequence, the CFB flows were emplaced as inflated compound pahoehoe flow fields with eruption rates similar to the highest observed for Iceland (e.g., Laki 1783; 8,700 $m^3/s$ Thordarson & Self, 1993; Guilbaud et al., 2005) and Hawaii (e.g., Kilauea 2018: 130-200 $m^3/s$ Neal et al. (2019), Mauna Loa a'a flow: 1,179-1,769 $m^3/s$ Finch and Macdonald (1953)), but sustained over years to decades (e.g., Ho & Cashman, 1997; Keszthelyi & Self, 1998; Self et al., 1998; Anderson et al., 1999; Bryan et al., 2010; Keszthelyi et al., 2006); See S. Reidel et al. (2018) for an alternative viewpoint).

This conclusion is supported by the observations that the typical thickness of individual flow lobes in the Deccan Traps is $\sim$ 15-20 m (Jay, 2005; Jay et al., 2009) with only a few thicker lobes (up to 100 m) in the Wai subgroup formation (Koyna Core Mishra et al., 2017). Although this lobe thickness is larger than the typical value for Hawaiian pahoehoe flow fields ($\sim$ 5 m lobes, HSDP2 core Katz & Cashman, 2003), the Laki 1783 eruptions are associated with 15-25 m thick basaltic flows Thordarson and Self (1993).

Finally, the lack of a'a flow fields in CFBs provides an upper limit for permissible effusion rate. Hawaiian a'a flow fields form when emplacement is rapid (days to weeks, average effusion rates $\geq$10 $m^3/s$ Rowland & Walker, 1990) with open lava channels near the fissure vents (e.g., Lipman & Banks, 1987; Wolff & Ramos, 2013). Since there is no clear evidence of such processes in most CFB flow fields (Self et al., 1998), we conclude that these flow fields (with much larger volumes than Hawaiian flows) must be emplaced over a long time-period (decades-centuries) rather than days-weeks. For the DT in particular, no clear lava tubes have been identified (Duraiswami et al., 2004; Pawar et al., 2015). However, the lack of unequivocally identified eruptive vents potentially allow for more frequent lava tubes and open channels. We also acknowledge that we do not preclude the possibility for some short-lived high effusion rate pulses that have been proposed to explain the rubbly pahoehoe structure in some CFB flows (N. R. Bondre et al., 2004; Duraiswami et al., 2008; B. Sen, 2017). In fact, Rader et al. (2017) have argued that effusive flux variations (of factor $\sim$ 2-5) are a necessary requirement for the formation of inflated sheet lobes. In both Columbia River Flood Basalts and Deccan Traps, the presence of multiple vesicle horizons in a single flow lobe may be evidence of this cyclic inflation (Hon et al., 1994; Cashman & Kauahikaua, 1997).

One potential challenge with using constraints on the effusion rate at the vent location to the estimate the duration of a CFB flow field is that each flow field may be fed by a long fissure system (e.g., $\sim$ 150 km for the CRB Rosa flow Self et al., 1998). If this fissure is active along a significant fraction of its length, the total time-period for a CFB emplacement will be short despite the restricted effusion rate. This hypothesis can be tested by using the thickness of a flow lobe's upper vesicular crust to constrain the duration of active inflation of the thickest flow lobe in a flow field. For the $\sim 1300$ km$^3$ Roza flow field in the CRB, Thordarson and Self (1998) estimated a minimum duration of 5 to 14 years (using the (Hon et al., 1994) $\sqrt{t}$ scaling for the thickness of the flow's upper crust). Using a similar method, Vye-Brown, Self, and Barry (2013) estimated similar minimum emplacement durations for other CRB flow fields: Palouse Falls (233 km$^3$, 19.3 years), Ginkgo (1570 km$^3$, 8.3 years), and Sand Hollow (2,660 km$^3$,16.9 years). These estimates are minimum estimates for the duration of a CFB field since they assume that the thickest flow was inflating throughout an eruption. If instead, different lobes undergo inflation at different times and/or have some time-hiatus between them, the total duration of a CFB field would be longer (order centuries) (See Vye-Brown, Self, and Barry (2013) for a more detailed discussion of other model uncertainties). Thus, we can conclude that individual CFB fields are emplaced over a long time-period ($\sim$ centuries) by sustained eruptions.

Another dataset supporting a multi-year to decadal-scale of CFB field emplacement is the low and high-temperature thermo-chronology around CRB dike segments ($\sim 1$ km long) from the Chief Joseph dike Swarm (H. L. Petcovic & Dufek, 2005; Karlstrom et al., 2019). The active duration of the $\sim 10$ m thick Maxwell dike is constrained to be 1.4–5.4 years (H. L. Petcovic & Dufek, 2005; Karlstrom et al., 2019) while the $\sim 10$ m Lee dike was active for 1.7 - 4.1 years with a long flow unsteadiness scale of 3.9 - 11.3 years (Karlstrom et al., 2019). Although more observations are needed to assess how representative these values are for other CRB dikes, as well as DT dikes, the general lack of melt-back in CRB dike swarm (Grunder & Taubeneck, 1997) suggests that analogous to the Laki 1783 eruption, different parts of the fissure may have been active at different times during an eruption (Thordarson & Self, 1993). Summarizing the various observations, we conclude the CFB flow fields were most likely emplaced over a decade to centuries with eruption rates comparable to the peak Laki 1783 eruption rate.

## 4 Geochemical characteristics of Deccan Flows - Implications for the magmatic system

We can indirectly constrain some of the characteristics of the DT magmatic architecture based on the stratigraphic pattern of geochemical variations in lava flows. Since the Deccan magmas underwent a complex sequence involving fractionation (potentially at multiple depths), assimilation of multiple components, and periodic replenishment of parental magmas before surface eruption, it is difficult to confidently ascertain the volume of cumulate material or their precise P-T histories and storage locations (See discussion in O'Hara and Herzberg (2002) and references therein). However, we can use stratigraphic changes in major, minor, and isotopic compositions to constrain the nature of the magmatic plumbing structure, especially whether an open system with constant recharge is required.

With regards to Deccan magma across all subprovinces, there is a general consensus that the AFC processes occur first in a lower crust reservoir(s) followed by shallow ($< 2$ kbar) fractionation and mixing (and in some cases assimilation) in upper crustal magma reservoir(s) (K. Cox & Devey, 1987; Devey & Cox, 1987; J. J. Mahoney, 1988; Hawkesworth et al., 1990; Lightfoot, Hawkesworth, et al., 1990; Z. Peng et al., 1994; J. Mahoney et al., 2000; G. Sen, 2001; Chatterjee & Bhattacharji, 2008; H. C. Sheth & Melluso, 2008; Natali et al., 2017). However, within with the general framework, is the magmatic plumbing system primarily a closed system with pulsed inputs of melts into the upper crustal system followed by closed system fractionation and assimilation?

### 4.1 Open System behavior

Firstly, many geochemical formations (e.g., Ambenali, Beane et al., 1986) as well various individual sections in other DT provinces (e.g., Malwa Plateau, Haase et al., 2019) typically have a restricted major and minor element composition. This observation can be explained by either of the two-model end-members: a quasi-steady state open magmatic system with reservoirs undergoing Recharge, Tapping via eruptions, Fractionation, and Assimilation (RTF, O'Hara & Mathews, 1981; Leeman & Hawkesworth, 1986; K. Cox, 1988; Arndt et al., 1993), or a thermally buffered, quasi-closed magma reservoir undergoing minimal chemical evolution (as suggested in Black & Manga, 2017). Similarly, continuous stratigraphic changes (e.g., from Poladpur to Ambenali) in ratios between incompatible elements (e.g., Ba and Zr) and isotopes can be explained by either of the scenarios since the observation-only requires some form of mixing between different magma bodies (K. Cox & Hawkesworth, 1985). However, if the system undergoes a sudden change in lava composition (e.g., Bushe to Poladpur), a model with large lower and upper crustal reservoirs will have challenges since it is difficult to change bulk compositions with a large volume of pre-existing melt without drastic amounts of contamination. Instead, it is easier to explain these transitions in a magmatic system with multiple chambers coupled with changes in lower crustal magma compositions (K. Cox & Hawkesworth, 1985).

Finally, the Black and Manga (2017) style model predicts an increasing contribution from crustal assimilation with increasing thermal maturity of the upper crust. In contrast, the least contaminated DT lava flows are the Ambenali formation, which erupts towards the end of the sequence. We note that over time, the increasing size of the crustal magma reservoir due to the addition of lower crustal melt (Black & Manga, 2017) will partly modulate this effect. However, without quantitative calculations, it is unclear how much this will buffer compositions. For the DT, multiple authors have proposed that minimal crustal assimilation of the Ambenali flows (even though they have undergone extensive fractionation) is a consequence of geochemical buffering due to plating of the reservoir by previous solidification zones (e.g., J. J. Mahoney, 1988; Devey & Cox, 1987). If this process were to be relevant for the large magma reservoir also, it could explain the Ambenali formation composition within that model context. However, it will also further increase the hiatus time between eruptive episodes since the crustal volatile source will be removed unless the parental magma has sufficient volatile content to be buoyant.

### 4.2 Intra-geochemical formation variability

Another avenue to distinguish among the two model end-members is to use the small but resolvable geochemical (e.g., Sr isotopes, Ni and Ti, Cr and Zr, Mg and Fe; J. J. Mahoney, 1988, and references therein) and petrological variations (e.g., phenocryst fraction and modal percentages of ol, plag, and cpx; Beane et al., 1986; Pattanayak & Shrivastava, 1999; G. Sen, 2001; Krishnamurthy, 2020; Basu, Saha-Yannopoulos, & Chakrabarty, 2020) within a geochemical formation. As illustrated by almost all sections in WG, the stratigraphic variation in a formation is not smooth even after accounting for possible biases due to surface alteration and lava flow fractionation (J. J. Mahoney, 1988; K. Cox & Hawkesworth, 1985; Krishnamurthy, 2020; Basu, Saha-Yannopoulos, & Chakrabarty, 2020). In Figure 3, we show a few representative sections from the Deccan Traps illustrating this behavior. In Figure 3a, we show the $^{87}Sr/^{86}Sr$ variation for a composite WG section with geochemical formations from Neral to Mahabaleshwar as well as an inset with just Ambenali, Mahabaleshwar, Panhala formations (Beane et al., 1986; Lightfoot & Hawkesworth, 1988). Figure 3b shows the dataset for Zr in a 480 m section spanning almost the whole Ambenali Formation section (the Torna hill-fort Devey & Lightfoot, 1986). Both these datasets clearly illustrate quasi-oscillatory variations within a single geochemical formation.

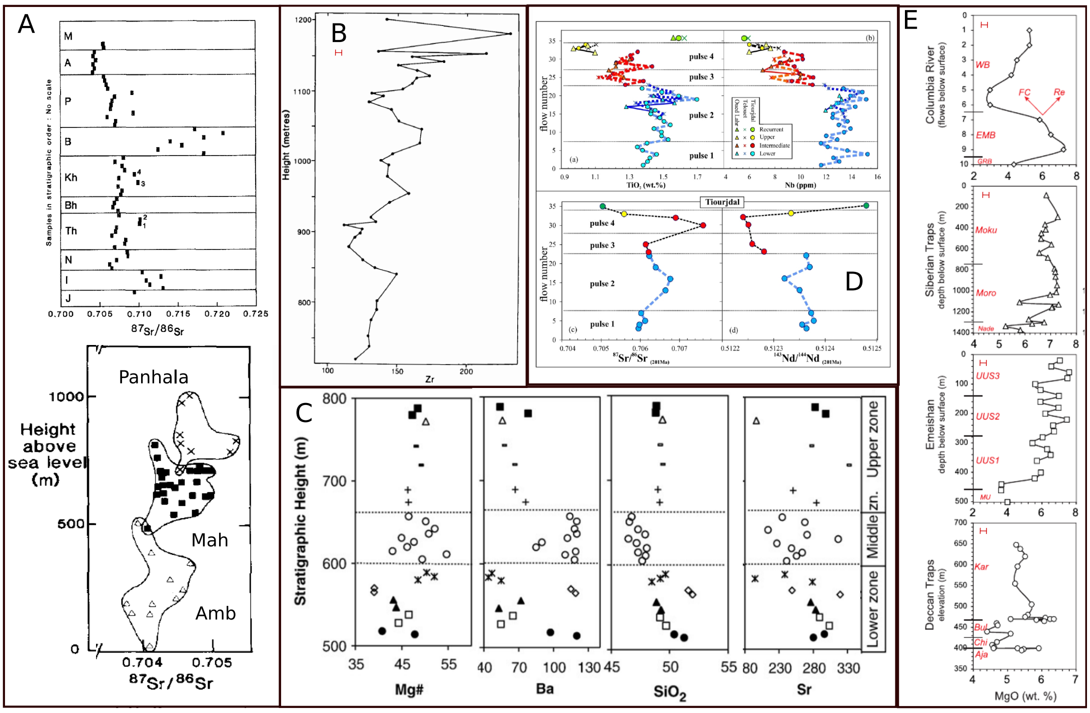

Figure 3: Representative sections showing geochemical variations within a stratigraphic section for Deccan Traps and some other continental flood basalts. A) $^{87}Sr/^{86}Sr$ variation for a composite WG section with geochemical formations from Neral to Mahabaleshwa(from Beane et al., 1986) with a zoomed inset showing the variations for the Ambenali, Mahabaleshwar, Panhala formations (from Lightfoot & Hawkesworth, 1988). B) Variation in Zr in a 480 m section spanning almost the whole Ambenali Formation section (the Torna hill-fort, Devey and Lightfoot (1986); K. Cox (1988)). C) Geochemical data from the Bhir section located about 300 km east of WGE in the Central Deccan province (from Talusani, 2010) with alkalic middle flow (this section has been correlated with the Ambenali formation). D) Geochemical datasets from the CAMP sections from Morrocco. The Nb (ppm), $TiO_2$ (wt %) data is shown for three sections from the Central High Atlas Mountains - Tiourjdal, Telouet, Oued Lahr while the isotopic dataset are from Tiourjdal section only. The Figure also shows the 4 eruptive pulses defined based on classical Directional Group analysis by Knight et al. (2004)(from Marzoli et al., 2019) E) MgO (wt %) for stratigraphic sections for stratigraphic sections for various flood basalts (from Yu et al., 2015). The dataset for Columbia River basalt include Eckler Mountain Basalt (EMB) and Wanapum Basalt (WB) flows. The Deccan Traps section is from K. V. Kumar et al. (2010) showing the Ajanta (Aja), Buldhana (Bul), Chikhli (Chi) formations (correlative to the Poladpur formation), and Karanja (Kar) formation (correlative to the Ambenali formation). The Yangliuping basaltic sequence from Emeishan large igneous province has been divided into four units by Song et al. (2006). Finally, the Siberian Traps section consists of three formations Mokulaevky (Moku), Morongovsky (Moro), and Nadezhdinsky (Nade) (Lightfoot, Hawkesworth, et al., 1990). The FC and Re arrows indicate the effect of Fractional Crystallization and magma recharge in an REAFC (REcharge Assimilation Fractional Crystallization) model. The typical error bar for each panel is of the size of the symbol unless otherwise specified. All of these datasets clearly illustrate quasi-oscillatory variations within a single geochemical formation across CFBs.

As another example, we show the geochemical data for the 282 m Bhir section located about 300 km east of WGE in the Central Deccan province Talusani (2010) in Figure 3c. Based on the trace element ratios (e.g., Zr/Nb, Ba/Y) most diagnostic of different WG geochemical formations as well as regional stratigraphy (Jay & Widdowson, 2008), the Bhir section lavas have Ambenali-type geochemistry (Talusani, 2010). However, within this section (10 flows), the middle flow is alkalic with a very different composition compared to the other tholeiitic basalt flows in the section. Additionally, the Bhir basalts show a wide range of chemical compositions for incompatible elements (e.g., Nb, Zr, Y, and Rare Earth Elements) between individual flows as well as within a single flow (e.g., the middle alkalic flow). Since the difference between alkalic and tholeiitic basalt is explained as a different degree of partial melt in the mantle, the presence of the flow in the same geochemical formation is indicative of a magmatic plumbing system that does not homogenize the melt in a large magma reservoir (Talusani, 2010).

We show another Central Deccan Trap stratigraphic section in Figure 3e (lowermost panel) where the local geochemical formations are correlative to the Ambenali and Poladpur formations (K. V. Kumar et al., 2010). Similar to the other examples, there is clear evidence of oscillations in MgO concentration within the same formation. We would emphasize that these examples are not unique and are instead representative of a large number of DT studies both within WGE as well as other subprovinces (e.g., K. Cox & Hawkesworth, 1985; Devey & Cox, 1987; Devey & Lightfoot, 1986; Lightfoot, Hawkesworth, et al., 1990; Z. Peng et al., 1998; J. Mahoney et al., 2000; Sengupta & Ray, 2011; S. K. Bhattacharya et al., 2013; Z. X. Peng et al., 2014; Haase et al., 2019). This conclusion is also illustrated by the fact that each geochemical formation typically traces out an extended region (much larger than analytical uncertainty) in the Sr, Pb, and Nd isotope parameter space instead of a single point (e.g., Vanderkluysen et al., 2011).

Our conclusion of continuous magma recharge and mixing during the eruption of a geochemical formation is further supported by the common occurrence of various zoning patterns (normal, oscillatory, reverse, complex) in plagioclase and clinopyroxenes phenocrysts in lavas. These zoning patterns are indicative quasi-continuous mixing between primitive and evolved magmas (Lightfoot et al., 1987; Pattanayak & Shrivastava, 1999; N. R. Bondre et al., 2004; Talusani, 2010). Similar to other geochemical variations, both the type of zoning patterns as well as phenocryst mineral composition (e.g., anorthite content of plagioclase) show clear changes in a stratigraphic section but without any smooth pattern (Lightfoot, Hawkesworth, et al., 1990).

### 4.3 Geochemical variability - Model

K. Cox (1988) proposed that these kinds of variations can be explained using a RTF-type magma reservoir with stochastic melt influx (z, in units of initial magma reservoir mass/model cycle), eruption rate(y, in units of magma reservoir mass/model cycle), and crystallization rates (x, in units of magma reservoir mass/model cycle) around the steady-state values. K. Cox (1988) showed that a WG Ambenali section (Figure 3c) could be statistically reproduced by the model with $x/y \sim 1$, z random chosen between 0-5, and $x \sim 0.05$ along with a slow increase in x up-section (to produce the long term trend). Each lava flow is considered the output of one model cycle and the lava's trace element composition oscillates around a steady-state value. Although this is a reasonable model to explain the data, it requires that a significant mass influx (up to five times the original magma reservoir mass) into the system throughout the eruption of a geochemical formation (i.e. an open system). This is extremely difficult to do for a large magma reservoir, especially the upper crustal reservoir where the expected influx is expected to be large ($z \sim$ 1-3) but only occur a few times at most during the eruption of a geochemical cycle (Black & Manga, 2017).

In order to explore a broader range of values for z (and corresponding values of x and y) can explain the geochemical datasets, we use a modified version of the Recharge-Eruption-Fractional crystallization model (ignoring assimilation to be consistent with K. Cox, 1988). Although this is not as thermodynamically rigorous as the Magma Chamber Simulator (e.g., Heinonen et al., 2020), this model captures the first order behavior of interest here. Following C.-T. A. Lee et al. (2014); Yu et al. (2015), and K. Cox (1988), the rate of change of the total mass $M$ and elemental mass $m_{ch}$ of a magma reservoir per eruptive cycle ($dN_c$) is :

$$\frac{dM}{dN_c} = \frac{dM_{re}}{dN_c} + \frac{dM_e}{dN_c} + \frac{dM_{cc}}{dN_c} + \frac{dM_x}{dN_c} \tag{1}$$

$$dm_{ch} = dM_x DC_{ch} + dM_e C_{ch} + dM_{cc} C_{cc} + dM_{re} C_{re} \tag{2}$$

$$\frac{dM_{re}}{dN_c} = z * M_{initial} \tag{3}$$

$$\frac{dM_e}{dN_c} = y * M \tag{4}$$

$$\frac{dM_x}{dN_c} = x * M \tag{5}$$

Here, the magma reservoir mass changes due to eruption ($dM_e$, negative), assimilation ($dM_{cc}$, positive), recharge ($dM_{re}$, positive), and fractional crystallization ($dM_x$, negative) during an eruptive cycle ($dN_c$). For the elemental mass balance term, $C_{ch}$ is the element's concentration in the magma reservoir, $C_{cc}$ is the wall-rock composition, and $C_{re}$ is the recharge composition. $D$ is the equilibrium partition coefficient between crystals and melt during fractional crystallization. For our calculations here, we set $dM_{cc} = 0$ to prevent crustal assimilation and x/y =1. Following Yu et al. (2015) and K. Cox (1988), we set bulk partition coefficients for Mg and Zr as 2 and 0.001 respectively. Since the actual composition of the magma recharge is not known, we assume a primitive magma composition with MgO = 9 wt % and Zr = 85 ppm (see K. Cox, 1988; J. J. Mahoney, 1988). We would emphasize our focus with these calculations is to look at broad features and hence the exact parameter choices are not important. For the starting magma composition in the magma reservoir, we set MgO = 5.8 wt % and Zr = 160 ppm, a range typical for the Ambenali formation (Beane et al., 1986; K. Cox, 1988). We randomly choose the value of x,y,and z between 0 and the maximal values $x^m$,$y^m$,and $z^m$.

In Figure 4, we show the results of the calculations for three sets of values : $x^m = y^m$ : 0.03 (red), 0.02 (black), 0.01 (blue), and $z^m$ : 0.06 (red), 0.04 (black), 0.02 (blue). The values have been chosen such that the magma reservoir is in quasi-steady state over 100s of eruptive cycles. The key result from this analysis is that sufficiently large values of $z^m$ (and corresponding $x^m$ and $y^m$) $\sim 0.06$ are required to produce the amplitude of variations in Figure 3b and 3e (ignoring any long term trends). Although the precise threshold value will be dependent on the exact parameter choice for recharge magma, the result that recharge must be at least a few percentages of the reservoir size is a robust conclusion. In Figure 5A and 5B, we show model results where we have set either recharge (z, Figure 5A) or crystallization (x, Figure 5B) to be zero. We set the other parameters at $x^m = y^m = 0.03$ and $z^m = 0.06$ with the three lines in Figure 5A and 5B showing three model realizations each with the same input parameters. These results clearly illustrate that without all three processes in REFC, we do not get the oscillatory pattern of geochemical variation but instead only a persistent trend.

Based on the model results, there are two primary requirements for explaining the observations: a) a combination of fractional crystallization, recharge, and eruption together in a RTF/REAFC (or a variant of these, see O'Hara & Herzberg, 2002) type magma reservoir, b) a magma recharge volume at least a few percent of the reservoir mass.

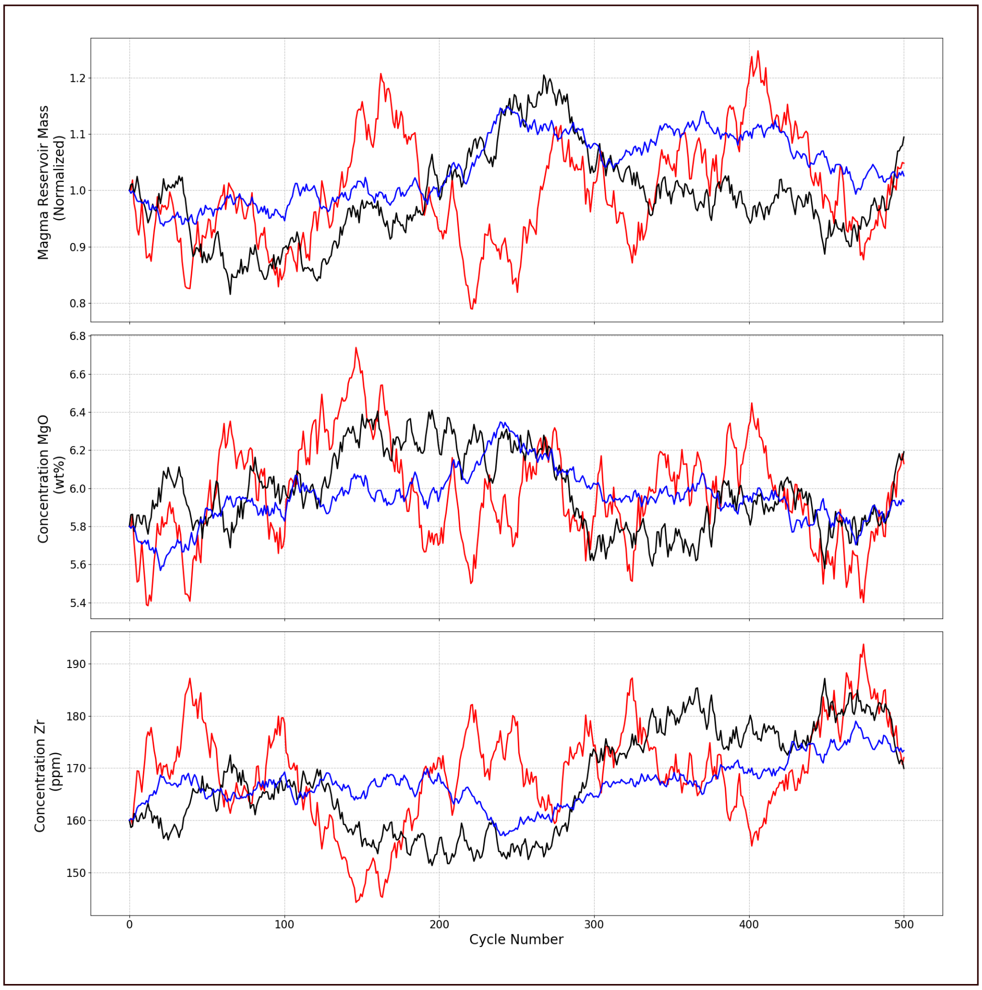


Figure 4: Results of REAFC model for Normalized magma reservoir mass, MgO (wt %), and Zr concentration (ppm). Calculations for three sets of values : $x^m = y^m$ : 0.03 (red), 0.02 (black), 0.01 (blue), and $z^m$ : 0.06 (red), 0.04 (black), 0.02 (blue) with stochastic variation around the mean value.

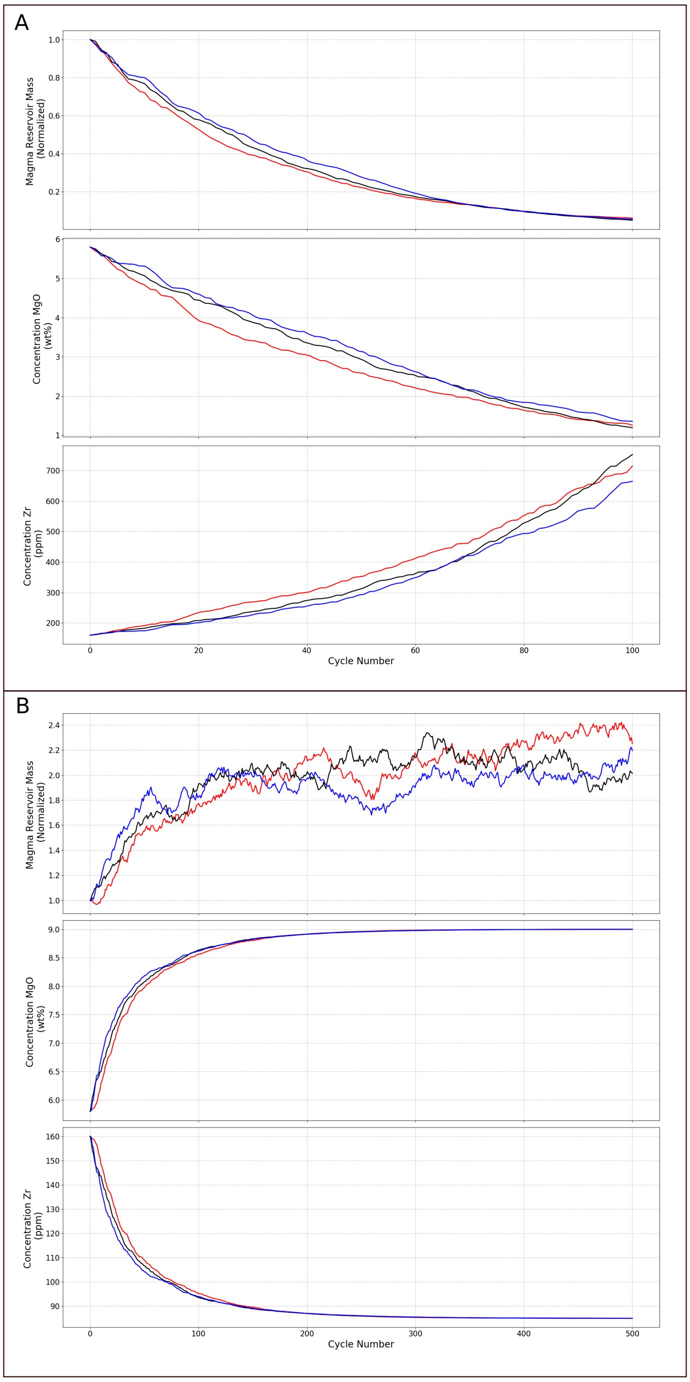


Figure 5: Results of REAFC model for Normalized magma reservoir mass, MgO (wt %), and Zr concentration (ppm). A) Calculations with no recharge (z = 0) B) Calculations with no crystallization (x = 0). We set the other parameters at $x^m = y^m = 0.03$ and $z^m = 0.06$ with the three lines each panel showing three model realizations each with the same input parameters.

### 4.4 Intra-Flow compositional variations

Although typically individual lava flows in a CFB are considered to be geochemical homogeneous, a few studies have illustrated that this may not always be the case (e.g., Philpotts, 1998; S. P. Reidel, 2005, 1998; N. R. Bondre & Hart, 2008); Passmore et al. (2012) for Laki 1783 eruption). The results of Vye-Brown, Gannoun, et al. (2013) and Vye-Brown et al. (2018) provide especially clear examples of the geochemical variations in products of a single eruption in the Columbia River Basalts.

Through the detailed mapping of the Sand Hollow ($\sim$ 2660 km$^3$) and Palouse Falls ($\sim$ 233 km$^3$) flow fields (Vye-Brown, Self, & Barry, 2013), these studies ensured that any observed variations were within a single eruption unit and not due to stratigraphic errors. In both cases (Sand Hallow and Palouse Falls), there is clear geochemical heterogeneity (major and minor elements) within both a single flow lobe, as well as laterally between different flow lobes of a single flow field. Based on the pattern of change across multiple elements, these studies concluded that either chemical weathering or post-emplacement fractionation could not have produced the observed variations. Instead, the intra-lobe and inter-lobe variations are indicative of heterogeneity within the magmatic system feeding the eruption. This conclusion is further supported by variations in $^{187}$Os/$^{188}$Os isotopes across a 35 m Sand Hallow flow lobe that require different amounts of crustal assimilation in the erupted melt (Vye-Brown, Gannoun, et al., 2013). These observations are inconsistent with a compositionally homogeneous magma reservoir feeding an individual eruptive episode (See Section 2.2.3). Instead, each eruptive episode is likely sourced from a magmatic system consists of a network of magma reservoirs with more primitive compositions over time (potentially from deeper magma bodies) (Vye-Brown et al., 2018). A counter example of this intra-flow geochemical hetereogeneity is a $\sim$ 2000 km$^3$ Nadayansky flow in Siberian Traps. Over the flow's few hundred km length, the lava major and trace element composition remains constant, within the analytical uncertainties (N. Krivolutskaya & Kedrovskaya, 2020).

Nevertheless, with the limited number of studies at present, it is unclear how common (or uncommon) intra-flow variations are for CFBs in general, especially isotopic variations which are less sensitive to post emplacement alteration and fractionation. In particular, no systematic stratigraphically controlled analysis of intra-flow variations has been done for lava flows in the Deccan Traps. This is partially due to the challenge of carefully mapping single flow fields over 10s-100s of km physically (Patil et al., 2020; Dole et al., 2020). Nevertheless, there are some potential examples from DT suggestive of geochemical differences within a single eruptive unit. For instance, K. Cox and Hawkesworth (1985) noted that the Kelghar Mafic Unit in the Mahabaleshwar Formation is restricted to only a single WG section. However, less mafic, but geochemically related Mahabaleshwar flows, are present at the same stratigraphic height in sections less than 40 km [away??]. Thus, one potential explanation for the Kelghar Mafic Unit is that it represents a small magma batch within the eruption of a larger, spatially extensive flow field. A similar interpretation has also been proposed for the DT lava flows with large concentrations of multi-cm long plagioclase phenocrysts - giant phenocryst basalts (GPBs P. Hooper, 1988b; Higgins & Chandrasekharam, 2007; H. Sheth, 2016). These flows occur at multiple stratigraphic heights within the WG sections though they are more common during the Kalsubai subgroup (e.g., Beane et al., 1986). They are also found in some of the other Deccan subprovinces (e.g., Talusani, 2012; Alexander & Purohit, 2019). Since GPBs are sometimes not continuous within a single physical eruptive unit, this suggests intra-flow variation within a single eruptive epsiode (Paul Renne and Steve Self, personal comm.).

Other potential evidence for geochemical variations within an eruptive event comes from compositional (including isotopic) differences along a single dike (N. Bondre et al., 2006; R. Ray et al., 2007; H. C. Sheth et al., 2009, 2013; Vanderkluysen et al., 2011; Cucciniello et al., 2015; H. Sheth et al., 2018). Additionally, closely spaced dikes in a dike swarm don't always have the same chemical compositions (N. Bondre et al., 2006; H. Sheth et al., 2019).

Finally, a substantial fraction of dikes in the Deccan Traps have multiple columnar jointed layers due to several magma injections (H. C. Sheth & Cañón-Tapia, 2015, and references therein). Frequently, each of these layers has a different geochemical composition indicative of a different batch of magma (H. C. Sheth & Cañón-Tapia, 2015; Cucciniello et al., 2015; Gadgil et al., 2019). A typical time-period for the emplacement of these composite dikes can be estimated based on the requirement that dikes need to cool sufficiently between each injection to form and preserve layered columnar joints. Using this principle, H. C. Sheth and Cañón-Tapia (2015) estimated that a multiple large dike (20-m-thick with ten columnar rows) would be emplaced over a minimum of several years and more likely up to a 100 years after accounting for latent heat of crystallization as well as the increasing country-rock temperature after emplacement of each dike segment (Evelyne et al., 2018). We readily acknowledge that there are counter-examples to this with DT dikes that are compositionally homogeneous over 10s of km (e.g., Vanderkluysen et al. (2011), see Kalsbeek and Taylor (1986); R. E. Ernst (2014); Buchan and Ernst (2019) for examples from other CFBs) and that some of the compositional differences may be due to dike-wall rock interactions. With existing literature, we cannot assess how ubiquitous intra-dike geochemical variations are. Still, observations from even a few dikes, along with physical evidence for multiple injections, suggests that a homogeneous, well-mixed magma reservoir does not always feed individual eruptive episodes. Additionally, the presence of multiple dikes in Deccan Traps hints at a long duration ($\sim$ 100 years) for each eruptive episode consistent with other datasets described in this section.

### 4.5 Geochemical variability - Implications for magmatic architecture

Considering the observations and the model results, we contend that it is very difficult to explain these with a large magma reservoir end-member. Firstly, this model predicts that a single geochemical formation with 10s of eruptive episodes is erupted very rapidly (order 100s to a few 1000 yrs, Black & Manga, 2017). Hence, there is generally not enough time-period between individual eruptive episodes for significant crystal fractionation or assimilation to change magma composition based on diffusion and crystal growth time-scales (Borges et al., 2014; H. Sheth, 2016). The large compositional inertia of the system due to the large magma volume makes this even more challenging. Secondly, the oscillatory stratigraphic, geochemical changes require an influx of some primitive melt input along with fractionation based on our calculations. The recharge into the crustal magma reservoir needs to quasi-continuous, which is contrary to the model results for a large magma reservoir model wherein there are generally very large but very sparse magma inputs for upper crustal reservoirs (Black & Manga, 2017, See description in Section 2.2.2, 2.2.3). The observations are best explained by a smaller (order 5 %) continuous melt input into the magma reservoir (s). Thus, we conclude that the pattern of geochemical variations strongly suggests that the large magma reservoir end-member is not realistic. Instead, these variations are a natural outcome in a RTF/REAFC model with one or multiple magma reservoirs (K. Cox & Hawkesworth, 1985; Yu et al., 2015).

### 4.6 Observations from other CFBs

The stratigraphic section in many other CFBs shows the same characteristic geochemical variability features as described above for DT (inter and intra-geochemical formation variations). In Figure 3d and 4e, we show a few example cases for the Central Atlantic Magmatic Province (Figure 3d, CAMP Marzoli et al., 2019), Siberian Traps (Lightfoot, Naldrett, et al., 1990), Columbia River Basalts (P. Hooper, 1988a;

P. R. Hooper, 2000), and Emeishan CFB (all Figure 3e, Song et al., 2006). All of these cases have the same oscillatory features within a single geochemical formation as in DT. Many similar examples have been described for most CFBs: CAMP (e.g., Tegner et al., 2019; Marzoli et al., 2019, and references therein) , Karoo-Ferrar CFB (Fleming et al., 1992; D. Elliot et al., 1995; Fleming, 1995; J. Marsh et al., 1997; Luttinen & Furnes, 2000), Siberian Traps (Lightfoot, Naldrett, et al., 1990; Fedorenko et al., 1996; Reichow et al., 2005; N. A. Krivolutskaya & Sobolev, 2016; N. Krivolutskaya et al., 2018; N. A. Krivolutskaya et al., 2018), North Atlantic Magmatic Province (NAMP, Andreasen et al., 2004; Peate et al., 2008; L. M. Larsen & Pedersen, 2009; Millett et al., 2016, 2017), Parana-Etendeka (Peate & Hawkesworth, 1996; Machado et al., 2018), Ethiopian Traps (Kieffer et al., 2004; S. Krans et al., 2018), and Columbia River Basalts (CRB, Brueseke et al., 2007; Wolff & Ramos, 2013; Moore et al., 2018; Potter et al., 2018). The CAMP section shown in Figure 3d is particularly illustrative since the stratigraphic section is divided into four individual pulses based on paleo-secular variation analysis each of which has been interpreted to have been erupted within a few hundred years (Knight et al., 2004; Font et al., 2011; Marzoli et al., 2019, although see the discussion in Section 3.3). Nevertheless, the individual lava flows in a "pulse" have different major and minor elements, as well as Sr and Nd isotopes with the same quasi-oscillatory oscillations as the DT examples. These variations are characteristic of continuous magma mixing, recharge, assimilation, and fractionation (Section 4) and are hence inconsistent with expectations from a single large homogeneous upper crustal magma reservoir (See Section 2.2.2). We readily acknowledge that precise REAFC characteristics (with respect to values of x,y,z) for each CFB province are different and a careful examination of each CFB, and the corresponding magmatic architecture constraints, is beyond the scope of this study. Nevertheless, the general prevalence of flow-by-flow geochemical variations is a robust feature.

In many CFBs, a significant component of the magmatic system is the shallow ($<$ 4 km, See Magee, Ernst, et al., 2019; Hoggett, 2019; S. M. Jones et al., 2019) sill complexes emplaced in cratonic and passive margin sedimentary basins. The typical sill radius, area, and thickness in these complexes as is illustrated by a dataset from NAMP are 1-7.5 km, 1-100 km$^2$, and 100-500 m respectively (S. M. Jones et al., 2019; Magee et al., 2018; Magee, Ernst, et al., 2019). Nevertheless, the volumes of some of the largest sills can be similar to a single DT eruptive episode e.g., CAMP Palisades Sill (1500- 5000 km$^3$, Husch, 1990), Karoo-Ferrar associated Peneplain Sill (4750 km$^3$, Gunn & Warren, 1962), and Basement Sill in McMurdo Dry Valleys (1050 km$^3$, B. Marsh, 2004; Petford & Mirhadizadeh, 2017), Karoo Basin sills (upto 3000 km$^3$, H. Svensen et al., 2012), and the Karoo-Ferrar Dufek-Forrestal intrusions (10,200-11,880 km$^3$, Ferris et al., 1998). Since most sill complexes have been discovered and characterized primarily through seismic datasets (with some boreholes) (e.g., Magee et al., 2018; Galland et al., 2018; Magee, Ernst, et al., 2019), any geochemical variation between individual sills is difficult to ascertain.

The one exception to this is the Karoo Basin where a large number of sills are exposed and accessible. Through careful sampling of 5 sills and a dike over 125 km$^2$ area (Golden Valley Sill Complex, Karoo basin, South Africa), Galerne et al. (2008, 2011) found that each of the sills, barring two, have different geochemical signatures and hence were fed by a different magma batch. This conclusion is supported by the few studies of larger sills which show evidence for multiple magma injections based on stratigraphic geochemical and petrological analysis e.g., the Beacon Sill (M. Zieg & Marsh, 2012), Palisades Sill (Husch, 1990; Gorring & Naslund, 1995), the Basement Sill (B. D. Marsh, 2004; Bédard et al., 2007), the Doros Complex (T. Owen-Smith & Ashwal, 2015), and the Black Sturgeon Sill (M. J. Zieg & Wallrich, 2018). Thus, these results suggest that large sills and sill complexes were emplaced as multiple magma batches over time (potentially within 10-100s of years to prevent sill solidification) with some of the magma injections having different compositions. We contend that it is difficult to obtain these eruption characteristics if the magma is sourced from a single, large well-mixed magma reservoir, especially considering the eruption rate results described later in this study.

# 5 Magmatic Architecture observations - Deccan Traps

The final, direct constraint on the crustal magmatic architecture of continental flood basalts is the exposed magmatic plumbing system - the dike swarms & sills. In CFB provinces, the primary mechanism for magma transport from crustal reservoirs to the surface is a combination of mafic dike swarms, sills - both deep crustal sills (Wrona et al., 2019; Buntin et al., 2019) as well as shallow sill complexes (Maccaferri et al., 2011; Magee, Ernst, et al., 2019). Since the Deccan Traps were emplaced primarily atop Archean cratons, there are no large scale sill complexes hosted within sedimentary basins like many other CFBs e.g., Karoo-Ferrar province, Siberian Traps, North Atlantic Magmatic Province, and Central Atlantic Magmatic Province (See H. Svensen et al., 2018; Magee, Ernst, et al., 2019, and references therein). The sills in Deccan are generally restricted to the outlying regions of the province with sedimentary basins e.g., Jurassic sedimentary rocks in Kutchh (Biswas, 1982, 1987; Karkare & Srivastava, 1990; Duraiswami et al., 2008; Karmalkar et al., 2016) and Gondwana sediments in Central Indian Satpura Range (Crookshank, 1936; Sengupta & Ray, 2011). In addition, a small lopolith is emplaced in an intertrappean at Amboli hill, Mumbai (Tolia & Sethna, 1990), after the main DT eruptive phase (See discussion in Section 3.2.2 about Mumbai section).

The saucer-shaped Mahad sill (Duraiswami & Shaikh, 2013) in the southwestern Deccan Traps is one of the few DT sills within the main DT section along with the olivine-gabbro Khopoli intrusion (Cucciniello et al., 2014). The Mahad sill, the larger of the two, is 7.1 km x 5.3 km long and is intruded into a Bushe Formation flow. Since the sill is geochemically associated with either the Poladpur- or Ambenali Formation, the sill was emplaced at $< 1$ km depth from the surface (Duraiswami & Shaikh, 2013). Based on existing work, none of these sills feed subaerial lava flows. The only known basaltic sill in DT that may have been responsible for magma transport to an overlying lava flow is the $\sim 200$ m Chakhla-Delakhari Intrusive Complex (CDIC) in the Central Deccan region (Crookshank, 1936; Shrivastava et al., 2008; H. C. Sheth et al., 2009). CDIC is the largest DT basaltic sill complex with an area of about 150 $km^2$ and is emplaced primarily in the Gondwana sandstone. The complex stratigraphic shape, spatially variable thickness, the mineralogical as well as chemical heterogeneity, and textural features of the CDIC strongly suggest that it is composed of multiple individual intrusions with different compositions (Crookshank, 1936; H. C. Sheth et al., 2009). Finally, We note that there are a number of small primarily alkaline intrusive complexes in the Saurashtra and Malwa Plateau subprovinces of DT such as Girnar, Osham, Barda (dominantly granophyre), Alech hills (dominantly rhyolite) of Saurashtra as well as the Pavagadh and the Phenai-mata igneous complexes in the Narmada-Tapti Tectonic Zone (Auden, 1949; Greenough et al., 1998; B. Singh et al., 2014a; Cucciniello et al., 2019). Although a number of these intrusions (e.g., Phenai-mata and Pavagadh) are associated with local tholeiitic and alkaline magmatism, there is as yet no geochemical, geochronological, or volcanological evidence suggesting that these intrusive complexes fed large DT lava flow units, especially any of the main phase WG flows.

## 5.1 Deccan Dike Swarm

The large Deccan Traps eruptive episodes are instead hypothesized to have been fed by tholeiitic dike swarms (Vanderkluysen et al., 2011). There is no concensus if individual eruptive episodes were fed by a single dike (with sequentially active fissure segments like 1783 Laki eruption, Thordarson & Self, 1993), a set of dikes fed from the same magmatic system, or eruptive centers 100s of km apart that all fed the same flow field (Self et al., 1998; Vanderkluysen et al., 2011; Óskarsson & Riishuus, 2014; Kale et al., 2020). Without a detailed field analysis of individual flow fields along with complementary geochemical and isotopic analyses, each of these models may be applicable for Deccan eruptive episodes with potentially different styles for different DT regions, subprovinces, and/or geochemical formations. The four main DT dike swarms are the Narmada-Tapi dike swarm extending across from Mandla Lobe region to Saurashtra (Sant & Karanth, 1990; Bhattacharji et al., 1996; Melluso et al., 1999; R. Ray et al., 2007; H. C. Sheth et al., 2013; Cucciniello et al., 2015; H. Sheth et al., 2019), the Saurashtra dike swarm (Cucciniello et al., 2020; Chatterjee & Bhattacharji, 2001), the Western Ghats Coastal dike swarm (Widdowson et al., 2000; P. Hooper et al., 2010; H. C. Sheth et al., 2014; Patel et al., 2020), and the Central Dike Swarm (also called Nasik-Pune dike swarm) located to the east of the WGE (N. Bondre et al., 2006; Vanderkluysen et al., 2011; R. Ray et al., 2007; Das & Mallik, 2020) (See Figure 6). Based on similar major and minor element geochemistry as well as Pb-Nd-Sr isotopes, the three (excluding the Saurashtra swarm which feeds that corresponding sub-province) major dike swarms have been correlated with individual geochemical formations across the Deccan Traps (e.g., Vanderkluysen et al., 2011; Patel et al., 2020, and references therein). Vanderkluysen et al. (2011) concluded that lower and middle subgroups were most likely fed by the oriented Narmada-Tapi and Coastal dike swarms while the Upper subgroup lavas were predominantly fed by Nasik-Pune (Central) dike swarm.

Since the spatial pattern of dike emplacement in the upper crust is primarily controlled by the crustal stress field (See M. A. Richards et al., 2015, and references therein), we can use the dike distribution in the Deccan Traps to test the large magma reservoir model. In particular, we are interested in assessing whether there is any clear radial or circumferential dike pattern as expected from a single magma body (R. Ernst et al., 2019; Buchan & Ernst, 2019; Bistacchi et al., 2012; Johnson, 1961). The other possible alternatives are that the dike orientation was controlled by a) overall tectonic extension pattern (especially in the Narmada-Tapi rift zone and the West Coast) in India (O. P. Pandey, 2020), or b) stress field associated with multiple individual magma reservoirs leading to a heterogeneous, and likely time-dependent, stress field. We test these end-members using a new data dataset from a combination of Deccan field mapping by the Geological Survey of India (GSI 1:50k maps, Raju (2016); GSI District Resource Map (2001); GSI Bhukosh (2020 (accessed December 1, 2020))) and joint analysis of satellite image analysis & digital elevation maps (National Geological Lineament Mapping, Jagannathan and members (2010)). Although there is some overlap between the dataset, the majority of the combined dike dataset ($\sim$ 29,000 segments) is from the GSI field mapping ($\sim$ 75 %) and the coverage of the datasets is fairly complementary. Thus, for simplicity, we use the combined dataset for analysis and leave a more careful analysis to remove some of the dike duplicates ¡ 1000 segments based on a preliminary analysis) to future analysis. Since our focus in this study is broad spatial patterns, we do not expect this simplification to introduce an appreciable bias.

In Figures 6 and 7, we show the full dike dataset with the dike segments colored by dike orientations and segment lengths respectively. The highest dike line density (dike number density per $km^2$ integrated over segment lengths) is in Western Narmada-Tapi rift zone with other local maxima in Saurashtra and Coastal dike swarms. The longest single mapped dike segment is $\sim$ 70 km. We anticipate that there are longer Deccan single dikes in reality but they have been mapped as distinct dike segments due to a combination of erosional breaks and lack of access. Since it is difficult to decide which dike segments are part of the same dike without further analysis, especially geochemistry, we have not merged dike segments here. Consequently, we would advise

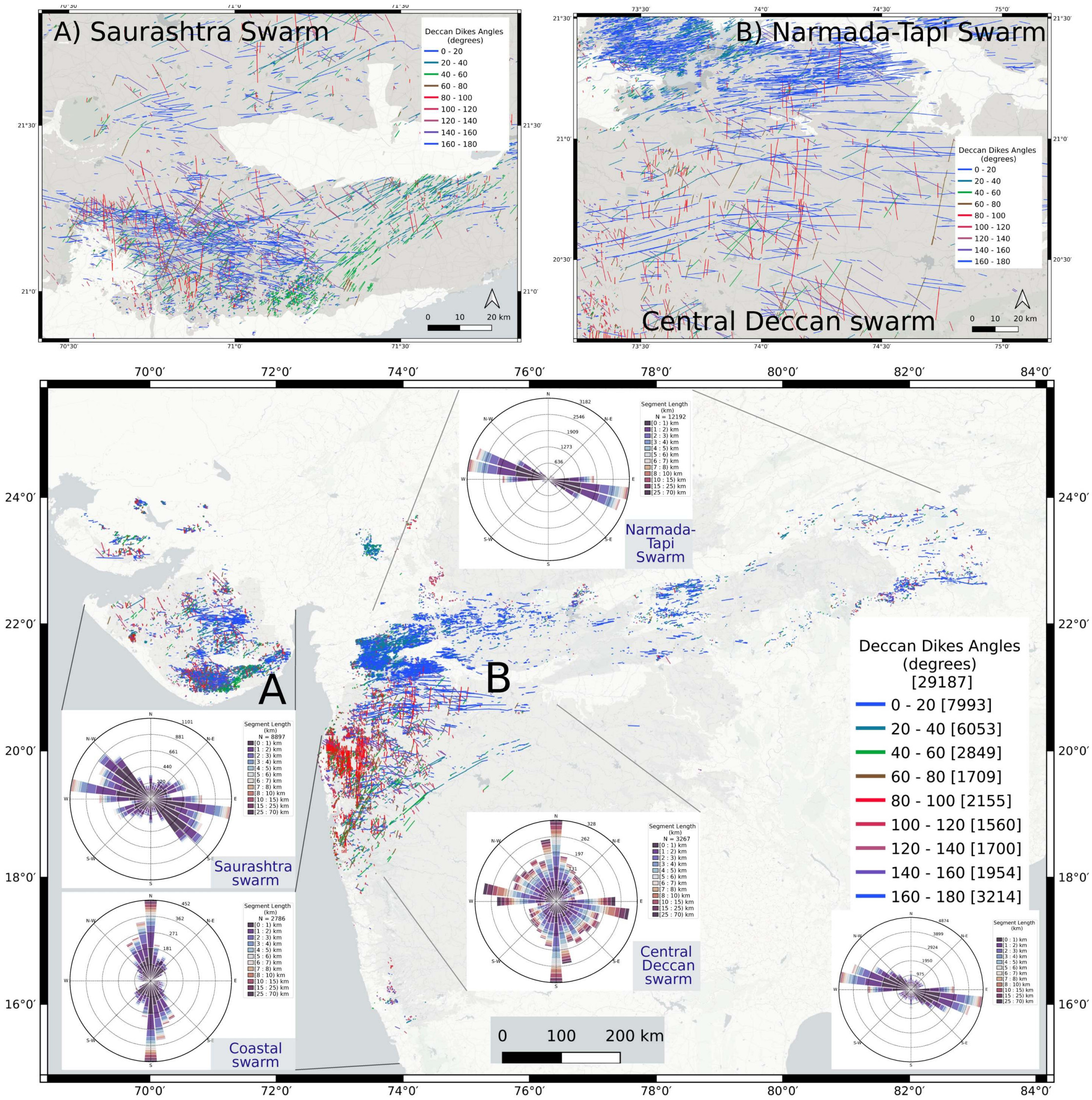


Figure 6: Spatial distribution of ∼29,000 Deccan-associated dike segments (using GSI 1:50 k geologic maps and NGLM dike dataset) with colors denoting dike orientation. Insets A and B show regional zoom-in for Saurashtra and Narmada-Tapi swarms each with multiple overlapping dike orientations. Figure also shows rose diagram for each dike subregion as well as for the whole dataset (in bottom right). For each dike orientation bin in rose diagram, color bars illustrate corresponding dike length distribution histogram.

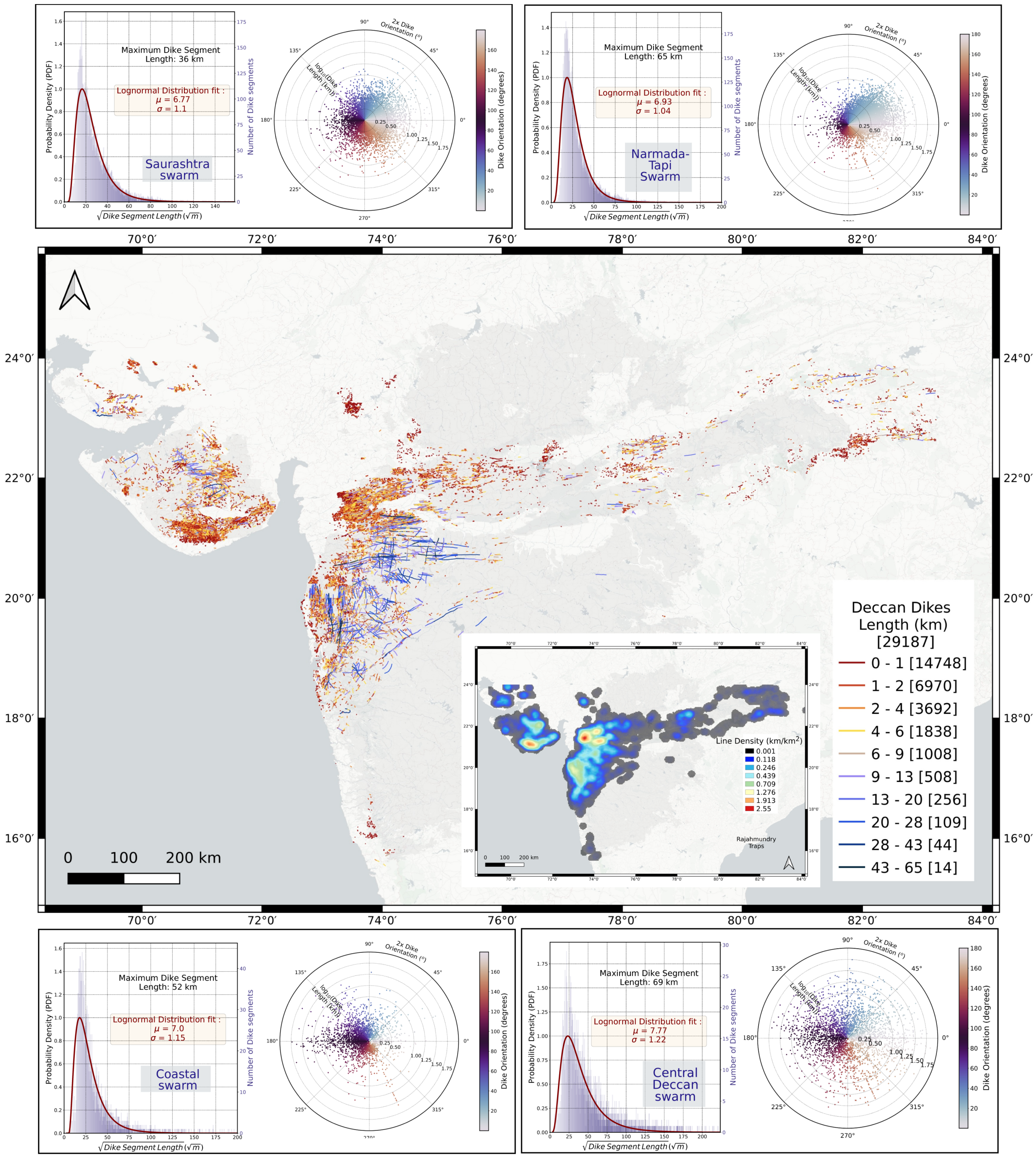


Figure 7: Spatial distribution of ∼29,000 Deccan-associated dike segments (using GSI 1:50 k geologic maps and NGLM dike dataset) with colors denoting dike segment length. Inset in main figure shows spatial map of overall dike density (weighted by dike segment length). For each sub-region, corresponding figure shows PDF of dike length (and best-fit lognormal distribution) as well as polar plot of (2x) dike orientation and (log10) dike segment length.

some caution in interpreting the dike segment length distribution since it is naturally biased against long dikes. Looking at the full dataset, we find that the Deccan dikes overall have a strong ENE-WSW orientation (Figure 6, lower right inset) consistent with previous work (Auden, 1949; Vanderkluysen et al., 2011). However, regionally, the four dike swarms have distinct characteristics with the Narmada-Tapi swarm (Median length – 954 m) generally striking ENE-WSW parallel to the rift-and-graben structure following the Narmada and Tapi rivers (Figure 6 , see top panel B). These dike segment orientations clearly highlight the significance of pre-existing crustal structure for the Deccan crustal magma transport. Next, the Coastal Dike Swarm (from the coast to the WG escarpment, Median length – 990 m) has a strong N-S approximately parallel to the west coast and the India-Seychelles rifting pattern. The Saurashtra dike swarm (Median length - 890 m) is broadly similar to Narmada-Tapi swarm in terms of orientations though there are much more orientation spread with intersecting dikes (Figure 6 , see top panel A). Finally, the Central Deccan swarm, associated with the eruption of the Wai subgroup flows (Median length – 2330m), has weaker preferred directions compared to the other swarms. In addition, the dike segment length is substantially larger than other swarms potentially indicative of a different magmatic overpressure regime and more localization into single dikes (as is evident with the low dike density, Figure 7 main figure inset). Like other CFB dike swarms (e.g. Columbia River Basalts, the log-normal density distribution provides the bestfit to the dike segment length data akin to other LIPs (e.g., CRB Chief Joseph Dike Swarm Morriss et al. (2020)), the log-normal distribution provides the best description of the dike segment length distributions (Figure 7). We expect that systematic under-prediction of the dike length distribution at the small end-member is an artifact of a single dike sub-divided into multiple parallel segments. A full analysis of the dike dataset, and its biases, is beyond the scope of the current study.

Each of the dike swarms is typically composed of a number of sub-swarms of 100s of dike segments each (Vanderkluysen et al., 2011; H. C. Sheth & Cañón-Tapia, 2015) with a wide range of thickness from 1 to 62 m (median of $\sim$ 10-20 m for different subswarms) and lengths (1 km to 79 km R. Ray et al., 2007; N. Bondre et al., 2006; H. Sheth et al., 2019). The only exception is the Goa dike swarm with typically shorter ($<$ 200 m) and thinner (average $\sim$ 6 m) dikes (Gadgil et al., 2019). A significant fraction of these dikes show evidence of multiple magma injections in the form of multiple columnar-jointed rows (2-5 injections) (See discussion in H. C. Sheth & Cañón-Tapia, 2015). Due to surface erosion, the dike tops in DT are generally truncated and thus a clear feeder relationship between lavas and flows is not established. Consequently, it is unclear if the dikes terminated in the crust/lava flow pile or if they fed lava flows directly.

We can compare the expectations of the dike spatial pattern with those from a large magma reservoir by computing the intersection points between dike segments. In order to reduce any biases due to missing/unmapped dike segments, we extended each dike segment by 50 km on either side before computing the intersection points. The results of this analysis (with a total $\sim$ 32,000,000 intersection points) are shown in Figure 8 with a few zoom-in insets for the Coastal dike swarm, Narmada-Tapi dike swarm, and Saurashtra swarm. The key result of note here is that there is no preferred location of dike intersection as would be expected for the case of a central magma reservoir (and a clear circumferential or radial dike swarm). Instead, the dike intersections form multiple local maxima within each sub-swarm as clearly seen for the Coastal dike swarm (Figure 8).

In conclusion, we find that the spatial pattern of Deccan dikes is inconsistent with the predictions of the large magma reservoir model. Instead, the dike distribution is more consistent with a combination of strong tectonic control and variations in small-scale crustal stress field due to small magma bodies. This conclusion is particularly true

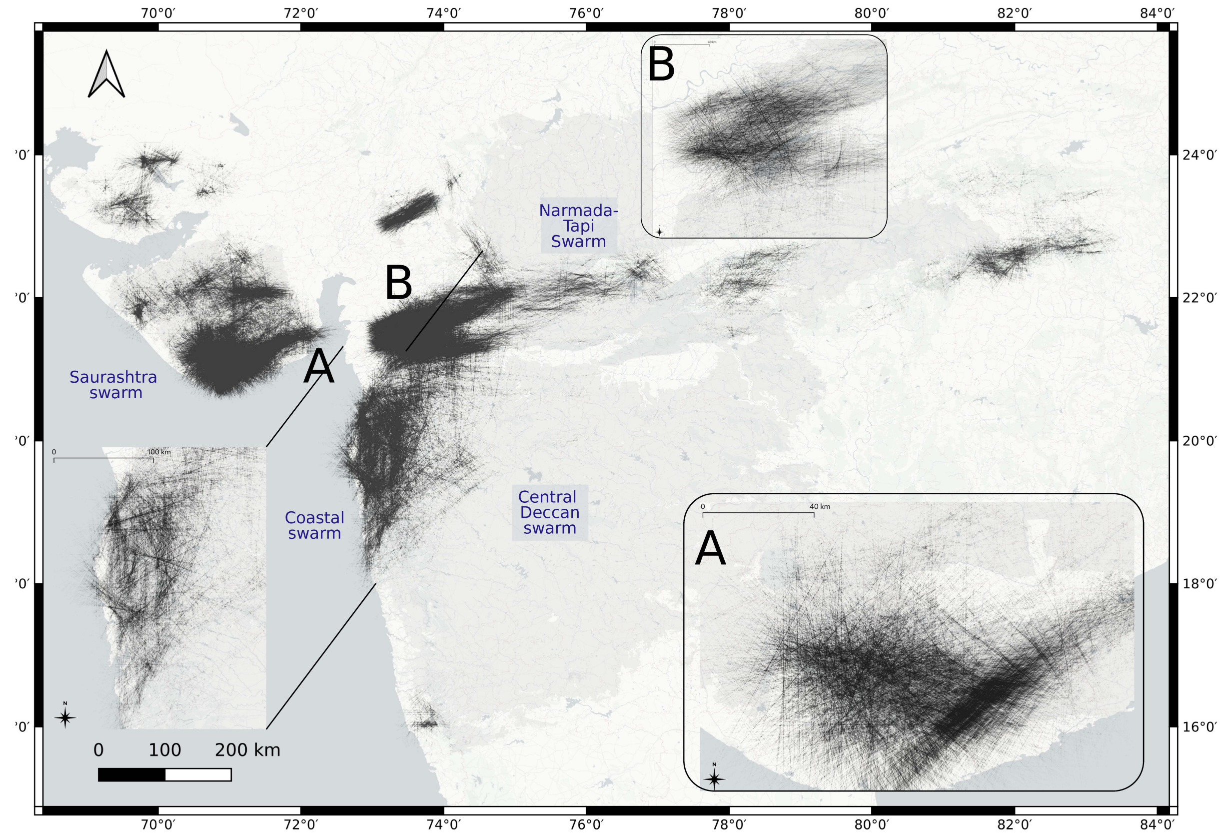


Figure 8: Spatial distribution of intersection points between the ∼29,000 Deccan-associated dike segment with each segment extended by 50 km on either site. The insets show zoom-in regions for the Coastal, Narmada-Tapi, and Saurashtra dike swarms.

for the Central Deccan swarm which shows much larger orientation spread vis-a-vis other dike swarms and geochemically, represents the most likely feeder for voluminous Upper subgroup Deccan lavas (Vanderkluysen et al., 2011). We would note that our conclusion from the dike dataset analysis is completely consistent with the analysis in M. A. Richards et al. (2015) based on a more restricted dataset. Analogous to their interpretation, we posit that the Deccan crustal stress field transitions over time from regional extensional stress dominated (e.g., Narmada-Tapi Swarm) to un-oriented crustal magma intrusion dominated (due to regional-scale lower crustal intrusion; e.g., Central Dike Swarm).

Our conclusion of a lack of large magma reservoir is further supported by the large number of dike segments mapped in Deccan. In typical shield volcanoes, only 30 % of the dikes are feeders (Galindo & Gudmundsson, 2012). Even if the number is higher for DTs (e.g., 50-50 ratio, H. C. Sheth & Cañón-Tapia, 2015), a large number of observed dike segments (e.g., Misra et al., 2014; Misra & Mukherjee, 2017) suggests magma reservoir failed even more frequently to form dykes (Kavanagh, 2018) than the rate of eruptive episodes. This observation makes it even more difficult for a single large magma reservoir to feed the surface eruptions given the various constraints on eruptive tempo described above (See Section 2.2.3).

## 6 Deccan Traps Intrusive structure - Geophysical Observations

Given the lack of direct observations of intrusive structures in Deccan Traps, we compiled results from various studies using a variety of geophysical methods (gravity, magnetics, magnetotelluric, and Deep seismic sounding) to constrain the crustal intrusive structure. Since the 1980s, geophysical methods have been used to image continental crustal structures, especially in Saurashtra and the Narmada-Tapi Rift zone, along with a few studies across the central Deccan Plateau and the WGE. If mafic magmatic bodies are emplaced at deep depths, they can be converted to a basic garnet-pyroxene-plagioclase-bearing granulite facies assemblage, which are good seismic reflectors (e.g., K. G. Cox, 1980). Similarly, the higher density of intrusive mafic bodies, especially mafic and ultra-mafic cumulates, would naturally lead to a signal in the gravity field as well as the change in seismic velocity (Ridley & Richards, 2010; M. Richards et al., 2013). We note that the high density of the mafic and ultramafic intrusions can lead to post-emplacement deformation and downward crustal flow and potentially crustal delamination if the surrounding crust has sufficiently low viscosity (Roman & Jaupart, 2016, 2017; Gorczyk & Vogt, 2018). However, the presence of the upper crustal Bushveld complex (Eales & Cawthorn, 1996) and the lower crustal Seiland Igneous Province (R. B. Larsen et al., 2018) suggests that this process does not completely remove the crustal intrusives. Thus, the present-day geophysically imaged intrusive bodies represent a lower bound on the total DT intrusives, especially at the Moho depth.

Overall, the shear wave velocity structure of the Indian crust (e.g., Maurya et al., 2016; Sharma et al., 2018; Saha et al., 2020, and references therein) suggest a thick crust (with an underplated mafic layer) in parts of the Indian subcontinent influenced by the Deccan Traps. This signature is particularly strong for parts of the Kutch-Saurashtra region (with the various intrusive complexes), Cambay rift, Narmada-Tapi rift, and the Western Ghat/Western Coastal region and much less for the Eastern/Central Deccan. There is an analogous signature of Deccan plume melting in the lithosphere thickness for India with a much thinner (100 - 120 km) lithosphere under the same regions vis-a-vis > 200 km under Central-Eastern Deccan (Maurya et al., 2016; Saha et al., 2020). Interestingly, this pattern in lithosphere thickness variations and crustal underplating is similar to the spatial (though not necessarily the number density) distribution of Deccan Dikes (Figure 6). This observation suggests that the Deccan crustal magmatic system follows the regional extent of Indian lithosphere-Réunion plume interaction. However, there is no evidence in these analysis of a large single/few upper or mid crustal magma reservoirs.

To test this conclusion further, we use only high resolution crustal datasets, preferentially local to regional scale in our compilation. This reduces any biases due to spatially variable data coverage in Indian continent scale seismic velocity studies, which typically have 50 km (or larger) spatial resolution. For our analysis, we have mostly followed the interpretation of the original studies to delineate the extent of intrusive mafic bodies vis-a-vis continental crust and sediments. Ideally, a re-analysis of all the datasets using the same model and set of assumptions would lead to more robust results. However, in most cases, the raw datasets are not publicly available for us to be able to do the analysis. In order to reduce biases, we have filtered our compilation only to include studies that utilize more than a single geophysical dataset, and preferably other geological observations, in the model inversion in order to reduce the non-uniqueness of the solutions. Another challenge with the available datasets is that they typically are for 2D sections except for some MT studies. Consequently, our results are not very informative about the full 3D geometry of intrusive bodies. Finally, we expect our calculated volumes to typically be a lower limit for the true intrusive volumes since geophysical methods may not detect small crustal intrusive bodies below the typical model resolution of 1-5 km (e.g., Patro & Sarma, 2016a). Although we have a complementary reduction in the volume of the surface lava flows due to erosion, we expect this would to be a smaller bias than the undetected small intrusive bodies.

The other, potentially much larger error in our compilation, comes from the interpretation of the inversion results with respect to the presence of an intrusive body vis-a-vis crustal faults, fluids, and background crustal structure. We refer the reader to individual studies in our compilation for the inversion method and interpretation of the results. In the end, we use results from 19 individual studies for a total of 53 individual measurements (Table 1). In Figure 9, we show the distribution of estimated Intrusive/Extrusive ratio for datasets with only large scale Moho-depth underplating (median of 5.5, 25th to 75th percentile of 2.5-8.5, 18 measurements), datasets include only upper crustal magma bodies (median of 1.3, 25th to 75th percentile of 0.5-2.0, 17 measurements), and datasets which include magmatic bodies throughout the crustal column (median of 8.5, 25th to 75th percentile of 5-14.0, 17 measurements). The primary result of this compilation is the absence of significant upper crustal magma bodies in the Deccan Traps with most of the intrusive volume in the lower crustal region. In the following, we describe the results of this analysis for Deccan Traps as well as some supporting datasets from other CFBs.

### 6.1 Mafic Underplating

In almost all the DT sections with geophysical datasets, we find significant (multiple km thick) layers of underplating in the lower crust. Typically, the Extrusive to Intrusive ratio is an average value of 6 and a large range between $\sim$ 0.5 to 15.5. Even in datasets with some upper crustal magma bodies (full crustal column, Figure 9), most of the mafic intrusives are still at Moho depth (except the Phenaimata and Pavagardh intrusion described in the next section). A large value for Moho depth intrusions is consistent with the models of K. G. Cox (1980) as well as Black and Manga (2017) who proposed that at least 30-40 % of the parental melt is emplaced at the base of the crust as mafic cumulates. Existing gravity and seismic data suggest that similarly, thick Moho/lower-crust mafic underplates are common for both oceanic and continental flood basalts (Furlong & Fountain, 1986; K. G. Cox, 1993; R. White & McKenzie, 1989; Coffin & Eldholm, 1994; Ridley & Richards, 2010; M. Richards et al., 2013; Mammo, 2013; Thybo & Artemieva, 2013; Deng et al., 2016). Thus, there is significant observational support for models of flood basalt volcanism (See Section 2.1

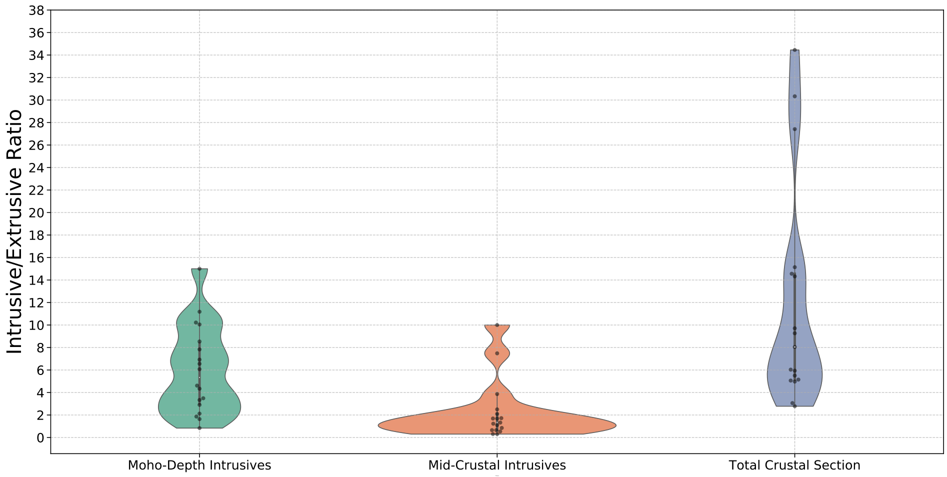


Figure 9: Kernel density estimates of Intrusive to Extrusive (I/E) Ratio for Deccan Traps using a literature compilation of gravity, seismic, and MT observations spanning the Western Ghats, Saurashtra, and Narmada Rift Valley in India (Table 1). The observations are divided into three groups based on whether the compiled datasets observed only Moho-depth intrusives, only mid/upper crustal intrusives, or mafic/ultramafic bodies across the whole crustal section (Moho plus crust). For each category, the individual measurements are plotted as circles while the bar plot shows the 25th to 75th percentile range. The high values for the I/E ratio (>20) are the Phenai-mata and Pavagadh intrusive complexes. The range of I/E ratio for the whole crustal sections is approximately equivalent to the sum of distributions of the Moho-depth underplating and mid/upper crustal intrusives.

for details) with deep crustal ponding and fractionation of ultramafic primary melts (Farnetani et al., 1996).

We note that the common presence of a thick ($\sim$ 5-15 km) underplated layer does not necessarily equate to a single magma body of equivalent size. Instead, frequent mafic intrusions(sills) at lower crustal depths along with visco-elastic deformation of the cumulate bodies can also lead to the formation of a large seismic underplate layer (J. D. White et al., 2009; Roman & Jaupart, 2016, 2017; Gorczyk & Vogt, 2018; Galland et al., 2018). We have observational support for this hypothesis from an exhumed lower crustal LIP section - the Seiland Igneous Province (SIP, R. B. Larsen et al., 2018, and references therein). The $\sim$ 5000 km$^2$ SIP is a dominantly mafic and ultramafic intrusive complex that was emplaced at the 25-35 km depths and represents the deep magmatic plumbing system of the Ediacaran age Central Iapetus Magmatic Province (e.g., Higgins & Breemen, 1998). With a total preserved volume of 17000 km$^3$ (Pastore et al., 2018), the SIP is made up of a number of multiple layered mafic plutons (85 % area, each $\sim$ 10-50 km$^2$), deep-rooted four ultramafic complexes (8-10 % area, each $\sim$ 20-100 km$^2$ and roots up to 9 km), as well as 2-5% carbonatitic rocks and alkaline and dioritic plutons. Based on field relationships, especially various chill margins between the mafic-ultramafic units and mafic-country rock metasediments (See R. B. Larsen et al., 2018, and references therein), SIP was clearly emplaced sequentially. The first mafic (and alkalic) magmas were intruded into lower crustal metasediments. Subsequent parental melts were emplaced into mafic gabbros, which were still close to solidus temperature. This is clearly evidenced in the field by extensive wall rock melting and assimilation between the gabbros and ultramafic intrusions (Griffin et al., 2013). The thermo-chemical insulation of emplacement into hot mafic rocks likely enabled the ultramafic melts to retain their parental composition with the deep ultra-mafic complexes acting as de facto LIP conduits for the shallow system (Grant et al., 2016; Degli Alessandrini et al., 2017).

In conclusion, observations clearly illustrate that the SIP magmatic system was never molten in total at the same time as a large lower crustal magma chamber but was rather assembled from multiple bodies. Additional support for this interpretation is provided by the seismic detection of a large (97 x 62 km) but thin (180 $\pm$ 40 m) igneous sill in the North Sea lower crust (17.5-22 km depths, Wrona et al., 2019) associated with the NAMP. As illustrated by a large number of intrusive steps and seismic reflection amplitude anomalies, the sill was formed sequentially with a complex pattern of lateral flow within the sill (Wrona et al. (2019), See Magee, Ernst, et al. (2019) for aerial examples of these features).

### 6.2 Upper Crustal Intrusive Bodies

With regards to upper crustal magma bodies, we do not find any evidence in the compiled sections for a large, continuous, upper crustal magma reservoir as hypothesized (See Section 2). Instead, the individual magma bodies, when detected, are typically small (5-15 km across and a few km thick, e.g., Patro & Sarma, 2016a; Prasad et al., 2018), are prismatic in shape, and are distributed in complex patterns depending on the pre-existing crustal structure. Since smaller magma bodies of this size are close to the resolution limit for many studies, our estimated Intrusive/Extrusive ratio may be biased to smaller values. We posit that, in reality, the upper crust may have more small magma bodies as petrologically required based on ubiquitous shallow fractionation (See Section 4 for details). A common feature across many studies, especially in the Narmada-Tapi rift zone region, is that the intrusive bodies are typically associated with pre-existing Precambrian age crustal and lithospheric fault zones (and other weak zones) (Kale et al., 2017, 2020). This suggests that Deccan Traps parental magma and magmatic volatiles utilized these pathways for melt transport to and through the crust (e.g., Bhattacharji et al., 1996; Naganjaneyulu & Santosh, 2010; Azeez et al., 2013;

Patro & Sarma, 2016a; Azeez et al., 2017a; G. P. Kumar et al., 2017; Patro et al., 2018). This is further reflected in the vertical magmatic features associated with fault zones (Chowdari et al., 2017). These magma bodies may have acted as conduits for upward transportation of magmatic material to other intrusive bodies (Patro & Sarma, 2016a; Naganjaneyulu & Santosh, 2011).

A number of other studies in LIPs and other magmatic systems have illustrated that pre-existing zones of lithospheric and crustal weakness play a critical role in magma transport (e.g., Begg et al., 2018; Peace et al., 2018; Alghamdi et al., 2018; Latyshev et al., 2019; Magee, Muirhead, et al., 2019). We note that the conclusions of our compilation with respect to the size of individual upper crustal magma bodies differs strongly from the results in Bhattacharji et al. (2004) (based on 2d and 3D gravity analysis). They inferred the presence of an extremely large (12 km thick, 300 x 30 km) magma upper crustal (6 km depth) body. However, G. S. Rao et al. (2018) used gravity datasets in combination with constraints from seismic observations, as well as magnetic datasets, and found that the observation can be better explained with much smaller crustal magma bodies along with a Moho depth underplate. Analogously, Patro and Sarma (2016a); Prasad et al. (2018) find that using higher-resolution datasets, the magma bodies in Narmada-Tapi Rift Zone are smaller and connected (if at all) thorough narrow zones instead of a larger, thicker, single magma body proposed originally by Bhattacharji et al. (2004).

Within our data compilation, the Phenai-mata and Pavagadh intrusive complexes in the Narmada-Tapti Rift Zone region are clear outliers (with very high intrusive/extrusive ratio, Figure 9). The Phenai-Mata complex consists of dominantly basaltic flows (2/3 by volume) along with alkaline plutonic rocks (1/3 by volume) and some orthopyroxene layered gabbro (Hari et al., 2011; Hari, Prasanth, et al., 2018). B. Singh et al. (2014a) used a combination of gravity and magnetic datasets to infer the presence of a large (11 km x 52 km) mafic body extending all the way from the surface to lower crustal depths ($\sim$ 20 km). The strong orientation of the intrusive body along the rift zone axis indicates the importance of a pre-existing tectonic structure for facilitating magma transport to the surface. The Pavagadh intrusive complex comprises twelve flows with a wide compositional range from olivine basalts and andesites to rhyolites (H. C. Sheth & Melluso, 2008; Hari et al., 2011). With similar gravity-magnetic analysis, B. Singh et al. (2014a) found a low-density rhyolitic body (3x 5 km) extending to about 10 km from the surface. This rhyolitic plug is, in turn, emplaced within a larger mafic intrusive body (20 km x 13 km x 10 km depth). Finally, both the rhyolite and the mafic rocks are underlain by an ultra-mafic body (20 km x 16 km) up to a depth of 20 km (and potentially larger). Thus, both the Phenai-Mata and Pavagadh intrusive complexes exhibit a trans-crustal DT magmatic system. Based on the different magnetization direction of the ultra-mafic and the rhyolitic magma bodies, B. Singh et al. (2014b) proposed that the Pavagadh complex may have been active between Chron 30N to Chron 29N/28N and was hence a long-lived system. Since most of the large WG flows are inferred to have been fed by dike swarms, we do not think that these igneous complexes are representative of the majority of DT magmatic architecture.

### 6.3 Relationship with Layered Mafic Intrusions

Another interesting conclusion of our data compilation is the lack of any intrusive bodies akin to large layered mafic intrusions such as the Rustenburg Layered Suite (RLS) of the Bushveld Igneous Complex. The RLS is the world's largest layered mafic intrusion that was emplaced about 2.06 Billion years ago. With an area of $\sim$ 65,000 km$^2$, a thickness of 7-9 km (Eales & Cawthorn, 1996; Cawthorn, 2015), and a shallow emplacement depth (0.15-0.25 GPa: Pitra & De Waal, 2001), R. E. Ernst et al. (2019) (and references therein) have suggested that RLS may be archetype for a typical CFB upper crustal magma reservoir. It has been typically assumed that the RLS represents a single, long-lived magma reservoir of the same size as its present-day extent with rapid assembly through multiple magma injections followed by a long period of closed system fractionation (Wager & Brown, 1968; Cawthorn & Walraven, 1998; Kruger, 2005). Such an interpretation serves as one of the original motivations for the large magma reservoir models discussed in Section 2. However, Robb and Mungall (2020) used a combination of high-precision U-Pb dates, plagioclase zoning observations, and thermal models to illustrate that RLS was instead accreted as an out-of-sequence stack of sills corresponding to individual magma intrusions over a 1.2 Ma time-period (C. Lee & Butcher, 1990; A. A. Mitchell & Scoon, 2007; Mungall et al., 2016; Scoon et al., 2020; Scoon & Costin, 2018; Hayes et al., 2018; Scoon & Costin, 2018; A. Mitchell et al., 2019). Thus, instead of a single magma reservoir, RLS more likely represents a region of extensive sill emplacement and subsequent thermo-chemical insulation of individual magma bodies akin to Seiland Igneous Province (as described in the previous sections). Thus, the presence of a large Bushveld sized intrusive body does not necessarily correspond to a large, well-mixed crustal magma reservoir. In fact, based on field observations, RLS was emplaced after the surface eruptions of the basaltic-rhyolitic Rooiberg Group (Lenhardt & Eriksson, 2012). Thus, the RLS was not an upper crustal magma body feeding the surface lavas, but was instead accreted as a set of sills under a lithostatic load.

For the CFBs in the past 200 Ma, we are not aware of any large upper crustal magnetic or gravity anomalies of the spatial scale and amplitude expected for a Bushveld sized large magma reservoir (or crustal deformation features suggesting their presence and subsequent delamination) (e.g., Mammo, 2013; Sharma et al., 2018). For instance, the Marzen et al. (2020) and Gao et al. (2020) found no seismic evidence for a large magma body in the upper crust in the Triassic South Georgia Rift associated with the CAMP volcanism in North America. Instead, both of these studies only found a thick ($> 5$ km) magmatic underplating layer at Moho depths. Although there are layered mafic intrusions associated with recent CFBs such as the Doros Complex ($\sim 20$ km$^3$, Parana–Etendeka CFB, T. Owen-Smith & Ashwal, 2015), NAMP associated central complexes such as Rum, Mull, Skye ($< 100$ km$^3$, O'Driscoll et al., 2006; O'driscoll, 2007; Namur et al., 2010), the Skaergaard intrusion ($\sim 280$ km$^3$, NAMP, Nielsen, 2004), La Balma-Monte Capio intrusion (CAMP, Denyszyn et al., 2018), and the Graveyard Point Intrusion (Snake River Plain Basalt, C. M. White, 2007), they are much much smaller than RLS. In conclusion, we posit that a Bushveld layered mafic intrusion does not represent a typical upper crustal magma reservoir for Deccan Traps definitely, and potentially for many other CFBs in the last 200 Ma.

### 6.4 Implications for Magmatic Architecture

The main conclusion from the Deccan geophysical compilation of the Intrusive/Extrusive (I/E) ratio is that there is a significant volume of intrusives: I/E = 5-15 with a few outliers. Additionally, the majority of the intrusive magmatic volume is in the lower crust with no clear evidence for a large spatially homogeneous upper crustal magma body. To first order, the lack of an upper crustal large magma reservoir is inconsistent with the buoyancy-driven magmatic eruption model (See section 2.2.2, 2.2.3), whereas the large underplated layer is consistent. Nevertheless, the observations from Seiland Igneous Province, as well as observations from various Layered Mafic Intrusions, show that the presence of a large magmatic intrusive body does not imply that it was necessarily molten at the same time as a well-mixed magma reservoir.

Another added complication with the geophysical observations is that we see only the end-result of the system. Since there are no geochronological constraints for most of the intrusive bodies, it is not clear if they were emplaced prior to or post the main phase of volcanism. A particularly illustrative example is the McMurdo Dry Valleys sill complex (B. D. Marsh, 2004) with four large sills (from top to bottom): the Mt. Fleming Sill, the Asgard Sill, the Peneplain Sill, and the Basement Sill. The sill system is capped by the Kirkpatrick flood basalts on top. Although the bottom-up sequence of intrusions for the system makes logical sense, at least as a first guess, B. D. Marsh (2004) used a variety of field, petrological, and textural observations to demonstrate that the system was instead formed top-down. The sills were emplaced after the eruption of the flood basalt lava flows instead of feeding the lava flows through sill-edge dikes (B. D. Marsh (2004); Jerram et al. (2010), see D. H. Elliot and Fleming (2018) for a detailed comparison with alternative scenarios). With an increasing lithostatic weight of the flood basalt sequence, the magma was emplaced in a progressively stepped down sills first through the Karoo sedimentary sequence (Mt. Fleming, Asgard, Peneplain Sill), and finally in the granitic basement (Basement Sill). This relationship is not unique for the McMurdo Dry Valleys and has been in multiple CFBs (e.g., Hansen et al., 2011; S. Burgess et al., 2017; Jerram et al., 2018) and has also been proposed for some CFB associate layered mafic intrusions (e.g., Higgins, 2005, Sept Iles intrusive Suite). Thus the intrusive structures and Intrusive/Extrusive Ratio calculated above should be considered as an upper limit (neglecting the methodological uncertainties) integrating over the whole CFB sequence.

## 7 Conclusions

In summary, the primary observations that strongly argue against the single, large magma reservoir model are as follows :

- Deccan Geochemistry (Section 3.1, 4) : In the Western Ghats sections for Deccan Traps, the crustal and lithospheric assimilates can vary rapidly between geochemical formations. Even within individual flows of a geochemical formation, there are quasi-oscillatory geochemical and petrological variations. Thus, the magmatic system has a relatively short geochemical memory. This is difficult to explain within a magmatic plumbing system is composed of only a few large magma chambers which integrate magma compositions over multiple 100s of kyr (Black & Manga, 2017). Additionally, the model result for a REAFC/RTF type shows that the intra-flow variations require a continuous recharge of at least a few percent of the reservoir mass which is inconsistent with a large reservoir model (Section 2.2.3). We find evidence for similar geochemical features in other CFBs including CAMP, Siberian Traps, Columbia River Basalts, and Karoo-Ferrar flood baslts. Finally, the Black and Manga (2017) style model predicts an increasing contribution from crustal assimilation with increasing thermal maturity of the upper crust. In contrast, the least contaminated DT lava flows constitute the Ambenali formation, which erupts towards the end of the sequence.
- Geochronology, Paleomagnetic, Mercury, Lava flow morphology (Section 3.2, 3.3, 3.4, 3.5) : The eruptive tempo of the Western Ghats sequence does not show any evidence for multiple 100 kyr hiatuses between eruptive time-periods contrary to the expectations of a large magma reservoir model. For a single WG composite section (spanning Kalsubai, Lonavala, and Wai subgroups), the typical time-period between eruptions is every 4000-6000 years (Section 3.3). Using Hg results, which integrate over the whole DT province, we find that individual eruptive episodes lasted approximately 100 to 1000 years and erupted around $\sim$ tens to hundred cubic kilometers of basalt per year. This conclusion is supported by eruptive estimates from lava flow morphology from Columbia River Basalt. These observational constraints on eruptive tempo do not match the large magma chamber model predictions. Furthermore, there is no clear eruptive/shallow sill emplacement hiatus for other CFBs with sufficiently high resolution datasets (Siberian Traps, Columbia River Basalts).

- Intra-Flow variation (Section 4.4) : Based on a few studies, there is evidence for isotopic, geochemical, and petrological variations within a single flow as well as composite dikes in the Deccan Traps and the Columbia River Basalts. These observations clearly suggest that a single eruptive episode was not necessarily fed by a homogeneous, well-mixed magma reservoir.
- Dike swarm distribution (Section 5) : No evidence for a single/few large magma reservoir sourcing the Deccan dike swarms since the spatial pattern of dikes is inconsistent with the crustal stress field from such a magmatic architecture. Instead, the spread in dike orientations and dike-dike intersection pattern, especially for the Deccan Central dike swarm, suggests a heterogenous (and likely time-dependent) crustal stress field from multiple magma reservoirs.
- Geophysical datasets (Section 6) : No geophysical evidence for a large, connected, upper crustal magma reservoir. We do find evidence for a large Moho-depth mafic underplating layer consistent with the presence of a large Moho-depth magma reservoir. However, based on the exposed section of analogous systems (e.g., Seiland Igneous Province), we conclude that a thick mafic layer inferred using geophysical methods does not necessarily imply a single magma body of equivalent size.

We posit that the most plausible magmatic structure that may be able to explain these (and other) observations is a multiply-connected magma reservoir model, with each reservoir undergoing REAFC-type processes. Driven by a large and spatially extensive mantle melt flux, the CFB magmatic system likely consists a number of small - medium ($<$ few thousand km$^3$) sized magma bodies instead of one/couple of large magma chambers. This new model of CFB magmatic architecture also likely has signficant implications for the impact triggering hypothesis for the most voluminous Deccan eruptions (M. A. Richards et al., 2015) since the stability of a multiply-connected magmatic system to seismic perturbations will be very different (likely higher) than the previous CFB architectures (Black & Manga, 2017) (low sensitivity to mantle flux variations). However we note that with existing studies, it is unclear if this proposed architecture can quantitatively explain the eruptive tempo associated with CFBs. In the Part II of this study, we use theoretical magma reservoir models to illustrate that this magmatic architecture can indeed describe the observations from Deccan Traps and other CFBs.

## 8 Tables

Table 1: Geophysical Estimates of Intrusive/Extrusive Ratio for Deccan Traps

| Count | Paper reference | Deccan Traps Region | Category | I/E ratio |
|---|---|---|---|---|
| 1 | O. Pandey et al. (2009) | Western Ghats | Moho-Depth Intrusives | 4.34 |
| 2 | O. Pandey et al. (2009) | Central Deccan | Moho-Depth Intrusives | 0.83 |
| 3 | Vedanti et al. (2018) | Central Deccan | Moho-Depth Intrusives | 3.33 |
| 4 | Vedanti et al. (2018) | Central Deccan | Mid-Crustal Intrusives | 10.00 |
| 5 | G. S. Rao et al. (2018) | Western Ghats | Moho-Depth Intrusives | 3.49 |
| 6 | G. S. Rao et al. (2018) | Western Ghats | Moho-Depth Intrusives | 2.12 |

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|---|---|---|---|---|
| 7 | G. S. Rao et al. (2018) | Saurashtra | Moho-Depth Intrusives | 6.07 |
| 8 | P. V. Kumar et al. (2018) | Saurashtra | Total Crustal Section | 2.78 |
| 9 | Tewari et al. (2018a) | Narmada Son Lineament-Central India | Mid-Crustal Intrusives | 7.48 |
| 10 | Tewari et al. (2018a) | Narmada Son Lineament-Central India | Total Crustal Section | 14.56 |
| 11 | Tewari et al. (2018a) | Narmada Son Lineament-Central India | Total Crustal Section | 14.32 |
| 12 | Tewari et al. (2018a) | Narmada Son Lineament-Central India | Moho-Depth Intrusives | 15.00 |
| 13 | Tewari et al. (2018b) | Western Ghats | Total Crustal Section | 3.06 |
| 14 | Tewari et al. (2018b) | Western Ghats | Moho-Depth Intrusives | 1.64 |
| 15 | Seshu et al. (2016) | Kutchh | Mid-Crustal Intrusives | 0.65 |
| 16 | Seshu et al. (2016) | Kutchh | Mid-Crustal Intrusives | 0.30 |
| 17 | Seshu et al. (2016) | Kutchh | Mid-Crustal Intrusives | 0.54 |
| 18 | Seshu et al. (2016) | Kutchh | Mid-Crustal Intrusives | 0.30 |
| 19 | A. Singh et al. (2015) | Narmada Son Lineament Central India | Total Crustal Section | 27.41 |
| 20 | Rajaram et al. (2017) | Koyna Central India | Moho-Depth Intrusives | 1.86 |
| 21 | Tewari et al. (2018b) | Saurashtra | Total Crustal Section | 9.27 |
| 22 | Tewari et al. (2018b) | Cambay Rift Zone | Moho-Depth Intrusives | 6.92 |
| 23 | Tewari et al. (2018b) | Cambay Rift Zone | Moho-Depth Intrusives | 6.55 |
| 24 | Tewari et al. (2018b) | Cambay Rift Zone | Moho-Depth Intrusives | 4.61 |
| 25 | Naganjaneyulu and Santosh (2010) | Cambay Rift Zone | Total Crustal Section | 5.15 |
| 26 | Naganjaneyulu and Santosh (2010) | Narmada Son Lineament Central india | Total Crustal Section | 4.97 |
| 27 | Azeez et al. (2017b) | Narmada Son Lineament Central india | Total Crustal Section | 9.71 |
| 28 | Naidu and Harinarayana (2009) | Narmada Son Lineament Central india | Moho-Depth Intrusives | 7.83 |

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|---|---|---|---|---|
| 29 | Patro and Sarma (2016b) | Narmada Son Lineament Central india | Total Crustal Section | 5.06 |
| 30 | Patro and Sarma (2016b) | Narmada Son Lineament Central india | Total Crustal Section | 8.05 |
| 31 | Patro and Sarma (2016b) | Narmada Son Lineament Central india | Total Crustal Section | 6.03 |
| 32 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 2.50 |
| 33 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 0.85 |
| 34 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.69 |
| 35 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.72 |
| 36 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 2.10 |
| 37 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.33 |
| 38 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.23 |
| 39 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 0.65 |
| 40 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.10 |
| 41 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 3.86 |
| 42 | Prasad et al. (2018) | Narmada Son Lineament Central india | Mid-Crustal Intrusives | 1.68 |
| 43 | Krishna et al. (2002) | Western Ghats | Moho-Depth Intrusives | 2.90 |
| 44 | Krishna et al. (2002) | Saurashtra | Moho-Depth Intrusives | 8.54 |
| 45 | B. Singh et al. (2014b) | Saurashtra | Total Crustal Section | 15.14 |
| 46 | B. Singh et al. (2014b) | Saurashtra | Total Crustal Section | 34.45 |
| 47 | B. Singh et al. (2014b) | Saurashtra | Mid-Crustal Intrusives | 5.94 |
| 48 | B. Singh et al. (2014b) | Saurashtra | Total Crustal Section | 30.33 |

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|---|---|---|---|---|
| 49 | Blanchard et al. (2017) | Saurashtra | Total Crustal Section | 5.51 |
| 50 | Chouhan et al. (2020) | Cambay Rift Zone | Moho-Depth Intrusives | 10.22 |
| 51 | Chouhan et al. (2020) | Cambay Rift Zone | Moho-Depth Intrusives | 10.05 |
| 52 | Chouhan et al. (2020) | Cambay Rift Zone | Moho-Depth Intrusives | 11.18 |

Table 1: Geophysical Estimates of Intrusive/Extrusive Ratio for Deccan Traps

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