Plume-related regional pre-volcanic uplift in the Deccan Traps: Absence of evidence, evidence of absence

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ABSTRACT

The mantle plume model predicts one to few kilometers of regional, domal lithospheric uplift 5-20 million years before the onset of flood basalt volcanism. The uplift results from heat conduction out of and dynamic support by the hot, buoyant, rising plume head. Field evidence for such uplift would comprise sedimentary sequences that reflect progressive basin shallowing before volcanism, or (in case of differential uplift along faults) widespread conglomerates derived from the basement rocks and underlying the first lavas. Local uplifts and subsidences cannot be used to invoke or rule out plume-caused uplift. Over large areas of the late Cretaceous Deccan flood basalt province, the base of the lava pile is in the subsurface. Basalt-basement contacts are observed along the periphery of the province, and in central India (Satpura and Vindhya Ranges), where substantial post-Deccan uplift is evident. Here, extensive, horizontal Deccan basalt flows directly overlie extensive low-relief planation surfaces cut on various older rocks (Archean through Mesozoic) with different internal structures. Locally, thin, patchy late Cretaceous clays and limestones (Lameta Formation) separate the basalts and basement, but some Lameta sediments are known to have been derived from already erupted Deccan basalt flows in nearby areas. Thus, the eruption and flowage of the earliest Deccan basalt lava flows onto extensive, flat planation surfaces developed on varied bedrock, and the near-total absence of basement-derived conglomerates at the base of the lava pile throughout the province, are evidence against pre-volcanic lithospheric uplift (both regional and local), and thereby the plume head model. There has been major (>1 km) *post*-Deccan, Neogene uplift of the Indian peninsula and the Sahyadri (Western Ghats) Range which runs along the entire western Indian rifted margin, well beyond the Deccan basalt cover. This uplift has raised the regional, late Cretaceous planation surface developed *on* the Deccan lava pile to a high elevation. This uplift cannot reflect Deccan-related magmatic underplating, but is partly denudational, is aided by a compressive stress regime throughout India since the India-Asia collision, and possibly also related to active eastward flow of the sub-lithospheric mantle. The easterly drainage of the Indian peninsula, speculated to be dome-flank drainage caused by the plume head, predates the uplift. Field evidence from the Deccan and India is in conflict with a model of plume-caused regional uplift a few million years before the onset of volcanism.

Keywords: Mantle plume; flood basalt; tectonics; uplift; planation surface; volcanism; Deccan

INTRODUCTION: FLOOD BASALTS, PLUME HEADS, AND LITHOSPHERIC UPLIFT

The deep mantle plume model of Morgan (1972, 1981) for flood basalt volcanism was further developed from fluid dynamical experiments and numerical modelling (Richards et al., 1989; Campbell and Griffiths, 1990; Hill, 1991; Farnetani and Richards, 1994; and several others). Such modelling indicated that thermal plumes rising buoyantly from the core-mantle boundary should develop large bulbous "heads" about a thousand kilometers (km) in diameter, by entrainment of surrounding mantle, and the heads remain connected to the source region by narrow "tails". A key premise is that mantle plumes are hotter than ambient mantle, and the large volumes of magma erupted in flood basalt provinces are due to the high-temperature plumes undergoing extensive decompressional melting.

Campbell and Griffiths (1990) calculated that a new ("starting") plume head, 1000 km in diameter, flattens to a disc twice as wide after impinging on the base of the lithosphere (Fig. 1). One key predicted consequence of plume upwelling is that the plume head raises the lithosphere over a broad area (~1000 km or more across) before flood basalt volcanism begins. This is a consequence of heat conduction out of and dynamic support by the hot plume head. Significant domal uplift is predicted 10-20 million years (m.y.) before flood volcanism, when the top of the plume head is still well below the lithosphere (Fig. 1). The uplift is followed by subsidence as the plume head begins to melt extensively and the magmas are transported towards and erupted at the surface. The amplitude of the surface uplift has been calculated at 1-4 km (Campbell and Griffiths, 1990; Farnetani and Richards, 1994) depending on model parameters such as plume excess temperature.

It has been claimed that geological and geomorphological evidence from most flood basalt provinces of the world fullfills the predictions and patterns of the above model (Kent, 1991; Cox, 1989; Campbell, 2005). Here, I examine in detail the evidence from the Deccan flood basalt province and India, and conclude that the model is inconsistent with the field evidence which rules out regional pre-eruption domal uplift in the province. The objective of this chapter is to systematically and clearly lay the field evidence and not to formulate or present an alternative model to the plume model, as such models, e.g., involving continental break-up and rift-related convection and mantle melting, have been presented elsewhere (e.g., Sheth, 2005a).

EVIDENCE REGARDING UPLIFT IN FLOOD BASALTS

Geological evidence for pre-volcanic uplift can take the form of pre-eruption sedimentary sequences that reflect progressive basin shallowing before volcanism (e.g., Rainbird and Ernst, 2001; Mazumder, 2005, Mazumder, www.mantleplumes.org/Dhanjori.html). Alternatively, if there is differential tectonic uplift along faults, rapid erosion of the basement rocks can be expected to produce beds of conglomerate or coarse clastics below the first lavas. If the uplift were regional, such sediments would be expected to have a regional distribution. In the Permo-Triassic Emeishan flood basalt province of China, He et al. (2003) have argued for domal, kilometer-scale regional pre-volcanic uplift based on palaeogeographic shallowing recorded in sediments, and conglomerate horizons underlying the initial basalt lavas. They support the plume head model and reject non-plume mechanisms. Pre-volcanic uplift has also been identified in the East and West Greenland flood basalt provinces of Palaeocene age (e.g., Dam et al., 1998).

However, evidence for pre-volcanic uplift is absent, or contentious, beneath several other flood basalts. In Russia, the Siberian flood basalts of Permo-Triassic age (~250 Ma) overlie the coal-bearing Tungusskaya Series (320-250 Ma) throughout most of their great areal extent (Czamanske et al., 1998). The Tungusskaya Series includes the Tungusska Basin, the largest coal-bearing basin in the world. This suggests continuous crustal subsidence before and during the flood eruptions, which Czamanske et al. (1998) considered incompatible with dynamic support from a plume head beneath the Siberian lithosphere. Subsequently, Reichow et al. (2002) reported and dated extensive subsurface basalts from the West Siberian Basin, considerably enlarging the area of the Siberian flood basalt province. Saunders et al. (2005) argue on this basis that pre-volcanic uplift did occur in the Siberian Traps, not over the present outcrop area of the Traps, but over the West Siberian Basin. They note that this uplift had been ascribed previously to a plate tectonic cause – a Hercynian tectonic event (Peterson and Clarke, 1991), but prefer a plume head as the explanation for the uplift. However, the plume head model predicts lithospheric uplift to *commence* 10-20 m.y. before the flood basalt eruptions and to reach a maximum 5-10 m.y. before the eruptions (Campbell and Griffiths, 1990; Hill, 1991). If the West Siberian Basin had been uplifted for tens of m.y. before the Siberian Trap eruptions, as has been suggested (A. Ivanov, pers. comm., 2005), this would be inconsistent with the plume head model.

In the Miocene Columbia River flood basalt province, the absence of pre-volcanic uplift has been noted by Hales et al. (2005), who observe that mild pre-eruptive subsidence was followed by syn-eruptive uplift of a few hundred meters and a long-term uplift of 2 km. In the 120 Ma Ontong Java oceanic plateau, world's largest, crustal uplift during or immediately preceding the eruptions was negligible or much less than expected. The entire plateau was constructed below sea-level with few sub-aerial eruptions (Tejada et al., 2004), and this observation and several others have prompted reconsideration of the plume model and exploration of alternative models (Ingle and Coffin, 2004; Korenaga, 2005; Clift, 2005).

The absence of uplift before volcanism in some flood basalt provinces has been explained within the framework of the plume model by proposing flattening and spreading of the plume head against a thick, cold and strong lithosphere, or by lateral migration of magma into the crust (Campbell and Griffiths, 1990). These variations make the plume model impossible to falsify or test. It is argued, for example, that uplift patterns can be complex depending on factors such as lithospheric strength and plume excess temperature, buoyancy and shape. It is thus argued that, all plumes do not cause uplift and flood basalt provinces *without* pre-volcanic uplift may well be plume-generated

(Burov and Guillou-Frottier, 2005.) If so, the plume head model makes no specific and testable predictions for field geologists and cannot be evaluated against their observations. The plume head model would become untenable, however, when a flood basalt province shows not merely an *absence of evidence* for pre-volcanic uplift, but also *evidence for absence* of such uplift. This is not only possible, but it is, I argue, the scenario in the Deccan flood basalt province, India.

INDIAN AND DECCAN GEOLOGY IN A NUTSHELL

The Deccan province, of late Cretaceous age (~65 Ma), was associated with the break-up of the Seychelles microcontinent from the Indian subcontinent along India's western margin (Fig. 2). Prior to this, at ~88 Ma, Greater India (India plus the Seychelles) broke off from Madagascar, an event which was associated with massive flood basalt volcanism on Madagascar and relatively minor volcanic-intrusive activity in India (Storey et al., 1995; Pande et al., 2001). The following description of the main features of Indian and Deccan geology (Figs. 2 and 3) is taken from Sheth (2005a), and many relevant references can be found therein. The Indian subcontinent has a rich rock record from early Archaean up to Recent, with at least six Archaean or early Proterozoic cratonic nuclei recognized. These are the Aravalli, Bundelkhand, Singbhum, Bastar (Bhandara) and Dharwar cratons, and the high-grade granulite terrain in the south (Fig. 2). Several major rift zones traverse the subcontinent: the Godavari and Mahanadi rifts in the east, the Cambay rift in the north-northwest, and the Kachchh rift in the northwest. The ENE-trending Satpura mountain range, considered by many a horst block, separates the Narmada and Tapi rifts, and this zone is a major, 1600-km-long, long-active tectonic zone along the central part of India. The Indian rifts are known to run along major Precambrian tectonic trends. The Narmada rift is considered by some to be a Proterozoic protocontinental suture. The western Indian coast and the Cambay rift parallel the NNW-SSE, Proterozoic, Dharwar orogenic trend. Another major Precambrian orogenic trend, the NE-SW Aravalli trend, splays out into two at its southern end: the E-W Delhi trend (along which the Mesozoic Kachchh rift has developed), and the main NE-SW Aravalli trend which continues into the Saurashtra peninsula.

Fig. 3 shows the main rock formations that make up the Indian shield: a large portion of the shield is made up of Archaean and Proterozoic crystalline rocks, and there are many Proterozoic and Phanerozoic sedimentary basins on them. The Deccan basalt pile, which obscures the basement from observation over 0.5 million km^2, is thickest (~2,000 m) along the Sahyadri Range (syn. Western Ghats) beside the west coast and thins progressively eastward and southeastward, such that along the eastern fringes of the province the lava pile is only ~200 m thick or less. Whereas the lava pile in the Western Ghats region and in the interior areas of the province is made up almost completely of fairly evolved, sub-alkaline tholeiitic basalts (Sheth, 2005b), felsic and alkaline magma types are also prominent along the rift zones and along the west coast. Considerable volumes of felsic and mafic tuffs, and rhyolite and trachyte lavas crop out along the coast, e.g., at Bombay (Sheth et al., 2001; Sheth and Ray, 2002). The west coast and the rift zones are also where significant tectonic-structural deformation has affected the lava pile, such as a pronounced seaward-dipping monoclinal flexure (the Panvel flexure) along the west coast (Sheth, 1998). Regional, dominantly N-S-oriented dyke swarms, of basalts, dolerites, lamprophyres, basanites and allied alkaline rocks, crop out along the coast. Significant volumes of felsic rocks and many alkaline complexes (several of which include carbonatites) are also found along the Narmada rift and the Cambay rift. The Narmada-Satpura-Tapi zone also contains major linear doleritic and alkaline dyke swarms that trend ~ENE-WSW.

The pronounced linearity of the west coast and the continental margin suggest structural control. The newly formed western Indian continental margin and the rift zones may have constituted major vent areas for the Deccan lavas, as inferred from abundant mafic dyke swarms and intrusions, high heat flow, and aligned thermal springs (Sheth, 2000). The Deccan basalts continue beyond the west coast and onto the continental shelf. The Cambay rift, and the region offshore of the west coast, are productive oil and gas field regions. Much of the Cambay region is covered by Tertiary and Quaternary sediments (up to 5 km thick), and at places the underlying basalts are known, from seismic data, to be over 4 km thick.

Over a large part of the province, the contact of the Deccan lavas with the pre-volcanic basement is not exposed. The Deccan lavas overlie a complex Archaean and Proterozoic basement along the the southern and southeastern periphery of the province. In the northern and northeastern parts of the province, i.e., central India, they overlie diverse geological formations: the great Vindhyan sedimentary basin (mid-late Proterozoic), the large Gondwana sedimentary basin (Carboniferous to Jurassic-early Cretaceous), late Cretaceous Bagh and Lameta sediments, as also Archaean and early Proterozoic crystalline rocks (granites, gneisses and metasediments). I will now consider the main geological features of the Deccan proper and southern India that have key significance to the plume-related uplift issue.

THE SAHYADRI RANGE (WESTERN GHATS) AND INDIAN PENINSULAR DRAINAGE

Many major Indian rivers originate in the Sahyadri Range (better known as the Western Ghats), not far from the west coast, and instead of draining into the Arabian Sea a few tens of kilometers to the west, they flow for hundreds of kilometers eastward to the Bay of Bengal (Fig. 4). Cox (1989) speculated that this pronounced easterly drainage was a consequence of regional domal uplift caused by the Deccan plume head. He noted that the Narmada and the Tapi, two major Indian rivers, flow westward. This was ascribed by him to their exploiting a rift system in the dome. Why such a rift system would produce a westerly drainage (*toward* where the topographically high centre of the uplifted dome would be) is unclear. He also presented drainage maps from the Karoo (South Africa) and the Parana flood basalt provinces, and considered each example as the preserved half of an originally complete dome-flank pattern produced by a plume head. Summerfield (1990) pointed out that the Cox (1989) model ignored aspects such as the Cenozoic drainage development in Africa. It is important to recognize that no quantitative data (such as apatite fission-track ages) have ever been offered in support of the plume-caused doming postulated by Cox (1989), and as I discuss here, the model ignores and is incompatible with well-known geological facts from India.

It should be noted that though the Sahyadri Range is where the Deccan basalt pile is best exposed, the Sahyadri is by no means confined to the Deccan lava field. It constitutes a NNW-SSE-trending and 1500-km-long "Great Escarpment" extending to the southern tip of India, well beyond the Deccan basalt cover (Radhakrishna, 1952; Ollier, 1990; Gunnell and Radhakrishna, 2001). The escarpment parallels the western rifted margin of India, and has been retreating eastward due to erosion since its formation along the line of rifting at ~65 Ma (e.g., Widdowson, 1997a). There is only one break in the escarpment throughout its great length, namely the "Palghat Gap" (Fig. 4). The Gap has a maximum elevation of 300 m above mean sea level, an average width of only 13 km, and is entirely rock-floored without any alluvium cover (Subramanian and Muraleedharan, 1985; Gunnell and Radhakrishna, 2001). Its origin has been controversial. On both sides of the Gap, the Ghats reach great heights of >2.5 km above MSL. The Nilgiri charnockite massif to the north of the Palghat Gap, and the Palni-Kodaikanal massif to the south of it, are made up of Precambrian hypersthene-bearing granites and granulites (the so-called "charnockites") (Rajesh and Santosh, 2004). These mountains are the highest in peninsular India, and rank among the highest mountains in shield areas anywhere in the world (Gunnell and Louchet, 2000). Summits of the Western Ghats and the Karnataka plateau, between the Deccan basalt outcrop and the southern charnockite massifs, also approach 2 km in height (e.g., Bababudan, Fig. 4) and are built of Precambrian metamorphic rocks such as quartzites and gneisses. In comparison, the highest peak of the Western Ghats in the Deccan basalt region, Kalsubai, stands at 1646 m (Fig. 4).

The substantially greater heights of the charnockite massifs may well reflect both the original topography and the greater resistance of charnockite to weathering than basalt, as these charnockites are among the hardest rocks known with an extremely low fracture density (Gunnell and Louchet, 2000). What is most important, however, is that *the entire Sahyadri is the precipitous western edge of an uplifted plateau that has been tilted eastward, and the plateau surface has an aged character – it is an ancient flat land surface that has been rejuvenated in relatively recent (Neogene) times by major tectonic uplift* (Radhakrishna, 1952, 1993; Vaidyanadhan, 1977), and the uplift has continued during the Quaternary (e.g., Powar, 1993; Valdiya, 2001). The Sahyadri, as at Mahabaleshwar (1436 m, Fig. 4) has a very youthful aspect today, with stupendous west-facing scarps and many great waterfalls. Mahabaleshwar sits atop a spectacular, ~1200-m-thick exposed Deccan basalt sequence (Fig. 5), and yet, the top of the Mahabaleshwar plateau represents a regional, low-relief, late Cretaceous palaeosurface or peneplain developed on the uppermost basalts after the cessation of the eruptions, represented by 25-50 m thick laterite (Widdowson, 1997a). In southern India, laterites or bauxites cap the high-elevation summits built of Precambrian rocks. The escarpment has much the same form whether it is developed on the Deccan basalts or on the massive and structureless charnockites in southern India (see Figs. 10.10 and 10.11 in Ollier and Pain, 2000). Sahyadri uplift is evidently not related to Deccan volcanism, but is more appropriately designated as "rift-shoulder uplift". Along with the Sahyadri, the Konkan Plain to the west of it has also been rising during the Tertiary (e.g., Powar, 1993), and places on the west coast, such as Mangalore (Fig. 4) are currently rising (e.g., Subrahmanya, 1996).

As regards the easterly drainage pattern of the peninsula, Widdowson and Cox (1996) showed that in the Mahabaleshwar area, the easterly drainage cuts through the axis of a N-S-aligned, south-plunging anticlinal structure identified from regional dips of the basalt lavas and the laterite cap. The anticlinal structure, they pointed out, developed subsequent to the drainage and had no effect on the drainage lines. Previously, Ollier and Powar (1985) observed that the drainage pattern of peninsular India is dendritic over both the region of the Deccan lavas and the older basement. They therefore suggested that the drainage developed subsequent to the eruption of the Deccan lavas. The newly formed Deccan lava field would have provided a regional slope to the east. However, they also noted that the drainage is *antecedent* to the uplift of the Sahyadri Range, i.e., the easterly drainage existed prior to the rise of the Ghats. The eastward-draining Godavari-Krishna River system also was already in existence by the time the Deccan lavas were in eruption. A thin (tens of meters), 64-Ma-old (Deccan-age) basalt flow sequence at Rajahmundry on the southeastern coast of India (Figs. 3, 4) has strong chemical similarities with some of the southernmost Deccan basalts, and Baksi et al. (1994) have proposed that the Rajahmundry basalts were intra-canyon flows (akin to many in the Columbia River basalt province) that erupted in the southern Deccan and flowed to the southeastern coast of India along existing river systems. Knight et al. (2003) consider this scenario of long-distance surface transport, and that of local eruptions at Rajahmundry, both equally possible. No dykes or intrusions that can be potential feeders to the local lava pile outcrop around Rajahmundry.

The easterly drainage of the Indian peninsula is also antecedent to the uplift of the *Eastern* Ghats, according to Ollier and Powar (1985). The Eastern Ghats, made up primarily of Precambrian charnockites, schists and other metamorphic rocks, form disconnected hill ranges in eastern India and are thus unlike the Western Ghats, though in southeastern India the Eastern Ghats have an aspect similar to that of the Western Ghats, and major rivers such as the Cauvery are antecedent to them and have cut steep gorges through them (Ollier and Powar, 1985). However, even the easterly direction of drainage in southern India may be an oversimplification, as there appears to be a distinct E-W axis of cymatogenic uplift (crustal arching) at 13^{o}N latitude, to the north of which the rivers flow NE, and to the south of which the drainage is to the SE (Subrahmanya, 1994; Fig 4). Mangalore on the west coast and Madras on the east coast are both located on this axis, and both are actively rising relative to the sea (Bendick and Bilham, 1999). The uplift of the Sahyadri (Western Ghats), and the attendant uplift of the Konkan Plain to the west of it, are definitely post-Deccan (Radhakrishna, 1952, 1993; Widdowson and Cox, 1996; Widdowson, 1997a), and both offshore sedimentary evidence and apatite fission-track data that have recently become available are consistent with this (Gunnell and Gallagher, 2001; Gunnell et al., 2003). The easterly drainage of the Indian peninsula is older than the uplift of the Ghats (both the Western and the Eastern), and the uplift of the Ghats is much younger than Deccan volcanism. The proposition of Cox (1989), that the easterly drainage is a *result* of a plume-generated lithospheric dome, is therefore without basis.

DECCAN BASALT – BASEMENT CONTACTS: CLUES TO PRE-VOLCANIC UPLIFT, OR LACK THEREOF

Over most of the Deccan province today the lava-basement contact is in the subsurface, often at considerable depth. There is a huge thickness (~1700 m exposed) of Deccan basalts in the Sahyadri region, and an additional ~500 m in the subsurface, as identified from seismic data (Kaila et al., 1981). Seismic data also suggest two linear, Mesozoic sedimentary basins in the subsurface *below* the Deccan basalts in the Narmada-Tapi region (Kaila, 1988; Sridhar and Tewari, 2001). The northern, Narmada basin is 1000 m thick and the southern, Tapi basin is 1800 m thick. To the north of the Narmada River, Deccan basalts overlie the mid-late Proterozoic sediments of the Vindhyan Basin, or older crystalline basement directly. The Vindhyan Basin does not extend to the south of the Narmada River. To the south of the river, in the Satpura Range, Deccan lavas overlie the thick Gondwana sedimentary basin (Carboniferous to Jurassic-early Cretaceous), or the Precambrian basement. The Gondwana Basin does not extend to the north of the Narmada River, and is particularly well-exposed in the Pachmarhi area (Fig. 3) due to substantial post-Deccan uplift along basin boundary faults.

Two other sedimentary rock formations – the Bagh and Lameta Formations – underlie the Deccan basalts locally in central India and Satpura region (e.g., Mohabey, 1996; Sahni et al., 1996; Tandon, 2002; Khosla and Sahni, 2003). Both are of late Cretaceous age, and much smaller than the Vindhyan and Gondwana Basins. Both are also known in Indian literature as "infra-trappeans" and locally separate the Deccan basalts from the Vindhyans, the Gondwanas, or the Archaean-Proterozoic crystalline basement, as the case may be. The Bagh Formation consists of sandstones and limestones formed during a late Cretaceous marine *transgression* in the western Narmada valley (Sheth, 2005a), the region where one would expect maximum uplift from a putative Deccan plume head at this exact time. The fluvial and lacustrine Lameta Formation with its type locality near Jabalpur (Fig. 3) is also of late Cretaceous age, and consists of thin (~20 m) limestones and clays. Its maximum thickness is 40 m near Jabalpur (Tandon, 2002). The Lametas overlie the Gondwana sediments or Precambrian rocks, and are themselves overlain by the Deccan basalt flows in some sections, based on which they were considered an older formation than the basalts. More recent field, geochemical and clay mineralogical studies have indicated, however, that the Lameta clays in the Jabalpur region were derived from the Deccan basalts themselves, and thus, basalts were already erupting in nearby areas and supplied material for these sediments which are of restricted lateral and vertical extent (Prasad and Khajuria, 1995; Salil et al., 1997; Tandon, 2002; Shrivastava and Ahmad, 2005). Clearly, the Lameta beds in some sections pre-date the Deccan basalts, and the Lameta beds in other sections are derived from (and so younger than) other Deccan basalts.

Pre-eruption uplift is recorded in the Lameta sediments of the Dongargaon Basin of the Nagpur region (Fig. 3), where Tandon (2002) recorded a clear "shallowing up" trend from shallow lake deposits to a palaeosol before the terrain was buried by the first Deccan basalt flow. He ascribed this, however, to pre-volcanic surface uplift of the area of the order of *meters only* (my italics) and possibly also to "mock aridity". Mock aridity (Harris and Van Couvering, 1995) refers to extreme local aridity resulting from active volcanism (e.g., Kilauea) while the region as a whole may be under humid tropical conditions (Hawaii). Evidence for mock aridity has been found in "inter-trappean" sediments interlayered with (not below) the Deccan basalt flows at Anjar, in the northwestern part of the province (Khadkikar et al., 1999; Fig. 3). If mock aridity was already at work when the thin, shallowing-upward Lameta sequence at Dongargaon was forming (Tandon, 2002), then volcanism was already active nearby.

To conclude, there were both local uplifts and subsidences just before volcanism in parts of central India, but local and vertically restricted uplift (as recorded in the Lameta beds of Dongargaon Basin) cannot be used to support the plume model, and local late Cretaceous subsidence and marine invasion (as in the Bagh area, Fig. 3) cannot be used to refute it. Such local tectonics are easily related to local processes such as the filling and emptying of magma chambers, emplacement of plutons and sills, and faulting. Relevant in this connection is a basalt- and basement-derived conglomerate bed reported in the Rajpipla area (Fig. 3) that unconformably overlies tilted late Cretaceous sediments and is overlain by the Deccan basalts (Widdowson, 2005). Rajpipla is located near the edge of the Deccan basalt outcrop and the basement is exposed in the vicinity. This outcrop suggests that uplift and/or tilting and erosion of the late Cretaceous sediments occurred in quick succession just before volcanism. A basalt flow, or possibly several flows, were then erupted nearby, and continuing uplift and erosion produced the conglomerate, which was in turn covered by the subsequent lava flows. This appears a case of local uplift such as can be produced by moderately large intrusions, or faulting. Dykes, intrusions and plugs have been known from Rajpipla (e.g., Krishnamurthy, 1971), and only 40 km to the east is the Amba Dongar carbonatite-alkaline complex, where the late Cretaceous marine Bagh sandstones have been domed up by an intrusive basalt plug (Ray et al., 2003). The Rajpipla conglomerate outcrop appears very similar to conglomerates interlayered with some Palaeogene basalt flows on the Isle of Skye, Scotland. At Preshal Beg on Skye, a spectacularly columnar-jointed olivine tholeiite lava flow (of maximum thickness 120 m) of the Talisker Bay Group is underlain by a conglomerate of the Preshal Beg Conglomerate Formation (Fig. 6). The conglomerate is a very poorly sorted and chaotic deposit, and was derived locally from the nearby volcanics by local uplift and probably slope instabilities (Williamson and Bell, 1994; Emeleus and Bell, 2005).

My own field work in the Pachmarhi-Chhindwara-Nagpur region (Fig. 3) reveals that the Deccan lavas directly overlie a varied basement without any intervening coarse clastics of late Cretaceous age. Kale et al. (1992), with field data from the southernmost fringes of the Deccan (e.g., around Phonda and Shahbad, Fig. 3), have reported a considerable variation (~200 m) in the elevation (above MSL) of the basalt-basement contact, based on which they have argued that the distribution of the lava flows was strongly controlled by pre-existing topography. They have not reported any basement-derived conglomerates under the lavas, however, and this indicates absence of local, immediately-pre-eruption uplift. Nor have Choubey (1971) or Dixey (1970) reported such clastics, from central India as a whole. Sections in the Deccan province with strata that reflect pre-volcanic uplift are, thus, very few in number, and the uplift here does not exceed a few meters. What, then, is the evidence in central India for or against regional pre-eruption uplift? To understand this we must first understand the geomorphological concept of a planation surface and its tectonic underpinnings.

PLANATION SURFACES: A KEY CONCEPT IN TECTONIC GEOMORPHOLOGY

Planation surfaces, discussed at length by Ollier and Pain (2000), are regionally extensive, flat or nearly flat surfaces produced by advanced erosion (usually fluvial) of any earlier topography to the existing base level of erosion, which is in most cases the sea level. (Large and high inland plateaus like the Tibetan Plateau have their own base level, and a large river may act as a sub-regional base level.) Planation surfaces are also called *erosion surfaces* or *peneplains*, and these terms will be used here interchangeably. *Palaeosurfaces* (Widdowson, 1997b) constitute a broader category of ancient land surfaces that can be exogenic (erosion/weathering) or endogenic (e.g., lava plains). A planation surface is therefore a kind of exogenic palaeosurface.

On a peneplain, relief is minor to absent, and erosion is therefore negligible, but deep in situ weathering and lateritization are typical. Summerfield (1991) states that low relief, absence of erosion, and sediment starvation are vital prerequisites for the development of indurated palaeosols of the nature of ferricretes or laterites. A planation surface thus indicates long-term tectonic stability, and the absence of uplift and erosion. If a low-lying peneplain is uplifted to form a plateau, the dissection of the plateau by further erosion works towards forming a younger planation surface at a lower level, broken only by some surviving remnants ("monadnocks") of the higher surface (Fig. 7a). Also, because there is no process that can produce a flat erosional surface out of a jagged terrain at a high elevation, the presence of a geologically young planation surface at high elevation today suggests rapid and recent uplift (Ollier and Pain, 2000).

It is important to be able to recognize planation surfaces correctly. First, planation surfaces are products of bedrock erosion, not sediment deposition. A vast, flat alluvial plain is not a planation surface. A planation surface is difficult to prove on horizontally-bedded rocks, because these tend to produce flat surfaces on erosion anyway. These are purely structural surfaces. A planation surface is most easily recognized when it cuts across, or bevels, diverse rocks with varied internal structures (Fig. 7b). The presence of bevelled cuestas over a wide area of folded or tilted rocks is a good indicator of a former planation surface, because in the absence of a former planation surface, a cuesta, which is a structurally controlled landform on dipping strata, would show a sharp ridge crest, and never a level top (Fig. 7c,d). However, even in the absence of bevelled cuestas, the phenomenon of "accordant summit levels" –complexly deformed rocks of widely varying types with their summits at more or less the same general level – suggests the existence of a former extensive land surface that lay somewhat higher, and that is now being dissected again (Fig. 7e). Planation surfaces are of great significance in tectonics, and a key concept in tectonic geomorphology, as Ollier and Pain (2000) show with many examples worldwide. For example, on the basis of the strikingly accordant summit levels seen in many major fold-thrust mountain belts (Himalaya, Alps, Andes), and identifiable planation surfaces and other evidence, they argue that fold-thrust mountains are essentially uplifted peneplains or plateaus, dissected subsequently into the present rugged topography, and the rock deformation reflected in their complex internal structure has nothing to do with the present mountainous topography. Planation surfaces at rifted continental margins have been used to track uplift. For example, Bonow et al. (2006a,b) show that the West Greenland rifted margin contains successive planation surfaces ca. 2 km above present sea level, and that this margin has experienced multiple uplift and erosion events since the culmination of rifting in the Labrador Sea during the mid-Eocene. They conclude that the present relief was formed during the late Neogene. Similarities with the Sahyadri, which has experienced major uplift in the Neogene, are striking.

Fig. 8 shows a schematic NNW-SSE profile (after Gunnell, 1998) through southern India showing the multiple planation surfaces that bevel varied bedrock types and structures (see Fig. 4 for locations of the summits). S_o (~2500 m elevation) is the original, Gondwanic planation surface (late Jurassic or early Cretaceous), S_1 (~2200 m) is a late Cretaceous surface well seen in the Nilgiri massif (around Ooty, Fig. 4) and surrounding highlands, S_2 probably an early- or mid-Tertiary surface (identified partly based on bauxite occurrences and suggested to be the top of the Deccan basalt pile), and S_3 a late Tertiary surface. S_4 is the lowest and youngest surface, Mio-Pliocene in age. Note that post-denudational upwarping has produced a fanning of the surfaces S_1 to S_3, which in the opinion of Gunnell (1998) is probably unrelated to the rift-flank uplift of the Western Ghats, but is related to crustal loading by the Deccan basalt pile and late Neogene intraplate crustal deformation in the Indian Ocean Basin that has been affecting peninsular India as well (Subrahmanya, 1996).

PLANATION SURFACES UNDER THE DECCAN BASALTS

As noted above, the top of the Deccan basalt pile in the Western Ghats, such as at Mahabaleshwar (1436 m), is a heavily lateritized planation surface of late Cretacceous to early Tertiary age (Fig. 5), developed on the uppermost basalts after the eruptions ceased (Widdowson and Cox, 1996; Widdowson, 1997a; Widdowson and Gunnell, 1999). The lateritization is consistent with observations of lateritized planation surfaces worldwide (Ollier and Pain, 2000) and the surface indicates a stable tectonic regime not punctuated by uplift and erosion but marked by advanced rock weathering. Its present high elevation today reflects purely post-Deccan uplift, amounting at least to several hundred meters if it originally lay a few hundred meters above the then sea-level (Widdowson, 1997a), and very probably to over a kilometer. This surface reflects stable tectonic conditions after the eruptions ceased. If a planation surface can also be recognized *below* the Deccan basalt pile in a particular region, however, long-term tectonic stability *before* flood volcanism would be indicated.

This scenario does appear to obtain. Widdowson and Gunnell (1999) note that with the exception of the thin infra-trappean Lameta and Bagh beds that indicate fluvial and shallow marine conditions at some localities in the province during the late Cretaceous, the basalts were erupted onto the pre-Deccan (Gondwanan?) land surface. The surface had, over very large areas, lain exposed to continental weathering for a long time. Evidence for pre-Deccan planation surfaces is best sought in central India, where the basalt-basement contacts are exposed over a large region, and the basement comprises a very varied assemblage of Archaean granites and gneisses, the Proterozoic Bijawar metamorphics (dominantly quartzites), the mid-late Proterozoic Vindhyan sequence (dominantly sandstone-shale-limestone), the Carboniferous to Jurassic Gondwana Supergroup, and the late Cretaceous Bagh and Lameta beds (Fig. 3). Dixey (1970, reprinted in Subbarao, 1999) and Choubey (1971) identified erosion surfaces under the Deccan lavas in central India, and Dixey (1970) commented on their usefulness in deciphering subsequent tectonic movements. Choubey (1971) noted that a significant feature of the landscape of central India was that most Vindhyan and Bijawar hills stand approximately at a level of 590 m, with a remarkable summit accordance. He suggested that these are the remnants of a vast Cretaceous peneplain. The Deccan basalts cap this surface. At lower elevations (440-360 m), the Deccan basalts cap another planation surface, developed on the softer Gondwana sediments. Both surfaces existed before the Deccan eruptions, and represented a long inverval of weathering with a deep lateritic saprolith. Below, I list the main observations from specific areas (Fig. 3).

Katangi-Singrampur

Deccan basalt flows cap Vindhyan rocks near Katangi and Singrampur (Fig. 3, 9), and five different flows with a total thickness of 150 m were recognized by Choubey (1971). A thin Lameta limestone bed separates the lowermost flow from the Vindhyan basement. He observed that nowhere do the basalts occupy valleys in the Vindhyans – they may be expected to, if any topographic relief existed before the eruptions – and the base of the basalts is remarkably flat, always at 590 m, irrespective of whether the basalts lie over the Lameta bed or the Vindhyans directly. These observations led him to suggest that the Vindhyans were completely peneplained before the eruptions of the Deccan lavas, and he considered this a spectacular example of a peneplain.

Sagar

Sagar (Fig. 3) is located near the northern margin of the Deccan outcrop, and the area exposes both Lameta beds and Vindhyans under the basalts (West, 1959; West and Choubey, 1964). Ten lava flows were recognized here by these authors, with a total thickness of 180 m. The base of the flows is at 425 m, with either the Lametas or the Vindhyans directly below. In the area some 25-35 km S-SE of Sagar, Choubey (1971) mentioned exposures of ferruginous laterite ~2 m thick, overlying the Vindhyans and overlain by the Deccan basalts. The laterite was interpreted by him to represent the sub-basaltic erosion surface or peneplain.

Dhura-Jamunia

In this area 40 km east of Sagar (Fig. 3), Choubey (1971) recorded three Deccan basalt flows, which he could correlate with the flows in the Katangi and Sagar areas. The basalts are horizontal, and overlie the Vindhyan rocks.

Jabalpur

In the Jabalpur area (Fig. 3), Choubey (1971) noted a late Cretaceous erosion surface extending over Archaeans, Bijawars and Gondwanas at 440-425 m, over which the (thin) Lametas were deposited, followed by the Deccan basalts. Flat-topped ridges of the Bijawar quartzites here are capped by 7-8 m of lateritic duricrust, which he reasonably interpreted as the remnants of the planation surface.

Dhar forest and Bari

In this area (Fig. 3), Choubey (1971) noted a remarkable summit accordance at 300 m on Archaean gneisses, Bijawar quartzites, and the Vindhyan sandstones, and remarked that these hard rocks, though considerably disturbed, form a flat terrace, most likely the remnant of an uplifted peneplain. He considered this area a fine area for examining an exhumed pre-volcanic landscape.

Monadnocks

100 km south of Sagar, at the village of Bitli (440 m), Choubey (1971) noted a conspicuous isolated hill of nearly vertical beds of Bijawar quartzites, with a very flat top. He suggested this to be an isolated monadnock (surviving remnant of an eroded peneplain). It is tempting to draw a similarity between this, and Ayers Rock, a huge monadnock made up of very steeply dipping feldspathic sandstone strata in a vast, flat, pre-Pliocene (>5 Ma) peneplain in central Australia (Ollier and Pain, 2000, p. 28-30). Together, the regional planation surfaces, the laterite that caps them, and the isolated monadnock, constitute very convincing evidence for long-term tectonic stability before the Deccan eruptions in central India. Yet more critical evidence comes from the Pachmarhi area of the Satpura dome.

THE SATPURA DOME

The sub-Deccan Trap Gondwana sedimentary basin is spectacularly exposed in the Satpura mountain range due to *post*-Deccan cymatogenic uplift, and the region of uplift mimics an ENE-WSW-elongated dome (Venkatakrishnan, 1984, 1987). Pachmarhi town, located on the dome (Fig. 10a,b), sits on a thick (~1000 m) lens of mid-Triassic (Upper Gondwana) Pachmarhi Sandstone containing subordinate shales and conglomerates, and capped by extensive ferricrete. It is arguably the doming that has exhumed these sandstones from beneath the Deccan basalt cover, and the most spectacular feature of the dome is the ENE-WSW-trending, 160-km-long, south-facing, free-face scarp that has been carved into the Pachmarhi Sandstone, with a maximum height of 280 m (Fig. 10a,b). Many outliers of this sandstone lie scattered south of the present position of the scarp, which has been retreating northwards.

The Pachmarhi sandstone strata dip north by 10-15^{o}, and are underlain by the Bijori Formation (Permian, Lower Gondwana) dominated by shales (Fig. 10b). Many large Deccan Trap dolerite dykes and intrusions can be found intruding the Pachmarhis and Bijoris at lower elevations, particularly in the gorges cut by rivers such as the Denwa. The dipping strata are beveled by planation surfaces. Venkatakrishnan (1984, 1987) recognized three distinct planation surfaces in the region (Fig. 10a,b). The lowest and youngest is the Bijori Surface (640 m) that is currently forming, and slopes northwards towards Pachmarhi from the basalt cliffs to the south. Older and higher than the Bijori Surface is the Pachmarhi Surface (920 m), and still older and higher is a planation surface (~1300 m) that is preserved in three accordant sandstone summits of Dhupgarh (1352 m), Mahadeo (1330 m) and Chauragarh (1308 m) that rise over Pachmarhi within a few kilometers of it. Venkatakrishnan (1984) named this the Dhupgarh Surface, and suggested that the Narmada river valley to the north of Pachmarhi has been recurrently acting as a local base level, periodically rising and falling, for streams eroding the plateau.

On Dhupgarh, which is the highest peak in the Satpura Range, a Deccan basalt flow unconformably overlies the north-dipping Pachmarhi Sandstone (Venkatakrishnan, 1984), and Venkatakrishnan (1984) considered the base of the basalt flow – the Dhupgarh Surface – to be the same as the late Cretaceous peneplain described by Dixey (1970), the sub-Deccan Trap erosion surface of Choubey (1971), and probably also the inter-continental Gondwana or African Surface of King (1953). Although the sandstone-basalt contact at Dhupgarh is an unconformity, the unconformity does not mean uplift and erosion right before the Deccan eruptions, because the entire rock record from mid-Triassic to late Cretaceous is missing here. Also, because the unconformity is a regional planation surface, indicating long-term tectonic stability, the missing rock record cannot be ascribed to rapid erosion just before volcanism.

Choubey (1971) regarded the Satpura doming as one of the finest and most spectacular examples of cymatogenic uplift in the world, and Venkatakrishnan (1984, 1987) observed that the multiple planation surfaces here were warped as a result of this post-Deccan uplift. He identified the doming and warping in the Satpura region on the basis of evidence such as the warped contact between the Pachmarhi sandstones and the Bijori shales, reversals of structural dips, warped planation surfaces, cave levels, and the drainage pattern. Major rivers in the region, like the Denwa and the Dudhi (Fig. 10a), maintained their pre-existing courses across the rising plateau, and in the process have cut profound gorges. These rivers are therefore good examples of antecedent drainage (Choubey, 1971; Venkatakrishnan, 1984). The smaller rivers, however, were defeated by the uplift, and therefore occupied litho-structural breaks in the Pachmarhi sandstones, producing the unusual rectangular and barbed drainage pattern (Fig. 10a). A key observation by Venkatakrishnan (1984) is that *the rivers such as Denwa and Dudhi originate well to the south of the present-day scarp, several hundred meters below the Pachmarhi Surface, and have cut their way northward through the evolving Satpura dome in spectacular gorges*. He suggests that the most reasonable explanation of this is drainage antecedence – the rivers were in existence before Satpura uplift, and kept cutting the domal upwarp that was forming in their path. To summarize, the uplift and warping of the Pachmarhi block, with its planation surfaces, is post-Deccan (just as the uplift of the Sahyadri and the Indian peninsula is post-Deccan).

What is the cause of this post-Deccan uplift? The Sahyadri uplift (all along the rifted margin) cannot be related to any magmatic underplating that may have occurred during Deccan volcanism, because the essentially thermal uplift associated with magmatic underplating (Cox, 1980; McKenzie, 1984) decays rapidly after the magmatism (e.g., Clift, 2005). The uplift and deformation of large regions of peninsular India, including the noticeable buckling and upwarping at 13^{o}N latitude and around Pachmarhi, may be related at least in part to the dominantly compressional stress regime throughout peninsular India since the suturing of India and Asia along the Himalaya at ~55 Ma (e.g., Klootwijk et al., 1992; Gordon et al., 1998; Valdiya, 1998), which is causing reactivation of ancient and older weak zones and much recent intraplate seismicity (e.g., Subrahmanya, 1994; Valdiya, 2001; Vita-Finzi, 2002). Sahyadri uplift is denudational in part (Widdowson and Cox, 1996; Widdowson and Gunnell, 1999), and may be potentially related also to active eastward flow of the sub-lithospheric mantle relative to the lithosphere as hypothesized by Doglioni et al. (2003) based on several lines of evidence.

DISCUSSION AND CONCLUSIONS

Campbell and Griffiths (1990) cited Pachmarhi as the center of a broad, uplifted dome produced by their conjectured Deccan plume head, quoting Choubey (1971) that the lava-basement contact at Pachmarhi is over 1 km above sea level. But as many lines of evidence discussed above show unambiguously, the doming of the Pachmarhi region (notably not in a circular pattern but elongated ENE-WSW along the Satpura tectonic trend), and the entire Satpura Range of which it is a part, is not pre-volcanic but post-volcanic. Casshyap and Khan (2000) argued for "pre-Deccan Trap doming" of the Indian landmass, based on sedimentological field studies in the Pachmarhi region, but the timing and location of this doming cannot be related to a putative Deccan plume by any means. They identified three separate uplift events, the youngest of which resulted in late Jurassic to earliest Cretaceous sediments, and these were derived from a source in northwestern India. An uplift event centered on northwestern India, and preceding the late Cretaceous-Palaeocene Deccan volcanism by ~70 m.y., cannot be considered pre-volcanic uplift from the Deccan plume.

I have presented and reviewed above the evidence for planation surfaces below the Deccan basalts in central India. The flatness of the pre-Deccan landscape constructed on various older rocks, the remarkable horizontality of the Deccan basalt flows over long distances, and laterites found on the pre-Deccan landscape, together form quite compelling evidence for pre-Deccan planation surfaces. This evidence, along with a near-universal absence of indicators of pre-eruption uplift throughout the extent of the province, runs counter to the idea that a large plume head upwelled beneath India in the late Cretaceous, producing regional domal uplift, the current drainage pattern, and Deccan flood basalt volcanism. Pre-volcanic regional lithospheric uplift is not to be found in the Deccan, and real field evidence (the planation surfaces) indicates its actual absence. There has been major, kilometer-scale post-Deccan uplift, however, which has brought the pre-Deccan planation surfaces from their originally low elevations (relative to the then-existing base level) to their present high elevations. A very similar scenario has been noted in the flood basalts of Yemen by Menzies et al. (1997), who note that these voluminous basalt lavas overlie thick palaeosols developed on underlying fluvial and marine sediments. The palaeosols are widespread, 5-70 m thick and ferruginous. Following Summerfield (1991), they suggested that these palaeosols indicate both low relief and a lack of intense denudation and sediment starvation. They noted ongoing sediment starvation during the eruption of some 3000 m of volcanics in 6 m.y. with few intervening sediments.

How are we to explain the lack of pre-eruption uplift in flood basalts, as described here? Campbell and Griffiths (1990) argued that pre-volcanic regional domal uplift due to a plume head may not be significant due to lateral migration of magma in the crust, or difficult to recognize as early uplift may be overprinted by later subsidence. These are not satisfactory explanations because uplift is predicted to begin 10-20 m.y. before volcanism, and maximum uplift is expected 5-10 m.y. before volcanism (Campbell and Griffiths, 1990; Fig. 1), when the plume head is still well below the lithosphere and initial low-degree melts may barely have begun to form, far less been injected into the crust. Later subsidence that overprints and erases evidence of early pre-eruption uplift cannot be invoked, because such subsidence is simply not seen in flood basalts like the Deccan and Columbia River basalts, which show major post-volcanic uplift instead (this work; Hales et al., 2005). And this uplift should expose evidence of early, pre-eruption uplift, if there ever was any.

There is a serious need to reconsider the plume head model for the Deccan Traps because few if any aspects of the plume model are compatible with geological and geophysical data from the Deccan province (Sheth, 2005a,b; see also Ravi Kumar and Mohan, 2005). The plume model is also under reconsideration for many other flood basalts (e.g., Ingle and Coffin, 2004; Hales et al., 2005 and several papers in Foulger et al., 2005). In a very recent summary of the plume model, Campbell (2005) claims that all the predictions of the model, including that of pre-volcanic regional uplift, are supported by field data from flood basalt provinces of the world, such as the Emeishan and the Deccan. But pre-volcanic regional uplift in the Emeishan province (He et al., 2003), and in the West and East Greenland provinces (Dam et al., 1998) are more of exceptions than the rule, and even in these provinces, it remains to be seen to what extent the pre-volcanic uplift reflects magmatic underplating as opposed to dynamic support from a thermal anomaly in the mantle.

In conclusion, the uplift of the Sahyadri Range (Western Ghats) of India is not related to a putative Deccan mantle plume: it is not domal, it occurred all along the western Indian rifted margin, well beyond the Deccan lava cover, and is most appropriately considered a rift-shoulder uplift affecting the Deccan basalt pile and the basement rocks equally, and maintained at least in part by denudational unloading (Widdowson, 1997a). Planation surfaces below the basalts reflect long-term tectonic stability prior to volcanism, and their significant elevations today reflect post-Deccan uplift, major uplift having occurred in the Neogene (i.e., <23 m.y.). The uplift has been aided by a compressional regime throughout the Indian shield since about 55 Ma. The easterly drainage of the Indian peninsula is not dome-flank drainage produced by a plume head but is antecedent to the uplift. The uplift of both the Western and the Eastern Ghats postdates Deccan volcanism and the easterly drainage. There is thus not only an *absence of evidence* for pre-volcanic regional uplift in the Deccan flood basalt province, but actual *evidence for absence* thereof. The abundant plume conjecture in the voluminous Deccan literature has no factual basis (see also Sheth, 2005a,b); the plume model is untenable there.

ACKNOWLEDGMENTS

Debate and interaction with colleagues at the 2003 GSA Penrose Conference “Plume IV: Beyond the Plume Hypothesis” (Iceland), and the 2005 AGU Chapman Conference “The Great Plume Debate” (Scotland) were invaluable, as were the field trips associated with both. I particularly thank Gillian Foulger for our extensive interaction over mantle plumes during the last five years. The first version of this paper benefitted considerably from very helpful reviews by Richard Ernst, Mike Widdowson, and an anonymous reviewer. Yanni Gunnell and Colin Pain provided valuable, constructive reviews of the revised manuscript, and Rajat Mazumder and Peter Japsen also offered helpful unofficial comments. Volume editors Gillian Foulger and Donna Jurdy offered editorial comments at both stages that helped improve the presentation.

Figure captions

Figure 1. Basic structure of a starting mantle plume, and the patterns of surface uplift predicted a few million years before, and after, plume head impingement on the lithosphere. Modified after Hill (1991).

Figure 2. Map showing the basic structural framework of the Indian subcontinent, including Precambrian cratons (boundaries approximate), structural trends defined by Precambrian fold belts (primarily Proterozoic), the rift zones (primarily Phanerozoic), and the present outcrop area of the Deccan flood basalt province (shaded). Modified after Sheth (2005a,b). The inset figures show the breakup of Greater India from Madagascar at ca. 88 Ma (Storey et al., 1995; Pande et al., 2001), and the breakup of the Seychelles microcontinent (located at the northern tip of the Mascarene Plateau, black) from India at ca. 65 Ma (Norton and Sclater, 1979).

Figure 3. Sketch-map of western and central India showing its main geological features, main geographic-physiographic features (italicized), and the outcrop of the Deccan flood basalts (shaded). The Bagh and Lameta sediments and the Bijawar metamorphics are too small to show at the scale of the map, but the locations of the Vindhyan and Satpura Gondwana Basins are indicated. Also shown are localities mentioned in the text. Modified after Sheth (2005a).

Figure 4. The main elements of the physiography of the Indian peninsula, based on Ollier and Powar (1985) and Sheth (2005a). Note the pronounced easterly drainage. The outcrop area of the Deccan flood basalts is shaded. The Sahyadri (Western Ghats) escarpment is shown by the heavy broken line, and some major summits of the Ghats are indicated, as are some localities cited in the text. The circumference of the plume head proposed by Cox (1989) and the axis of cymatogenic uplift at 13^{o} S latitude proposed by Subrahmanya (1994) are also shown. The St. Mary's Islands rhyodacite lavas (~85 Ma) along the west coast are older than the Deccan Traps, and related to the India-Madagascar break-up event (Pande et al., 2001).

Figure 5. View of the Sahyadri (Western Ghats) escarpment at Mahabaleshwar, showing the 1200-m-thick, horizontally disposed Deccan flood basalt lava pile. The top of the lava pile is a heavily lateritized late Cretaceous planation surface. Photo by Hetu Sheth.

Figure 6. Chaotic, completely unsorted Preshal Beg Formation conglomerate under a spectacularly columnar-jointed olivine tholeiite lava flow of the Talisker Bay Group, Isle of Skye, Scotland. Person near lower right corner of photo provides a scale. Photo by Hetu Sheth.

Figure 7. Schematic sketches showing (a) Multiple, successive planation surfaces in an area undergoing periodic uplift, i.e., a "polycyclic" landscape. (b) A planation surface cut on complex rocks with a resistant monadnock left behind. (c) Cuestas on dipping resistant rocks normally have sharp crests. View is along strike direction. (d) Bevelled cuestas representing a former planation surface. (e) Accordance of summit levels in a

lithologically diverse and complexly deformed terrain suggesting a former planation surface. Based on Ollier and Pain (2000). The vertical scale is greatly exaggerated in (a), (b), and (c).

Figure 8. Schematic NNW-SSE profile through the polycyclic landscape of southern India, showing the multiple planation surfaces. After Gunnell (1998).

Figure 9. Geological map of the Singrampur-Katangi area (Fig. 3), simplified after Choubey (1971). Note that the top of the gently dipping Vindhyan sandstone is a flat planation surface.

Figure 10. (a) Geomorphological map of the Pachmarhi region on the Satpura dome, showing the Pachmarhi scarp, planation surfaces, structural elements, Pachmarhi Sandstone outliers, drainage lines, and some localities. Innumerable Deccan Trap dykes and intrusions in the sandstones are not shown for clarity. D is Dhupgarh (1352 m), M is Mahadeo (1330 m) and C is Chauragarh (1308 m). Based on Venkatakrishnan (1984). (b) Generalized geological cross-section of the Pachmarhi area showing the main landscape elements, slightly modified from Venkatakrishnan (1987). Note the upwarped planation surfaces, and that the surfaces bevel the internal structure of the Pachmarhi Sandstones and the underlying Bijori Shales. The surfaces are also characterized by laterite/ferricrete development. The highest and oldest planation surface is the Dhupgarh Surface (only approximate in (a), and it is the pre-Deccan-Trap erosion surface of Choubey (1971), and the late Cretaceous peneplain of Dixey (1970).

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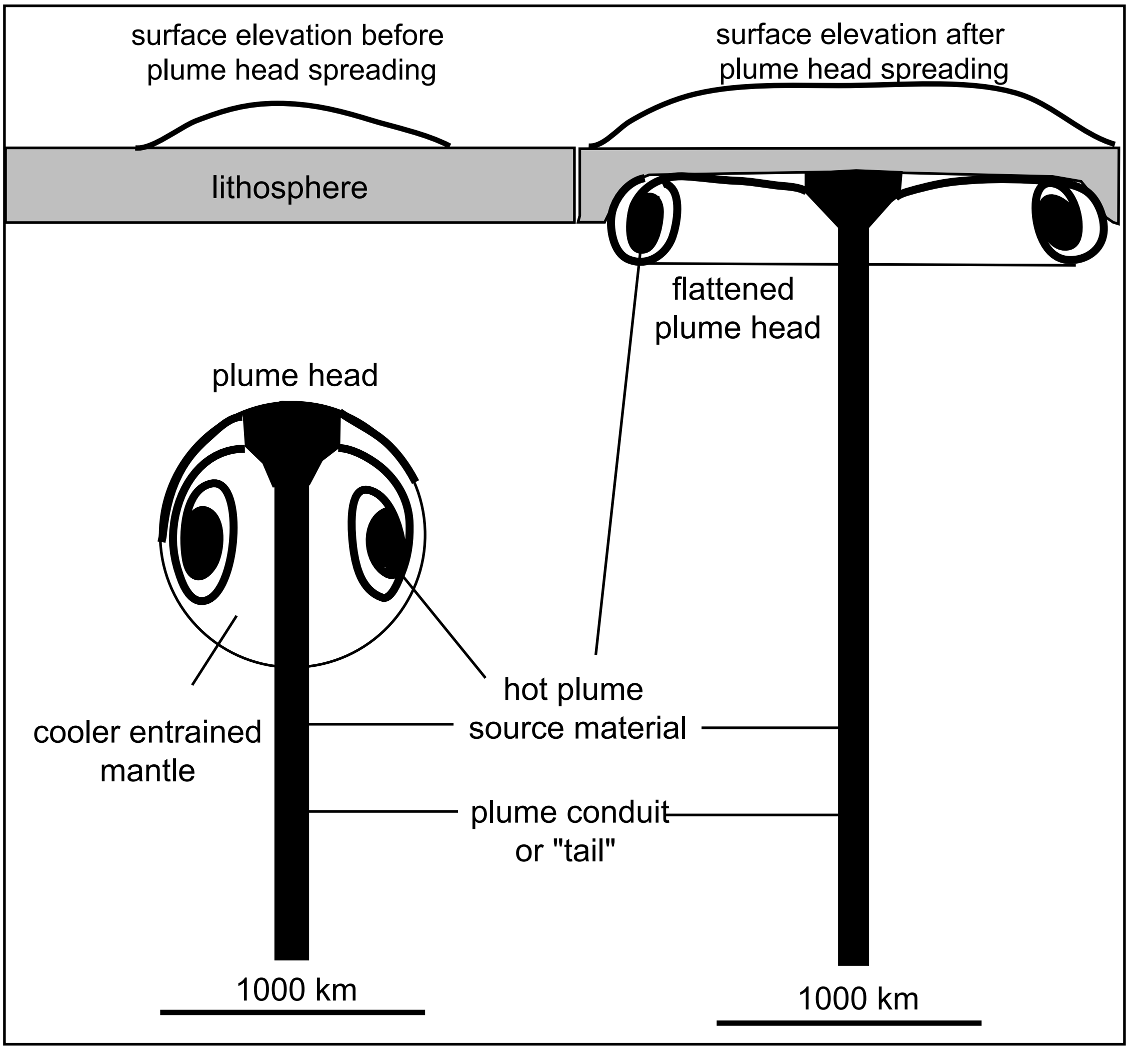

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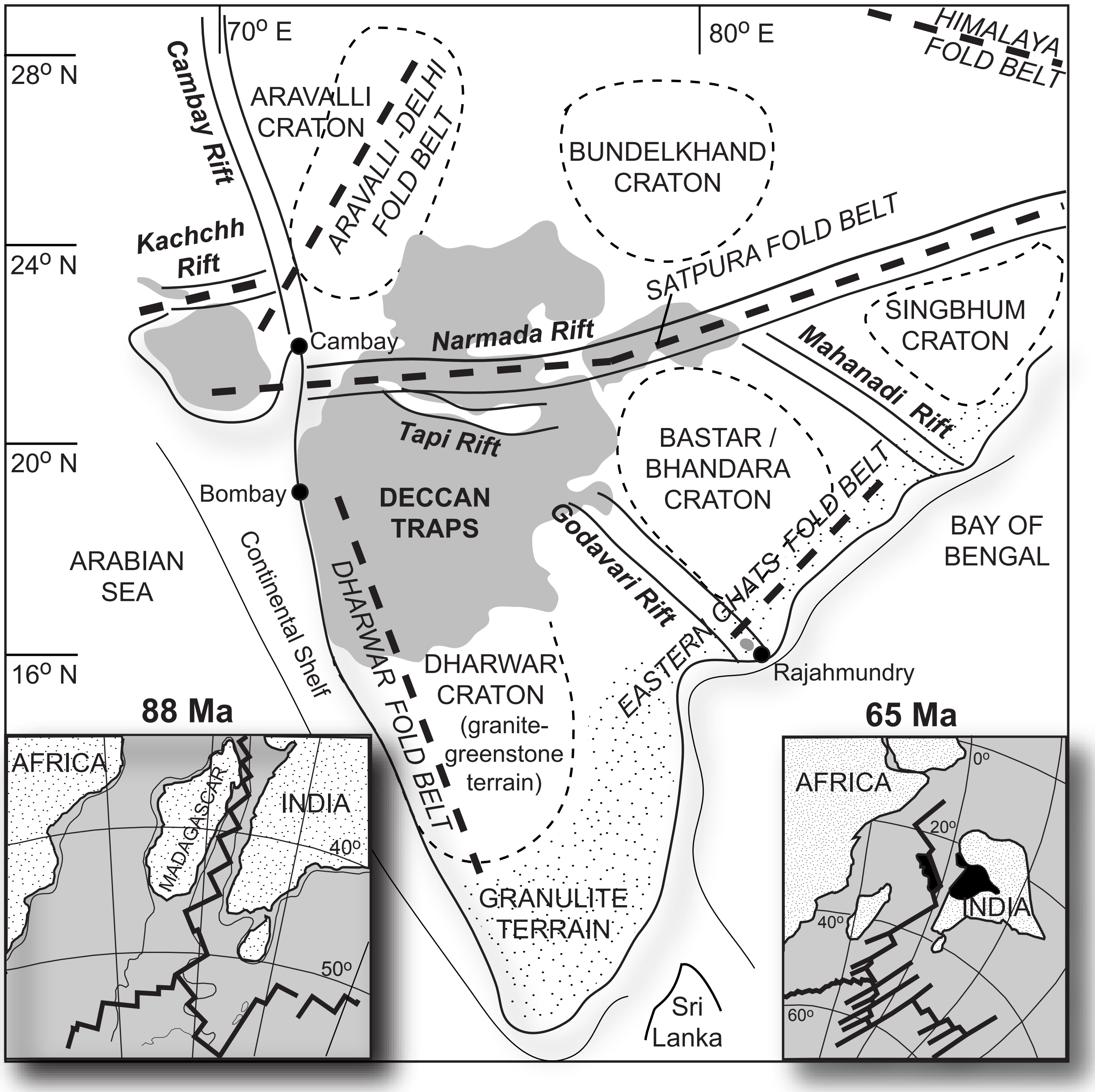

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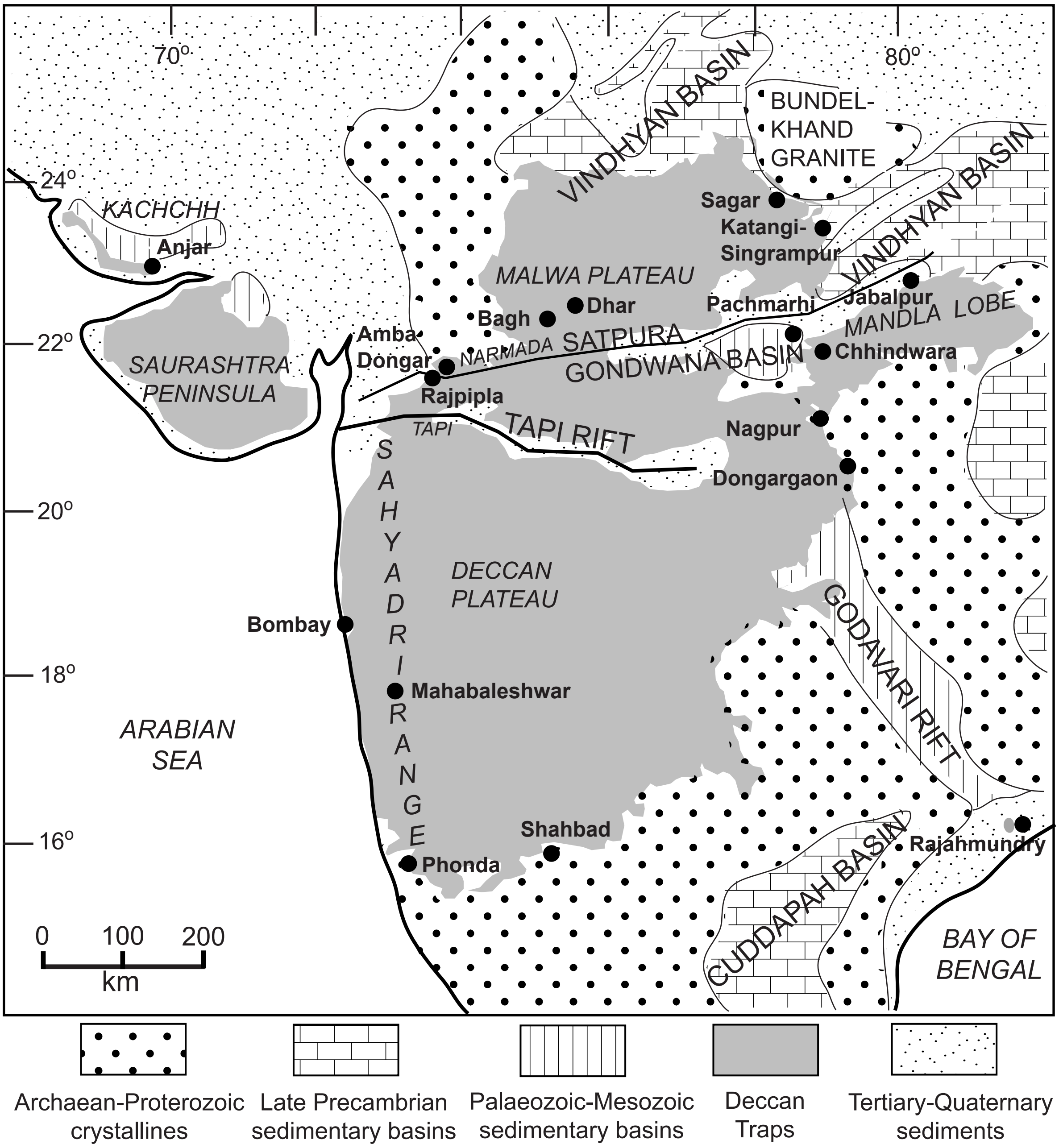

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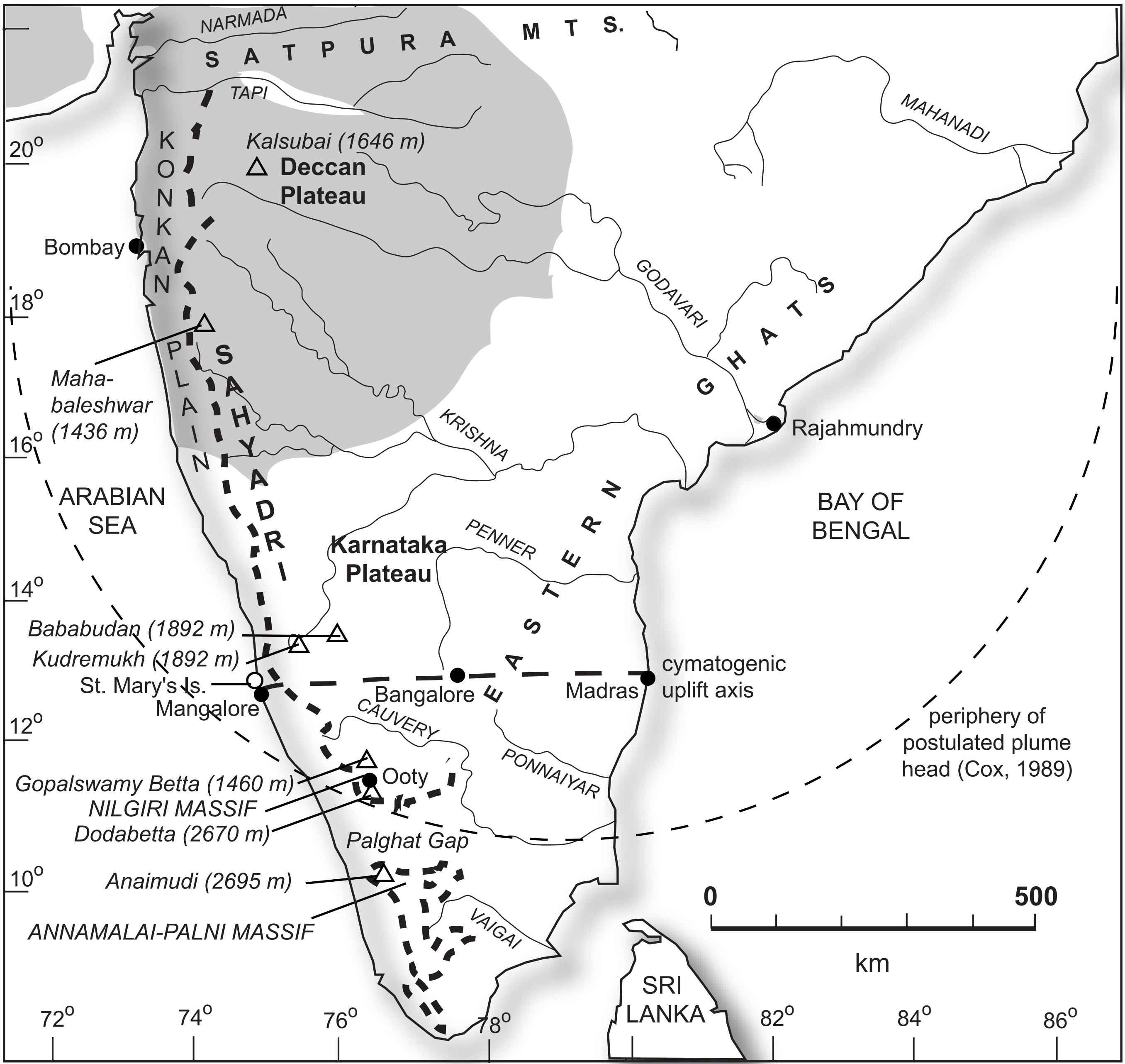

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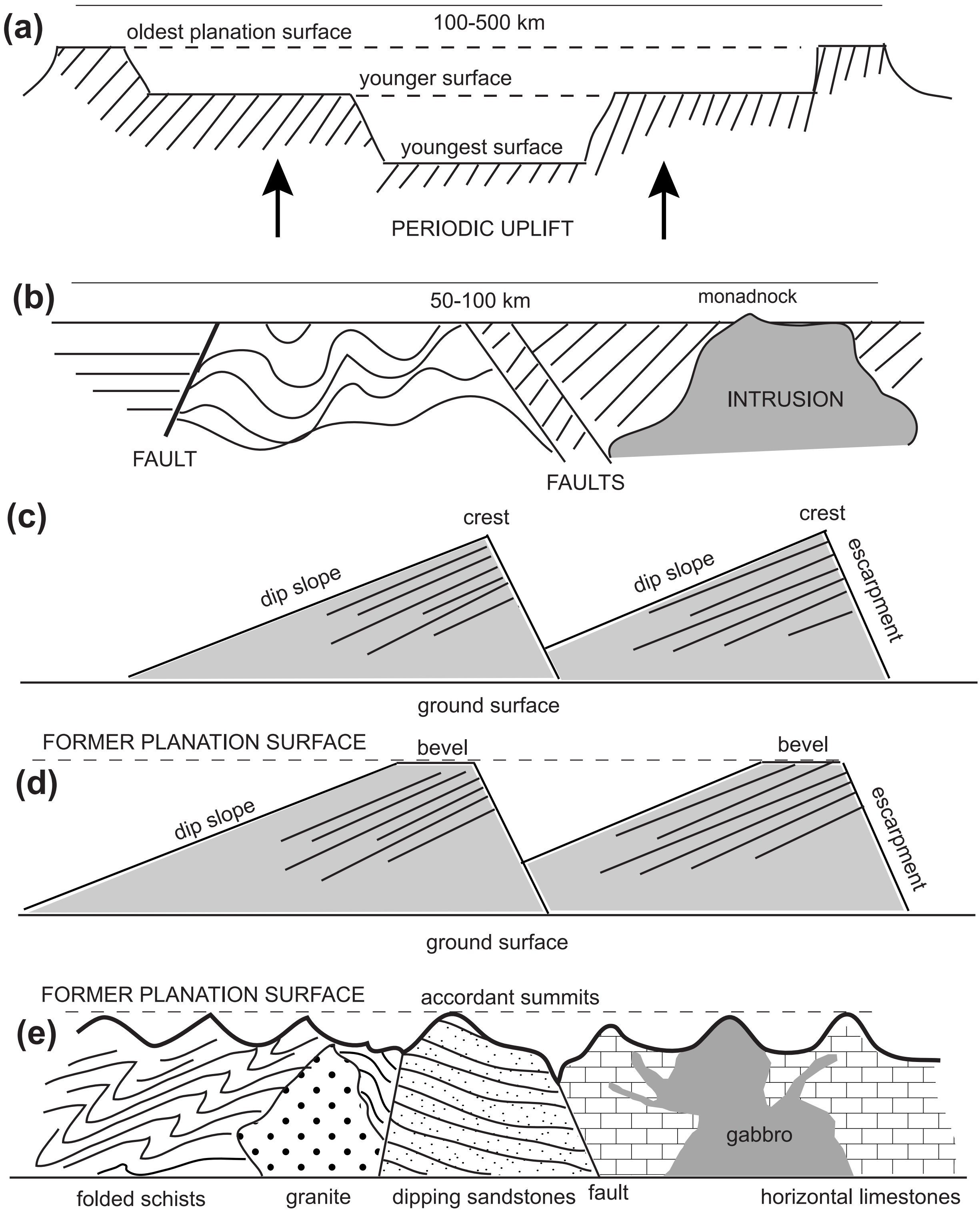

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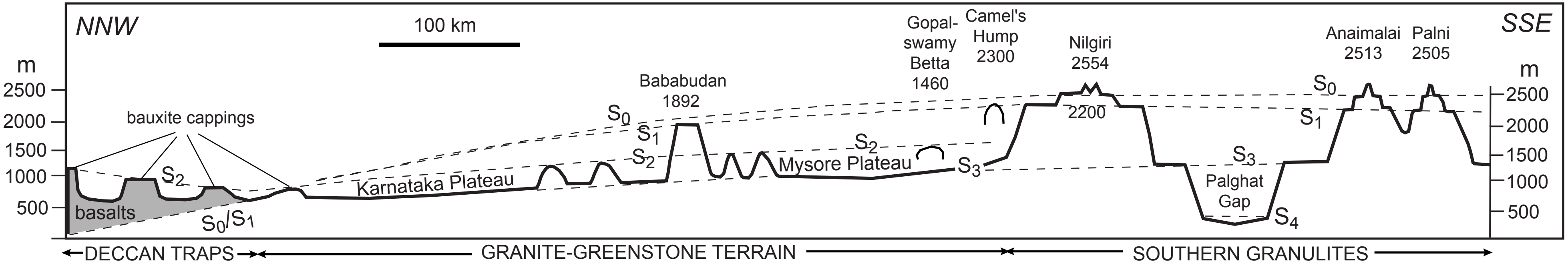

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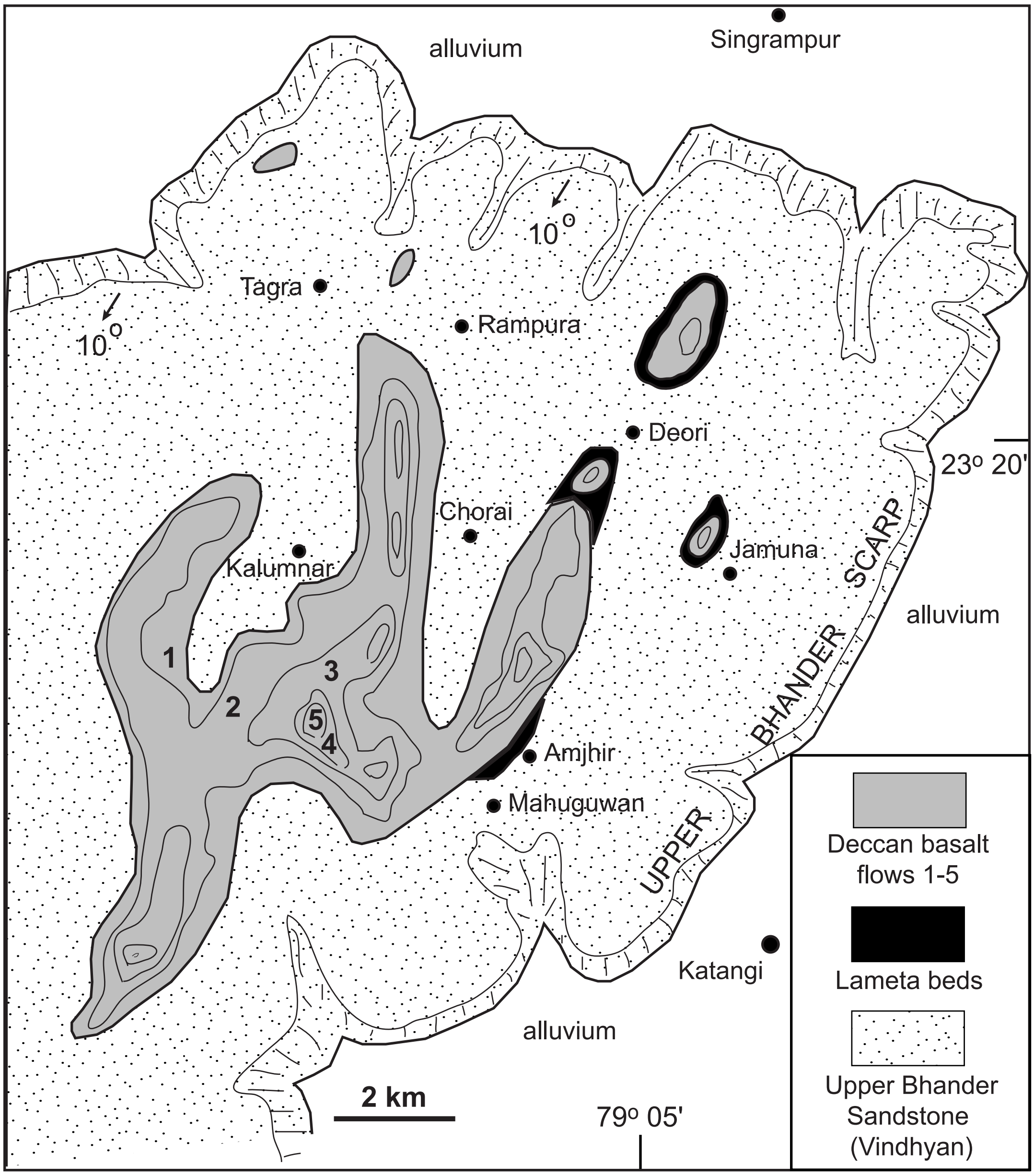

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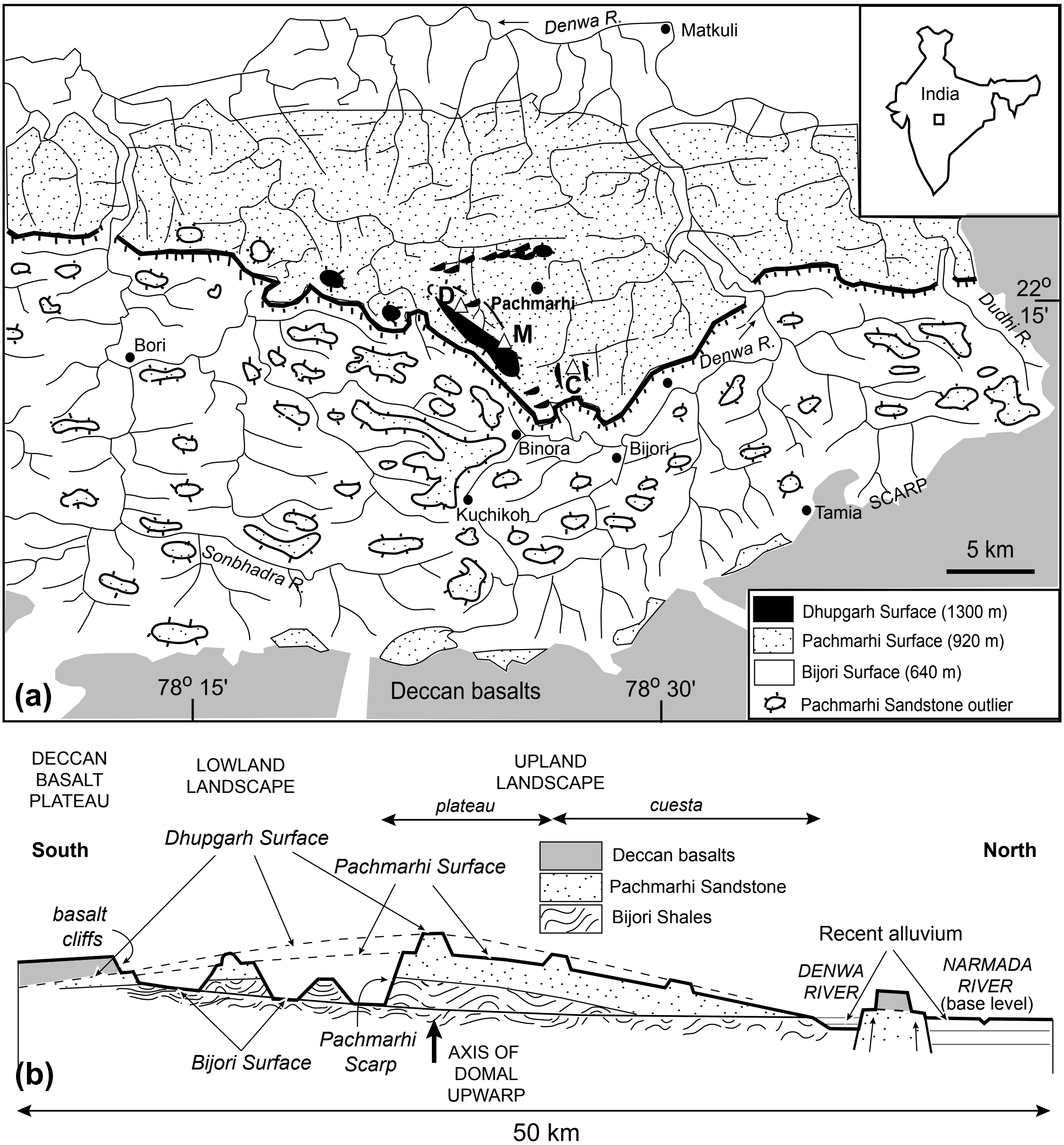

Fig. 10 (H. C. Sheth)