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RESEARCH ARTICLE

Volcanology and eruptive styles of Barren Island: an active mafic stratovolcano in the Andaman Sea, NE Indian Ocean

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Abstract Barren Island (India) is a relatively little studied, little known active volcano in the Andaman Sea, and the northernmost active volcano of the great Indonesian arc. The volcano is built of prehistoric (possibly late Pleistocene) lava flows (dominantly basalt and basaltic andesite, with minor andesite) intercalated with volcaniclastic deposits (tuff breccias, and ash beds deposited by pyroclastic falls and surges), which are exposed along a roughly circular caldera wall. There are indications of a complete phreatomagmatic tephra ring around the exposed base of the volcano. A polygenetic cinder cone has existed at the centre of the caldera and produced basalt-basaltic andesite aa and blocky aa lava flows, as well as tephra, during historic eruptions (1787–1832) and three recent eruptions (1991, 1994–95, 2005–06). The recent aa flows include a toothpaste aa flow, with tilted and overturned crustal slabs carried atop an aa core, as well as locally developed tumuli-like elliptical uplifts having corrugated crusts. Based on various evidence we infer that it belongs to either the 1991 or the 1994–95 eruptions. The volcano has recently (2008) begun yet another eruption, so far only of tephra. We make significantly different interpretations of several features of the volcano than previous workers. This study of the volcanology and eruptive styles of the Barren Island volcano lays the ground for detailed geochemical-isotopic and petrogenetic work, and provides clues to what the volcano can be expected to do in the future.

Keywords Volcanism · Barren Island · Andaman and Nicobar Islands · India · Indian Ocean · Andaman Sea

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Introduction

The Barren Island active volcano lies in the Andaman Sea, northeastern Indian Ocean. It is situated ~70 km east of India's Andaman Islands chain, where sequences of oceanic volcanic and metavolcanic rocks (pillow basalts, ultramafics, serpentinites, greenstones) as well as flysch sediments are exposed (e.g., Allen et al. 2007). The Andaman Trench, along which the NE-moving Indian Plate currently subducts beneath the Burmese Plate, is 250 km west of the volcano. The tectonic scenario is complicated by the presence of active back-arc spreading in the Andaman Sea ESE of Barren Island (Fig. 1; Curray et al. 1979; Kamesh Raju et al. 2004; Khan and Chakraborty 2005; Subba Rao 2008). Barren Island is the only active volcano in Indian territory, and the northernmost active volcano of the great Indonesian arc. To the north of Barren Island are two important dormant volcanoes: Narcondam (India) and Popa (Myanmar). Narcondam, an island volcano ~140 km NNE of Barren Island, may have erupted in the Holocene (Simkin and Siebert 1994; Siebert and Simkin 2002). If so its name (from the Sanskrit "Narak kundam", hell pit) may well be

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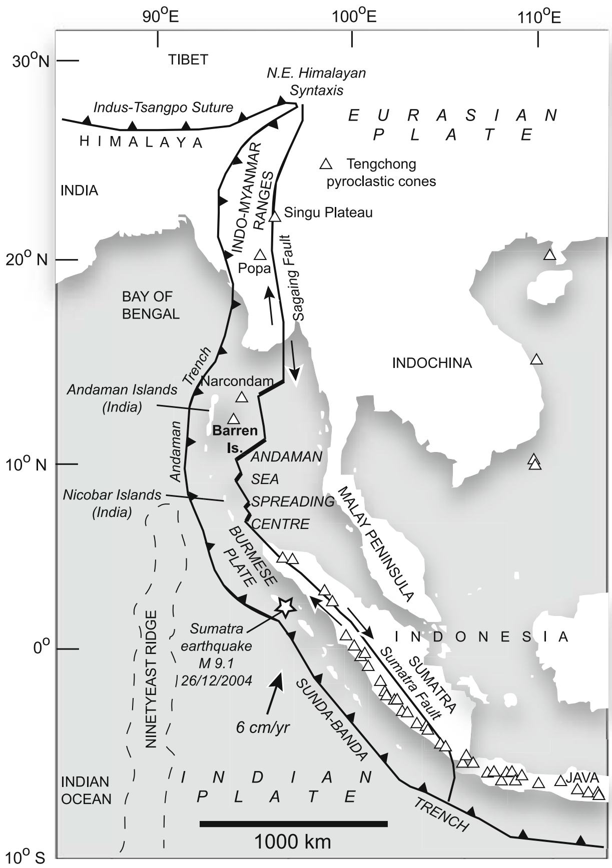


Fig. 1 Map showing the major geological and tectonic features of the northeastern Indian Ocean and southeastern Asia, along with the locations of the Andaman and Nicobar Islands, Barren Island, and Narcondam. *White triangles* are Holocene volcanoes (Simkin and Siebert 1994; Siebert and Simkin 2002). Based on Luhr and Haldar (2006) and references therein

appropriate, rather than having been meant for Barren Island, as Shanker et al. (2001) suggest.

Barren Island is only ~3 km across. It is not all barren and has a lush jungle on its southern and eastern sides, with some freshwater springs (Chandrasekharam et al. 2003). But it is uninhabitated by man, which may explain its name. Wildlife on the island includes feral goats, fruit bats, rats, and parrots. The climate of Andaman Islands (and the contiguous Nicobar Islands just to the south) is described as tropical with only two (rainy and summer) seasons and virtually no winter (Indian Meteorological Department 1999; see also the website of the Government of Andaman and Nicobar Islands at http://andamandt.nic.in/default.htm). The islands experience southwest and northeast monsoons from May to December. Average annual rainfall in the islands is 3,000 mm (30-year average), with 135 days of rain in a year. The mean relative humidity is 79%, and the mean maximum and minimum temperatures 30.2°C and 23.8°C.

Barren Island has restricted public access and can be reached by Indian Navy or Indian Coast Guard vessels from Port Blair (135 km), the capital of the Andaman and Nicobar Islands. A few scientists from the Geological Survey of India (GSI) have been studying the volcano for nearly two decades; however, with rare exceptions their studies have been published as the GSI's internal reports and as conference abstracts, not easily accessible (e.g., Haldar 1989; Haldar et al. 1992a, 1994, and several others). The volcano has therefore had quite low visibility internationally. There is also no continuous monitoring of the volcano. A GSI photographic atlas dedicated to the volcano (Shanker et al. 2001) gives some valuable first-hand information on the recent eruptive activity in the form of eyewitness accounts; however, the atlas is targeted in part at the layman, and our identifications and interpretations of several volcanic features on the volcano differ significantly from those offered in the atlas. The few studies by the GSI and others that exist in peer-reviewed Indian and international literature (Haldar et al. 1992b; Haldar and Luhr 2003; Alam et al. 2004; Luhr and Haldar 2006; Pal et al. 2007a) address aspects of magma evolution at the volcano and provide accounts of its recent eruptions.

Present and past eruptive styles if correctly understood can be a guide to the future activity of the volcano. Here, we describe our geological observations and interpretations of the volcanic sequences on Barren Island, based on two field trips to the volcano (January 2007 and April 2008) aboard Indian Coast Guard vessels.

Geology

Barren Island (Figs. 2, 3) is roughly circular with a diameter of ~3 km and represents the topmost part of a submarine volcano rising more than 2 km above the sea floor. In the absence of drilling or dredging no information is available about the rocks that make up its submarine mass. Luhr and Haldar (2006) estimate its submarine volume to be ~390 km^3 based on the bathymetry of Shanker et al. (2001), and the subaerial volume to be only 1.3 km^3. The volcano has a nearly circular caldera of ~2 km diameter, with a breach in the caldera wall on the northwestern side, which has existed since at least 1787 as the earliest sketches of the volcano (by Colebrooke and Captain Blair, reproduced in Shanker et al. 2001) show. A cinder or scoria cone rising to about 400 m above sea level exists roughly at the caldera centre. The caldera wall exposes prehistoric volcaniclastic deposits (terminology of White and Houghton 2006) and lava flows, which are interbedded with radial outward dips (Fig. 2a, b, c). By "prehistoric" is meant that these deposits formed at some (unknown) time before the first "historic" eruptions which began in 1787 and continued till 1832 (Shanker et al. 2001; Luhr and Haldar 2006). No radioisotopic dating work has been undertaken on Barren

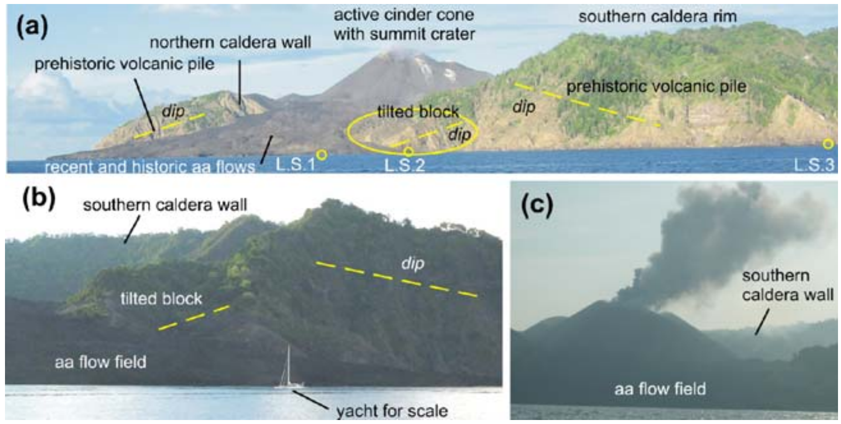
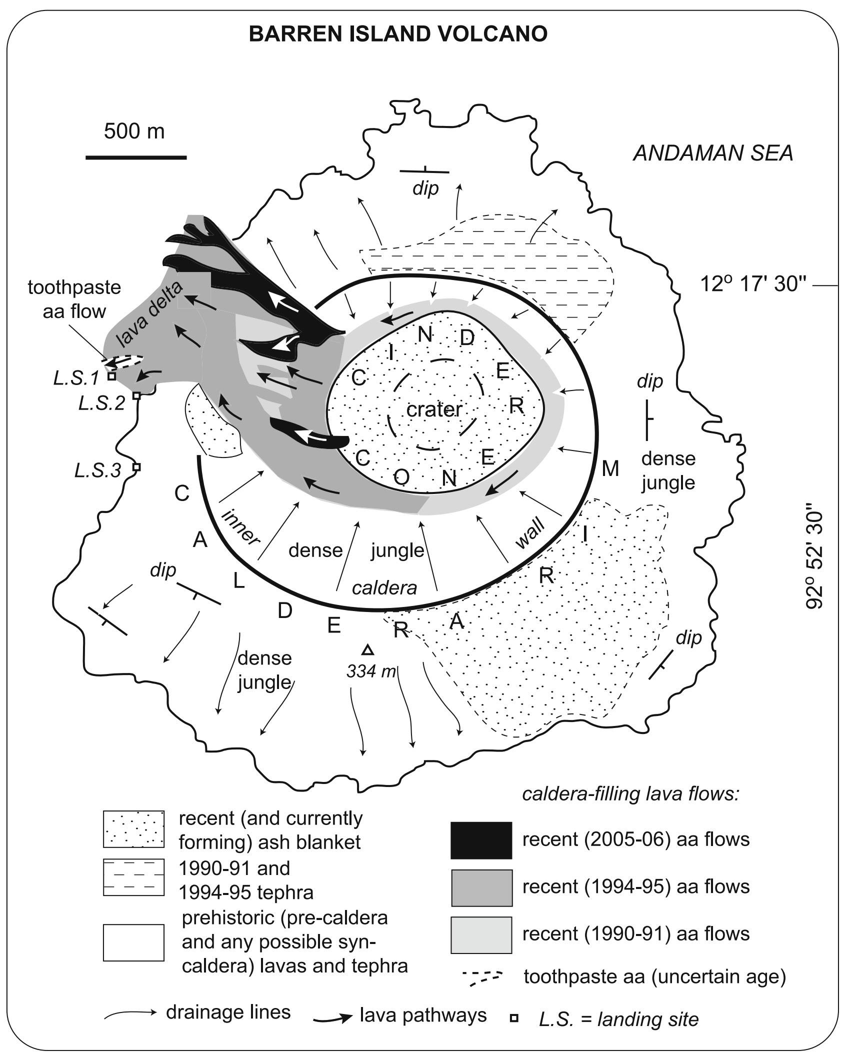


Fig. 2 **a**–**c** Panoramic views of Barren Island from the Indian Coast Guard vessels, showing the broad geology and morphology of the volcano. Photo **a** was taken April 2008, looking approximately NE, photo **b** taken April 2008, looking roughly ESE, and photo **c** taken January 2007, looking ESE. White patches on the cinder cone in **a** are fumarolic deposits. L. S. 1, 2 and 3 are the three Landing Sites, marked on map in Fig. 3. Their latitudes and longitudes given by G. P. S. are as follows: L. S. 1: N12° 17.08′/E93° 50.92′, L. S. 2: N12° 16.86′/E93° 50.95′, L. S. 3: N12° 16.24′/E93° 50.98′

Fig. 3 Geological map of the Barren Island volcano. Subsidiary vents on the central cinder cone are omitted for clarity. The authors landed on the island at the Landing sites 2 and 3 on the 2007 trip, and 1 and 2 on the 2008 trip. Site 1 (12° 17.078′ N, 93° 50.915′ E) has a small sheltered beach about 15 m wide and is the safest to approach; site 2 has a rocky shore and no beach; site 3 has a ~100 m long beach but approaching it is difficult because of breakers



Island rocks. Except for a 5-m-wide, NNE-SSW-trending basaltic dyke reported by Alam et al. (2004) as cutting the prehistoric lava flows on the southeastern inner caldera wall, no intrusions are known on the volcano.

Barren Island has had three recent eruptions in 1991, 1994–95, and 2005–06, during which it has erupted aa lava flows of basalt and basaltic andesite and pyroclastic materials (meaning the category of volcaniclastic materials that forms from volcanic eruption plumes or jets or pyroclastic density currents, White and Houghton 2006). Previous workers have used variable terminology for the various eruptions. Shanker et al. (2001) call the 1991 and 1994–95 eruptions, which they describe, as "recent", or the "third cycle", their second and first cycles referring to the historic (1787–1832) and the prehistoric eruptions respectively. Luhr and Haldar (2006) include the 1991 and 1994–95 eruptions among "historic" volcanism. Here we use the terms (recent, historic, prehistoric) following Shanker et al. (2001), and also introduce two new useful terms. Because the prehistoric deposits exposed on the caldera wall must predate the formation of the caldera itself, we call them the "pre-caldera" volcanic deposits. The historic (1787–1832) as well as the recent eruptions have occurred from the central cinder cone, well within the caldera. We therefore address both the historic and recent eruptive products as "caldera-filling" volcanic deposits.

A question is whether there are also "syn-caldera" deposits that accompanied the formation of the caldera, and which may be preserved between the pre-caldera deposits and the caldera-filling deposits. Shanker et al. (2001) speculated that the caldera may have formed in late Pleistocene time, by an original, 1,100-m-high cone blowing its roof off in a giant eruption. It is true that caldera-forming explosive eruptions are important in the evolution of not only dacitic and rhyolitic oceanic volcanoes, such as Kuwae in the Vanuatu (New Hebrides) arc (Robin et al. 1994a), but also of many mafic- and intermediate-composition island volcanoes, such as Tambora and Krakatau in Indonesia (Sigurdsson and Carey 1989; Self 1992), Mount Pelée on Martinique (Smith and Robol 1990), Tanna and Santa Maria in Vanuatu (Robin et al. 1994b, 1995; Allen 2005), and Aso in Japan (Miyabuchi et al. 2006). However, the formation of summit calderas on many basaltic stratovolcanoes and shield volcanoes is *not* associated with eruptions. Studies indicate that the formation of these calderas occurs by subsidence of a volcano's central floor along ring-shaped faults, the subsidence caused by rapid magma withdrawal (the "cauldron subsidence" concept predating the 1930's; see e.g., McBirney 1956, 1990; Macdonald 1965; Lockwood and Lipman 1987; Scandone 1990; Hirn et al. 1991; Robin et al. 1993; Miura 1999; Rymer et al. 1998; Geshi et al. 2002; Michon et al. 2007). There is no evidence for a high proto-Barren Island volcano blowing its roof off in a Krakatau-1883-like eruption. Shanker et al. (2001) also did not identify the deposits left on the volcano itself by this proposed event, and some that we describe below might correspond to such an event, or to older (pre-caldera) volcanism itself. We believe that the caldera of Barren Island is bounded by a ring fault, and note that there is a small tilted block on the western end, rising to about 100 m above sea level but without a clear relationship to the caldera wall, which shows the otherwise south-dipping prehistoric pile forming the southern half of the volcano to dip northwards, *towards* the caldera-filling lava flows (Fig. 2a, b). Cole et al. (2005) provide a recent review of calderas, and the Barren Island caldera can be considered a "simple, single-event, symmetric collapse, circular basaltic caldera" following their terminology. This, as the modelling of Roche et al. (2000) suggests, may indicate a shallow-level magma chamber. This would be consistent with the observations of Luhr and Haldar (2006) that several Barren Island lavas contain disaggregated troctolitic (olivine + plagioclase) cumulates from a shallow magma chamber under the volcano.

The caldera of Barren Island is small in comparison to some others on basaltic volcanic islands, such as the 12-km-wide caldera on Ambrym, Vanuatu (Carniel et al. 2003). Some others are of comparable size, such as the 1.5 km Rano Kau caldera, Easter Island (Baker et al. 1974). Tanna in Vanuatu has a 4 km wide caldera (Allen 2005). On many calderas the exposed caldera walls are considerably outward of the original ring fault, i.e., calderas once formed enlarge themselves by mass movements of their steepened walls. Shanker et al. (2001, their Fig. 1.28) described "rootless masses" of older lava flows in Barren Island's historic (1787–1832) aa lavas. These "rootless masses" may represent talus fallen from the inner caldera wall and rafted away by the lava flows. Or, they may be blocks of the central cinder cone which is said to have been born at the time rather than to have pre-existed (Luhr and Haldar 2006, following older accounts); such rafting of cinder cone blocks by lava flows is described from the 1973 eruption of Eldfell cinder cone on Heimaey, Iceland (Mattsson and Höskuldsson 2003), as well as the 1943–1952 eruptions of Paricutin, Mexico (Luhr and Simkin 1993; Valentine and Gregg 2008). We have not seen these "rootless masses", as access to the inner caldera wall from the solitary landing site, located within the breach in the caldera wall to the northwest, is now severely hampered by the extremely rugged surface of the recent aa flows, which also seem to have completely covered the historic flows.

A geological map of Barren Island, based on Luhr and Haldar (2006), monthly accounts, maps, and photographs on the Smithsonian Institution's Global Volcanism Program website (www.volcano.si.edu), and our own field work, is given in Fig. 3. A major difference in this map and the previous ones is the considerably enlarged area of recently erupted lava flows at the western coast of the volcano that is represented here.

Petrology and geochemistry

Figure 4 shows the total alkali-silica diagram of Le Bas et al. (1986) with published data for prehistoric through recent Barren Island rocks. The rocks are entirely subalkalic, and dominantly basalt and basaltic andesite, with only minor prehistoric andesite lavas. According to Luhr and Haldar (2006), all Barren Island rocks are porphyritic, with phenocrysts of both olivine and plagioclase, and usually also clinopyroxene. Hornblende constitutes a rare phenocryst phase in some prehistoric lavas. Groundmass is glassy to microcrystalline, with microphenocrysts and microlites of the same minerals, as well as titanomagnetite and ferrian ilmenite. Many lavas, particularly those from the 1994–95 eruption, contain small (2–5 mm) xenocrysts of extremely calcic plagioclase (up to $An_{95.7}$) and minor olivine (up to $Fo_{79.0}$) which strongly reduce bulk-rock SiO_2 contents while raising Al_2O_3 contents. Luhr and Haldar (2006) interpret these xenocrysts as products of disaggregation of troctolitic crystal mushes or cumulates located at shallow depth.

According to Luhr and Haldar (2006), Barren Island rocks show trace element signatures typical of subduction zone magmas, including large relative depletions in Nb and Ta and to a smaller extent in Ti, along with strong relative enrichments in K and Pb, and to a lesser extent in Sr. Isotopic ratios of Sr, Nd, and Pb measured by them on six samples show only a limited variation ($^{87}Sr/^{86}Sr$=0.70380 to 0.70405, ε_{Nd}=+4.1 to +6.8, and $^{206}Pb/^{204}Pb$=18.20 to 18.29). A prehistoric MgO-rich (8.2%) basalt has the most primitive isotopic values (lowest Sr and Pb, highest Nd) in the suite and the lowest abundances of the incompatible elements. Assuming a lava such as this to be parental to the rest, Luhr and Haldar (2006) inferred the following processes to have occurred: (1) melting of subduction-influenced mantle and subducted sediments, (2) selective scavenging of elements by magmas rising through the crust, (3) incorporation and disaggregation of troctolitic mushes, and (4) fractional crystallization of olivine (+ spinel inclusions), plagioclase, and clinopyroxene from the mantle-derived melts.

In being a mafic-dominated island arc stratovolcano Barren Island resembles the Quaternary volcano Tanna (basalt-basaltic andesite, with some andesite) in the Vanuatu arc (Allen 2005), or the Miyakejima volcano (basaltic andesite) in the Izu-Bonin arc (Geshi and Oikawa 2008). Barren Island contrasts with island volcanoes that erupt a fuller range of magmas, such as Anatahan in the Marianas arc (basalts through dacites, Wade et al. 2005), Easter Island (tholeiites, olivine tholeiites, various alkaline basalts, trachyte, and rhyolite, Baker et al. 1974), the Sumisu submarine caldera in the Izu-Bonin arc (basalt, basaltic andesite, dacite, and rhyolite, Shukuno et al. 2006), or Raoul volcano in the Kermadec arc (basalt, basaltic andesite, and dacite, Smith et al. 2006). If, in the absence of continental crust, crystal fractionation in magma chambers is a major mechanism by which mafic oceanic volcanoes generate highly silicic magmas such as dacite and rhyolite (e.g., Blake and Rogers 2005), then the lack of these magma types on Barren Island may be an indication that its magmatic system is of relatively young establishment. In comparison, the dormant volcano Narcondam has andesite and dacite and shows evidence for mafic and silicic magma mixing (Pal et al. 2007b).

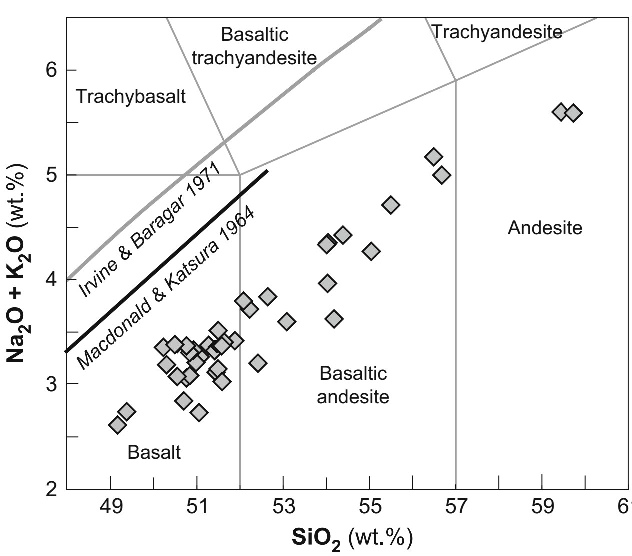


Fig. 4 Total alkali-silica (TAS) diagram showing the published data for prehistoric through recent volcanic rocks (45 samples) of Barren Island. Also shown are boundaries between the sub-alkalic and alkalic fields from Macdonald and Katsura (1964) and Irvine and Baragar (1971). Data sources are Alam et al. (2004), Luhr and Haldar (2006) and Pal et al. (2007a)

Prehistoric eruptions

Pre-caldera tuff breccias: deposits left by lahars and debris flows

The lowermost exposed prehistoric deposits on the inner caldera wall of Barren Island are polymodal deposits with sharply angular blocks of basalt as big as a metre, dispersed in a matrix of fine, ash-size, clay-like material (Fig. 5). These can be called tuff breccias. The best exposures are on the western shore of the volcano just south of the 1994–95 aa flows, where several of these deposits underlie a

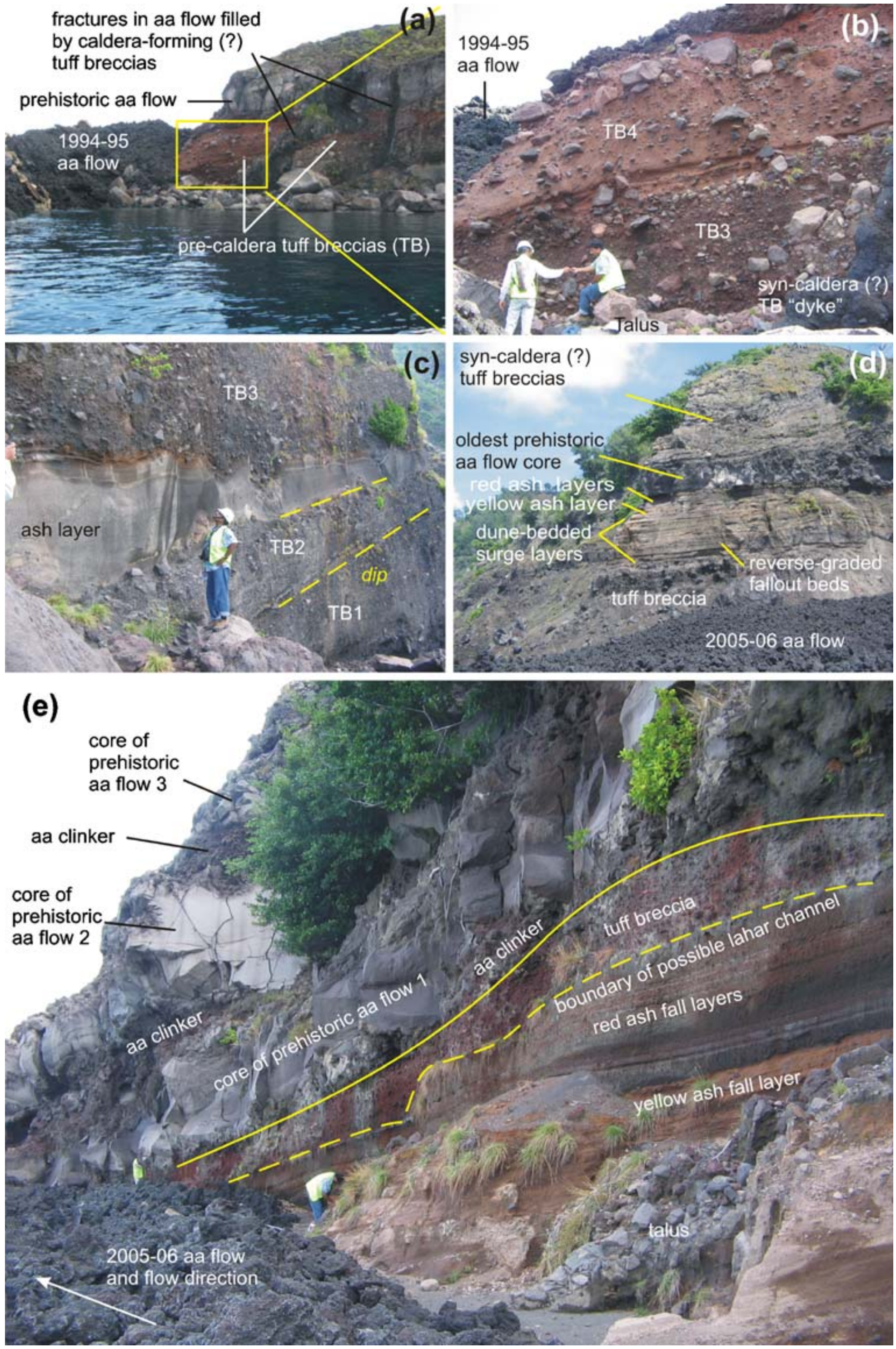


Fig. 5 **a**–**e** Prehistoric, pre-caldera tuff breccias (TB, interpreted as lahar and debris flow deposits and numbered TB1 through TB4 from oldest to youngest) and lava flows exposed on the inner caldera wall. Photos (**a**), (**b**) and (**c**) are of exposed sections on the tilt block, just south of the southern limit of the 1994–95 aa flows, at Landing Site 2, on the western shore of the island. Photos (**d**) and (**e**) are of sections exposed on the northern caldera wall and its northwestern end, just north of the northern limit of the 2005–06 aa flows

prehistoric aa flow (Fig. 5a). Two distinct tuff breccia deposits are visible in Fig. 5, and only a few metres southwards, four distinct tuff breccia deposits are exposed, with an unconsolidated, well-sorted ash bed in the middle (Fig. 5c), interpreted here as a pyroclastic fall deposit. The prehistoric lava flow and the tuff breccias below have steeply dipping fractures filled by still younger coarse fragmental material almost to the base of the exposed section (Fig. 5a). The aa flow may correspond with the lowest of the three aa flows exposed on the northern caldera wall, below which are seen ash and lapilli beds, followed by a tuff breccia deposit (presumably corresponding to the youngest on the southern side) (Fig. 5d; please see the interpretative logs and correlations in Fig. 6).

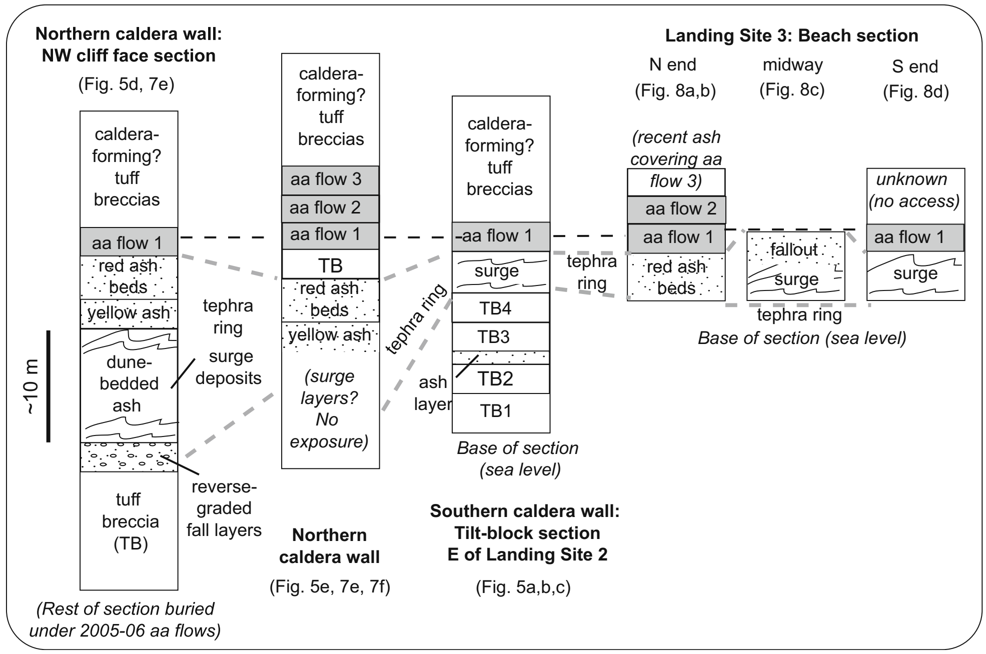


Fig. 6 Logs for the six locations or sections examined on Barren Island in the present study, with their inferred correlations. TB1 to TB4 are pre-caldera tuff breccia deposits, identified in the photographs in Figs. 5, 7 and 8

Shanker et al. (2001) have described these tuff breccia deposits as agglomeratic flows (e.g., their Fig. 1.9), but the total absence of bombs in them suggests that no molten rock was involved in the flows. Such tuff breccias may represent deposits of pyroclastic flows, particularly the sub-type of pyroclastic flows known as block and ash flows. Pyroclastic flows are common on arc volcanoes, with historic and recent examples of pyroclastic flows at Merapi, Indonesia (Bardintzeff 1984; Boudon et al. 1993), Unzen, Japan (Miyabuchi 1999), Mt. Pelée, Martinique (Fisher and Heiken 1982), Redoubt, Alaska (Gardner et al. 1994), and Soufriere Hills, Montserrat (Cole et al. 1998, 2002; Calder et al. 1999). Basaltic pyroclastic flows and their deposits are less common than those of more evolved composition, but are known, for example, from the large (25 km × 18 km) Aso caldera, Japan (Miyabuchi et al. 2006). Pyroclastic flows are typically associated with lava dome collapse, and move along a volcano's slopes, along channels, and the channel used or cut by such flows is often recognizable in cross-section (e.g., Németh and Cronin 2007).

Alternatively, the tuff breccias may be deposits left by debris avalanches or lahars, simply representing mass movements. Both are common and abundant on large oceanic volcanoes. Whereas debris avalanche deposits are typically hummocky, lahar deposits are channelized. The prehistoric tuff breccias on Barren Island appear to be channelized (Fig. 5e). We have not encountered carbonized plant remains that can help distinguish the deposits of the (high-temperature) pyroclastic flows from those of lahars. On the other hand, lahars, especially highly fluid hyper-concentrated flow varieties, contain substantial water. Considering the above, the tuff breccia deposit immediately underlying the prehistoric aa flow (Fig. 5b) is interpreted here as a lahar deposit with rainwater as the mobilizing agent. The two similar, but block-rich, matrix-poor deposits that underlie it are interpreted here as deposits left by debris flows. Accumulations of loose, coarse (~1 m), angular rubble that can be seen on the present slopes of the volcano, only some hundred metres to the south of these outcrops, support the interpretation that the prehistoric tuff breccias were formed essentially by mass movements.

Ash beds: pyroclastic surge and fall deposits

In the southern inner caldera wall section, in the tilted block described in the "Geology" section above, the lowest prehistoric aa flow and the lahar deposit beneath it are separated by ash and lapilli ash layers (Fig. 7a). They show mantle bedding (Fig. 7b) as well as dune bedding, sometimes with sags produced by angular basalt blocks (Fig. 7c, d). On the northern caldera wall, between the oldest prehistoric lava flow and the tuff breccia deposit are layers with reverse grading, followed upwards by ash and lapilli ash layers with dune bedding (Fig. 7e). They are overlain by well-bedded, well-sorted yellow and red ash layers (Fig. 7f). Shanker et al. (2001) report rare normal grading in these tephra at other locations. We interpret the ash and lapilli ash beds with reverse or normal grading, or which are well sorted (e.g., Fig. 5c), to represent pyroclastic fall deposits, and the ash beds with dune bedding to have been deposited by pyroclastic surges.

Surge-deposited beds overlain by fall deposits are seen also on the outermost southwestern side of the island, below the prehistoric aa flows (Fig. 8a–d). Therefore we

Fig. 7 a–**h** Pyroclastic surge and fall deposits preserved beneath the prehistoric aa flows and above the pre-caldera tuff breccias. Outcrops in (**a**) through (**d**) are located on the tilt block, just south of the recent lava flow field, at Landing Site 2. (**e**) A section on the northwestern end of the north caldera wall showing equivalent (presumably same) surge, fall, and lahar/debris flow deposits, as well as the oldest prehistoric aa flow. (**f**) Northern caldera wall a few tens of metres to the east of (**e**), showing the detail of the *red* and *yellow* ash fall layers

believe that these prehistoric, pre-caldera pyroclastic surge and fall deposits, exposed on both the northern and the southern caldera wall sections (sandwiched between the aa flows above and the tuff breccias below), as well as on the western shore of the island, represent a complete *tephra ring* that existed before the eruption of the prehistoric lavas. Tephra cones, rings, and maars are characteristic products of phreatomagmatism, the explosive interaction between magma and shallow surface water or groundwater (Hamilton and Myers 1963; Sheridan and Wohletz 1983; Sohn and Chough 1989; Zimanowki 1998; Thouret 1999; White and Houghton 2000). In contrast to tephra cones, dominated by fallout, tephra rings form when the bulk of the material has been deposited by pyroclastic surges, particularly base surges which propagate outward from a vent but close to the ground (Sohn and Chough 1989). Phreatomagmatism is an important eruption style of many eruptions on ocean islands (e.g., Fisher and Schmincke 1984; Cole et al. 1999). For example, the tuff cones and rings of the post-erosional Honolulu Volcanics, on Oahu, Hawaii, formed by dominantly phreatomagmatic activity on the slopes of the Koolau shield volcano, several million years after it had become inactive. These tuff cones and rings have incorporated blocks and fragments of shattered coral reefs that were growing around the sinking Koolau volcano by that time (Macdonald et al. 1983; Hazlett and Hyndman 1996).

Based on Shanker et al.'s (2001) observation, that "coral beds around the western shore (of Barren Island) were badly affected (by the 1991 eruption)", it is possible that coral reefs were in existence at the time the prehistoric pyroclastic surge and fall deposits formed on Barren Island. But the fact that these deposits contain no coral reef fragments, and few other accidental lithics (basalt blocks) suggests the absence, or at least non-involvement, of reefs, and certainly the dominance of juvenile magma and

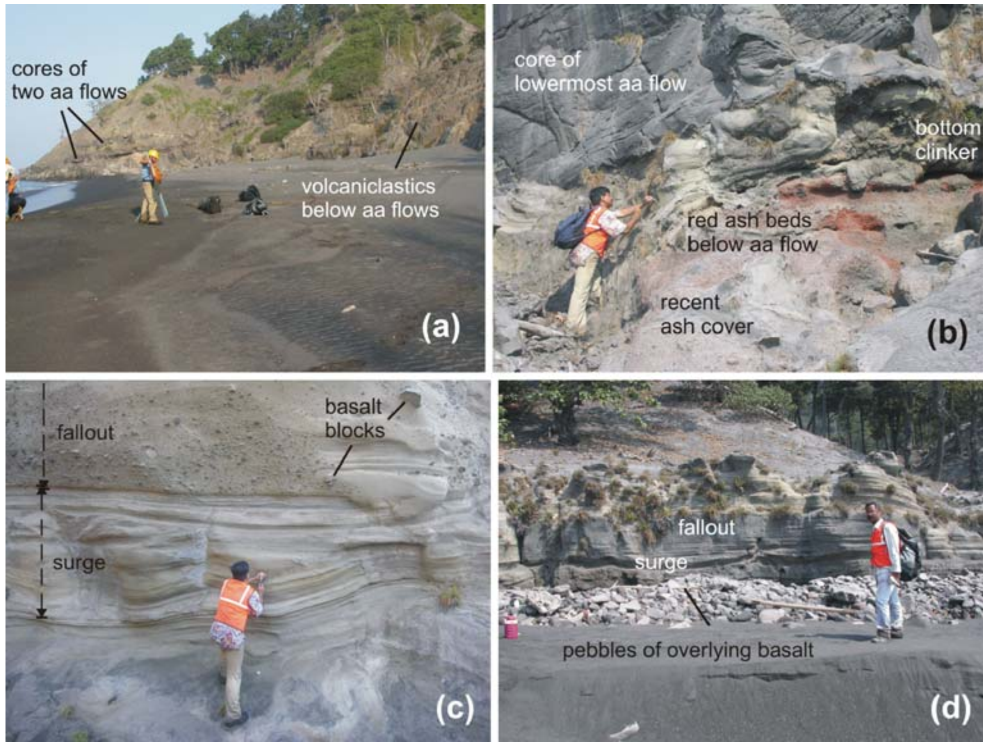


Fig. 8 The beach section at Landing Site 3 on the western side of the island, showing **a** two prehistoric aa flows underlain by tephra deposits, **b** close-up of the lowermost aa flow and underlying red ash beds, **c** fine, dune-bedded ash representing surge deposits overlain by airfall-deposited ash with a few lithic blocks of basalt, and **d** surge deposits underlying fallout ash beds. Frames (**a**) and (**b**) are from the northern end of the beach, frame (**c**) midway on the beach, and (**d**) from the southern end of the beach at Landing Site 3

minimal vent quarrying. In fact there is nothing to suggest that sea water was involved in magma fragmentation. The absence of accretionary lapilli throughout the outcrops examined suggests that the pyroclastic surges were more or less dry, with little external water involved (e.g., Wohletz and Sheridan 1983).

Shanker et al. (2001) have interpreted the cross-beds characteristic of the surge deposits on Barren Island as aeolian in origin. Whereas syn- or post-eruptive erosion by slumping, debris flows or wind does operate on tephra rings (Leys 1983; Sohn and Chough 1989; Chough and Sohn 1990), there are no palaeosols or erosional contacts within these deposits, suggesting rapid eruptions without breaks, and no horizons of coarse, angular rubble that might suggest reworking.

The prehistoric aa flows

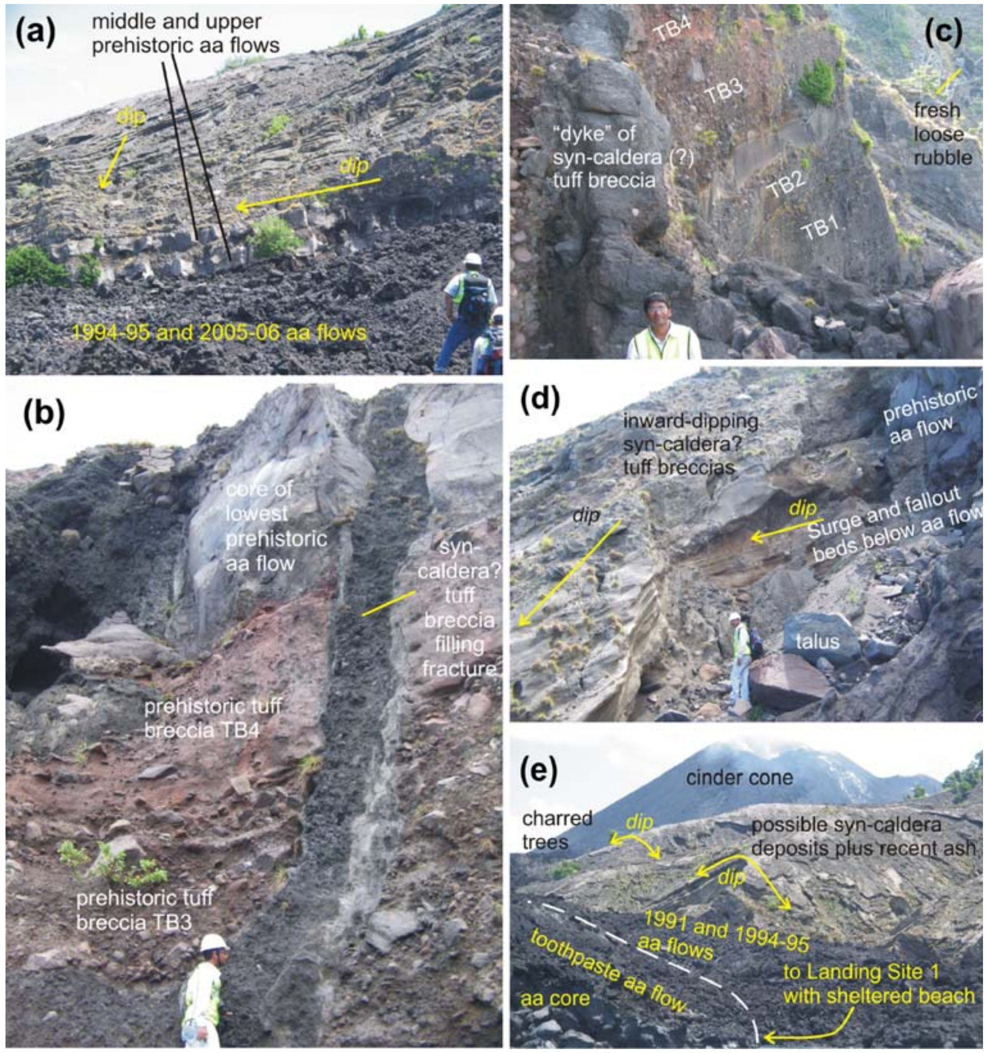
Each of the three aa flows on the northern caldera wall exhibits a distinct clinkery base and top, and a thick, massive, unjointed or poorly jointed core between the upper and lower clinker (Fig. 5e). Each clinkery horizon is about a metre thick, and the cores are about 4 m thick. The upper two lava flows appear to merge into one towards the northwestern end of the cliff face. Alam et al. (2004) report six lava flows on the prehistoric caldera wall along with scoria beds, and we wonder if their scoria beds are in fact the top and basal clinker like that accompanying these aa lavas.

Possible syn-caldera deposits

On Barren Island, it is unclear if there are deposits associated with the caldera-forming event. Shanker et al. (2001), who speculated about a giant caldera-forming prehistoric eruption, did not identify the associated deposits. Luhr and Haldar (2006) write that “Outcrop photographs in Shanker et al. (2001) show sequences of unconsolidated pyroclastic-surge deposits more than 15 m thick with prominent cross beds and bomb sags, likely from the caldera-forming eruption.” We believe that these photographs are of the outcrops we have described as pyroclastic fall and surge deposits *underlying* the prehistoric lava flows, which means that they were emplaced long before the caldera-forming event. We have also not observed bombs in these deposits (only blocks, and very few of them).

The youngest deposits exposed in the caldera wall are ~50 m thick pyroclastic deposits above the prehistoric aa flows. These include mainly well-stratified ash beds with some tuff breccias (Fig. 9a). Following many workers (e.g., Nakamura 1964; Decker and Christiansen 1984; Robin et al. 1993) we interpret these uppermost deposits as *possibly* representing the caldera-forming eruption. Some of the tuff breccias have entered the fractures in the prehistoric lava flows and underlying tuff breccias (Figs. 5a, 9b, c), and these fracture filling materials sometimes have a strong superficial resemblance to dykes (Fig. 9c). In places the ash layers and the tuff breccias dip inward towards the centre of the caldera (Fig. 9d) and show

Fig. 9 a–e Uppermost prehistoric deposits that may represent the caldera-forming eruption, overlying the prehistoric lava flows and tephra layers. Frame (a) is from the northern caldera wall, and (b) through (e) are from the tilt block just south of the recent aa flow field (Landing Site 2)



mantle bedding, suggesting that they were deposited on whatever topography existed at that time (Fig. 9e). Inward dips are common in such situations (e.g., Heiken 1971; Mattsson and Höskuldsson 2003).

Historic eruptions

Based on the accounts of Hobday and Mallet (1885), Ball (1888, 1893), Mallet (1895), Washington (1924), Raina (1987), Haldar et al. (1992a, b, 1999), Shanker et al. (2001), Haldar and Luhr (2003), Luhr and Haldar (2006), and reports from the Bulletin of the Global Volcanism Network (Smithsonian Institution, Venzke et al. 2002), Barren Island volcano had its first historically recorded eruptions in 1787, observed by passenger ships crossing the Andaman Sea. The earliest sketches of the island (sketches by Colebrooke and Captain Blair reproduced in Shanker et al. 2001) show the volcano in much its modern shape, including the breach in the caldera wall. However, the height of the central cinder cone has been shown to be a little less than that of the caldera wall, and, rather unrealistically, the cinder cone has been shown to have extremely steep (50°) slopes.

The activity in 1787 is said to have started with the formation of a new cinder cone near the centre of the caldera during a Strombolian-style eruption. "Strombolian" is not well defined in the volcanological literature but typically denotes intermittent bursts of incandescent material (see Valentine and Gregg 2008 and references therein). Activity continued until 1832 with breaks of 2–29 years between individual eruptions. The Smithsonian Institution lists the following eruptions during this interval: 1787, 1789, 1795, 1803–1804, and 1832 (Siebert and Simkin 2002). The cinder cone grew to 305 m height and developed a summit crater ~60 m across. The eruptions occurred from three subsidiary vents about 80 m below the crater rim on the northeast, west, and south flanks of the cinder cone (Raina 1987, and references therein). Basaltic aa lavas emerged from the three subsidiary vents, flooded the annular moat between the new cinder cone and the caldera wall, and ultimately flowed westward to cascade into the sea (e.g., see Fig. 3b of Luhr and Haldar 2006). Luhr and Haldar (2006) estimate a volume of ~25 million m^3 for them (but see below).

Cinder or scoria cones (and spatter cones) are typical of Hawaiian and Strombolian eruptions, the least violent (and entirely subaerial) forms of eruptions, which are typical of basaltic or basaltic-andesite compositions. Hawaiian eruptions (such as the ongoing eruptions of Pu'u' O'o' on Kilauea) are characterised by sustained jetting of magma into the air, whereas Strombolian eruptions involve discrete explosions (Valentine and Gregg 2008). The coarse ejecta and low eruption columns associated with both result in only local dispersal of the ejecta, which fall back and build spatter or scoria cones. Surtseyan and Taalian eruptions (phreatomagmatic eruptions building tephra cones and rings respectively) can be considered the phreatomagmatic equivalent of such eruptions (e.g., White and Houghton 2000; Schmincke 2004).

Recent eruptions

The recent (1991, 1994–95 and 2005–06) basalt and basaltic andesite flows have largely, apparently completely, covered and hidden the older historic (1787–1832) lava flows. Shanker et al. (2001), Haldar and Luhr (2003), and Luhr and Haldar (2006), as well as the Smithsonian Institution's Global Volcanism Program website (http://www.volcano.si.edu) offer valuable first-hand information on the recent eruptions. The 1991 flows erupted from the central cinder cone and flowed to the sea (Fig. 3), and the 1994–95 flows flowed along the southern margin of the 1991 lava field (Figs. 3, 10a) according to eyewitness accounts (Shanker et al. 2001), though they also appear to make up much of the lava front along the island's west coast. The 2005–06 flows travelled along the northern part of the lava field, close to the caldera wall, and reached the sea. The three eruptions have by now created a sizeable lava delta (term used by Luhr and Haldar 2006) at the western shore of the Island (Figs. 2a, b, 3, 10a). We describe these eruptions only briefly, as detailed descriptions can be found in the references cited above.

The 1991 eruption

This eruption began in late March 1991 from the existing cinder cone, producing thick jets of gas and red-hot lava fragments. The eruption began at the NE subsidiary vent from the 1787–1832 eruption, about 80 m below the rim of the crater of the cinder cone, and formed a new spatter cone. Lava flowed from that vent and also from the other two subsidiary vents of the historic eruption, and filled the annular moat between the central cone and the caldera by 6 April. Two new small, ~10-m-high lava driblet cones formed ~100 m and ~130 m west of the cinder cone atop these basaltic andesite lava flows, which were mostly blocky aa. These lava flows travelled westward to the sea where they buried a 12-m-high gas lighthouse on the shore and caused profuse boiling of the sea water and generation of thick steam clouds. The lava flows were individually 5–6 m thick, but by the late stages of the eruption they became ~25 m thick near the base of the cone and at the ocean entry.

Based on the mapped distribution of the 1991 lava flows, Luhr and Haldar (2006) estimate that they covered an area of ~0.26 km^2. Multiplying by an average lava thickness of 10 m over this area gives a total lava volume of 2.6 million m^3 (not 26 million as reported by them). If the ratio of tephra to lava was roughly 2:1, as estimated by eye (Haldar and Luhr 2003), then the associated 1991 tephra volume is 5.2 million m^3 (not 52 million as stated by them). Therefore, while we cannot account for the lava or tephra volumes that are under the sea, we consider that a good estimate of the total 1991 magma volume (lava plus tephra) that remains on the island is ~8 million m^3. Luhr and Haldar's (2006) figure of ~78 million m^3 is about an order of magnitude too large.

By early May 1991 the NE flank vent was ejecting incandescent fragments in pulsing pyroclastic columns (fire fountains), and this has been described as Strombolian activity. A pulsing fountain without pauses would probably be considered Hawaiian rather than Strombolian (e.g., Valentine and Gregg 2008). By August–September this activity advanced into sub-Plinian behavior as incandescent columns reached heights of about 1 km. Scoria and ash rained down upon the island, and a 3 m thickness of these is said to have covered the active lava flows. By late September all the three subsidiary vents on the cinder cone had broadened and merged with the central crater. The uppermost 80 m of the pre-1991 cone was removed and the summit crater, greatly broadened, was now 400 m wide, with the cone standing only 225 m high above sea level. Shanker et al. (2001) reported that the crater viewed from its rim resembled a giant funnel narrowing down to a conduit only about 25 m wide at a depth of 200 m, i.e., just 25 m above sea level. The eruption was continuing on 17 October, when the NOAA-11 satellite captured an image of a plume extending ~150 km WNW from Barren Island, but the eruption waned soon thereafter and was over by 31 October.

The 1994–95 eruption

This eruption started in mid-December 1994 and persisted into mid-June 1995. The event sequence for the first weeks of the 1994–1995 eruption is not known because the first

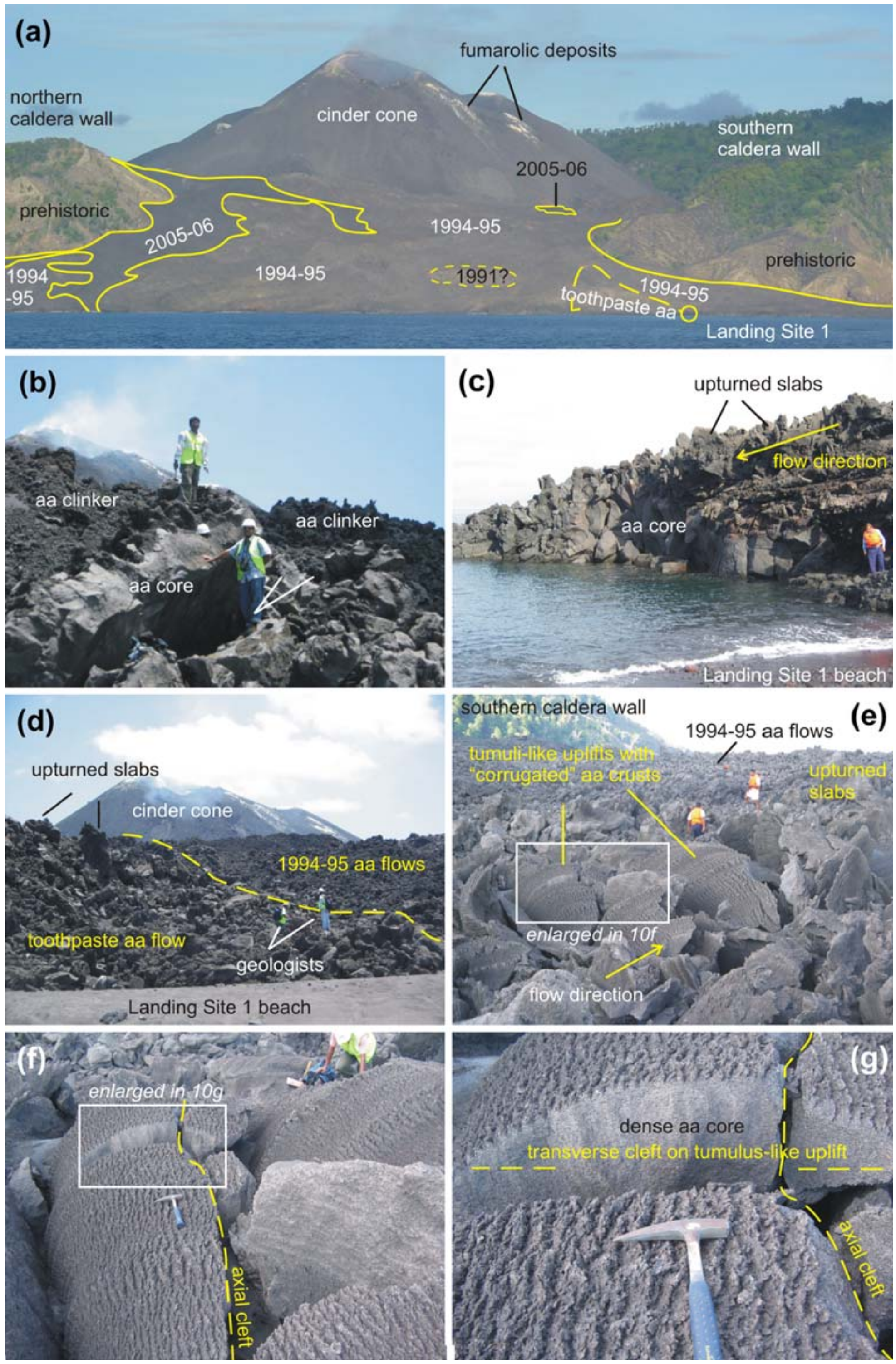


Fig. 10 a–g Recent aa flows. (**a**) is a panoramic view (looking east) of the recent lava flow field. *Dotted* boundaries are conjectural. Frame (**b**) is between the Landing Site 1 and the northern caldera wall. *Frames* (**c**–**g**) show the toothpaste aa flow at the Landing Site 1

GSI expedition did not arrive at Barren Island until 24 January 1995. Like the 1991 eruption, the 1994–1995 eruption has been termed mainly Strombolian in character and gained momentum over time to reach its paroxysmal stage during March 1995. It began with thick, dark smoke-like jets of ash and coarser clasts from a newly created flank vent on the cinder cone, 50 m below the 225-m-high crater rim. Three flank vents subsequently developed, which along with the central vent form a N-S-trending alignment. This alignment presumably reflects the orientation of the feeder dyke underneath. The only reported dyke on the island, the dyke cutting prehistoric lavas on the southeastern caldera wall (Alam et al. 2004), has a closely similar trend (NNE-SSW), perhaps suggesting that the regional tectonic stresses have remained much the same over the exposed history of the volcano.

The 1994–1995 lavas were basalt, and the eruptive activity did not change into sub-Plinian as it did for the 1991 eruption involving basaltic andesite. By late January, a spatter cone ~100 m high formed on one of the flank vents, and stood ~50 m higher than the summit crater rim of the central cinder cone. Block-lava flows fed from this vent travelled 1.5 km to reach the sea following a route between prehistoric lavas to the south and the 1991 lavas to the north, and caused profuse steaming at the ocean entry. The lava stream from a flank vent was about 50 m wide and 6 m high. Incandescent columns ~100 m high were produced from the vents, accompanied by vigorously pulsing plumes of gas, ash, and steam at every 30 s or so and rising to 800 m before being drawn by winds into a horizontal plume. A Space Shuttle image, taken on 14 March 1995 (Luhr and Haldar 2006; their Fig. 3d), shows the eruption plume drifting west towards the Andaman Islands. The eruption probably continued until the second week of June 1995, after which the volcano entered a strong but waning fumarolic stage.

Luhr and Haldar (2006) mapped the 1994–95 lava flows over an area of ~0.23 km^2. Assuming an average lava thickness of 10 m over this area gives a total lava volume of 2.3 million m^3, not 23 million as quoted. Based on the tephra to lava ratio of roughly 1:1, as estimated by eye (Haldar and Luhr 2003), the associated 1994–1995 tephra volume must also have a volume on the order of 2.3 million m^3. Therefore, the total erupted magma volume during the 1994–95 eruption (excluding the lava and tephra volume that entered the sea) is ~4.5 million m^3, an order of magnitude less than the ~46 million m^3 reported by Luhr and Haldar (2006).

The 2005–06 eruption

This is an extremely brief summary written based on the Smithsonian Institution's Global Volcanism Program website, where eight individual contributors are identified (http://www.volcano.si.edu). This eruption began in late May 2005 with an ash plume and fresh black basalt lava flows which did not reach the sea but produced a lot of steam from heavy rainfall. In June the eruption is stated to have become Strombolian, with periodic fire fountains ~100 m high and a dark plume rising 1 km (again, fire fountains, especially sustained ones, are typical of "Hawaiian" eruptions, e.g., Valentine and Gregg 2008, but there is no report of clastogenic lava flows fed by fire fountains). More lava erupted from the vent and flowed down the cinder cone. By September the lava flows reached the sea. Fire fountains from the cinder cone reached 300 m height. The eruption column's top formed a spectacular mushroom of gas and smoke, blowing to the north. Subsequent reports received from the Indian Coast Guard indicated that the eruption was continuous until at least 25 September. All active vents observed during the 2005 eruption lie in a zone trending almost N-S, an alignment also noted for active vents during the 1991 and 1994–95 eruptions. This activity continued through January 2006. Several earthquakes of moderate magnitudes (4–5) occurred in the region around Barren Island between November 2005 and January 2006. In January 2006 geologists observed dense clusters of incandescent pyroclasts of various sizes ejected forcefully from the crater with ballistic trajectories, presumably in discrete "Strombolian" bursts. By September 2006 the activity had slowed and become sporadic. Several ash plumes and red night glows over the crater have been observed since, well into July 2007.

The 2008 (and ongoing) eruption

The central cinder cone was sending periodic plumes of dark ash during our first field trip in January 2007 (Fig. 2c), but on our second trip in April 2008 it was in a rather quiet, fumarolic stage (Figs. 2a, 10a). Red glows over the cone at night, as well as ash plumes rising up to 2.5 km height, are reported between May and November 2008 (http://www.volcano.si.edu). A team of GSI's geologists visited the Island on 7th January 2008, and landed apparently at the same site as our Landing Site 1. It reports the formation of a new cinder cone south of the existing cone, and both cones sending "Strombolian" tephra columns upward in pulsating fashion every 10–60 s. The photographs in their report (http://www.gsi.gov.in/news.htm) suggest instead the development of a parasitic vent southwest of the crater of the existing cinder cone. A new (2008 and ongoing) eruption of the volcano can therefore be said to have begun.

The nature of the caldera-filling aa flows

Aa and pahoehoe are the two fundamental types of basaltic lava flows (e.g., Macdonald 1953; Peterson and Tilling 1980; Rowland and Walker 1990). All Barren Island lava flows, prehistoric through recent, are aa, including blocky aa. There is no pahoehoe on Barren Island, arguably due to (i) somewhat lower eruption temperatures of the lavas, consistent with melt inclusion studies and water-present melting in arcs (Luhr and Haldar 2006), and (ii) high strain rates experienced by the flowing lavas due to the steep ground slopes.

It is difficult in the field to distinguish between the aa flows issued in 1991, 1994–95 and 2005–06. The flows are distinctly channelized, as is typical of aa flows worldwide, and the whole aa flow field is made up of ridges of the aa flows sloping towards the sea but with an overall surface amplitude of as much as 25 m. From a distance the aa lava streams can be distinguished by subtle colour differences (shades of grey through black, with darker shades for younger lavas), and the younger lava streams can be seen to have left some “islands” of the older lavas between them (Fig. 10a). All these aa flows have jagged, very sharp and highly vesicular clinker at the top (Fig. 10b), based on which they can be characterized as proximal aa (terminology of Rowland and Walker 1987). They also show massive cores at several places (Fig. 10b, c), as do all Barren prehistoric aa flows (Fig. 5e), and aa flows do in general. Rowland and Walker (1990) describe how massive cores of aa flows can climb up from below the surface clinker along ramp structures, once the aa flow front has come to a halt and aa behind the front continues to flow forward. This is particularly seen in what Rowland and Walker (1990) term distal aa. One such ramp structure is shown in Fig. 10b.

The aa flows resemble aa flows elsewhere (e.g., the aa flows of the Stromboli 2002–03 eruption, Lodato et al. 2007; aa flows of Neá Kaméni in the Santorini caldera, Greece, or those filling the Valle del Bove caldera on Etna, Sicily, as illustrated in Scarth and Tanguy 2001; or Mauna Ulu 1969) and have behaved like them. For example, the 1991 lavas buried a 12-m-high lighthouse on the western coast that was the only man-made structure on the island. This is reminiscent of the hawaiite-mugearite blocky aa flows (Jakobsson et al. 1973) from the 1973 Eldfell cinder cone on Heimaey (Vestmannaeyjar, Iceland) that buried houses (Mattsson and Höskuldsson 2003), or the famous aa flow from the newborn 1943 cinder cone of Paricutin (Mexico) that half-buried a church in the nearby village (Luhr and Simkin 1993).

An intriguing morphological type of aa makes up a small part of the aa flow field near Landing Site 1 on Barren Island (Figs. 3, 10c–g). Its surface crust has been extensively broken into plates or slabs throughout, whereas it shows well-developed aa cores underneath (Figs. 9e, 10c). Evidently, the broken slabs of the surface crust—many of them razor-sharp—were carried along atop largely molten, mobile aa lava (which later solidified as the cores) and in the process experienced rotations, overturning, and even collisions.

An intriguing feature of this aa flow is that, locally, its surface crust is distinctly bent into convex upward shapes, resembling the tumuli common in pahoehoe flows which form by localized inflation (e.g., Anderson et al. 1999; Fig. 10e–g). Tumuli are most common in tube-fed pahoehoe flows (e.g., Walker 1991; Rossi and Gudmundsson 1996; Mattsson and Höskuldsson 2005; Sheth 2006), though they are also known from aa flows (e.g., Duncan et al. 2004; Lodato et al. 2007). These tumuli-like uplifts, elliptical in plan, and with long axes of 5–10 m, are approximately tranverse to the overall flow direction. They also exhibit well developed, smooth, axial and axis-perpendicular clefts (Fig. 10e–g) usually associated with true tumuli. By analogy, these clefts should indicate periodic crack propagation from the surface downward into viscous lava, consistent with tensional cracking of bulging, uplifting crust (e.g., Anderson et al. 1999).

An interesting aspect of the convex tumuli-like features is the long, linear, closely-spaced wrinkles or “corrugations” that run across the surface, parallel to their long axes and axial clefts (Fig. 10f, g). Shanker et al. (2001), without giving location information, have illustrated very similar, possibly the same, outcrops, judging from their photographic figures 1.23 (showing upturned slabs) and 1.24 (showing slabs with “corrugated” surfaces). They described the latter as “pahoehoe with ropy structure”, but these lavas are neither ropy nor pahoehoe. Note that the strong visual impression of continuous grooves and ridges imparted from a distance (Fig. 10f) is illusory; when traced these terminate and splay laterally, and are actually arranged en echelon (Fig. 10g).

Guest and Stofan (2005) have described comparable examples from the 1983 eruption of Mount Etna. These Etna flows have continuous, level slabs of pahoehoe crust normally 1–2 m wide and tens of metres long, and they typically erupt from ephemeral boccas which develop at a late stage of development of an aa flow field. They call these flows “slab-crusted” flows, and note their similarities with the “toothpaste aa” flows described from Hawaii by Rowland and Walker (1987). Toothpaste aa (also called spiny pahoehoe) typically issues from boccas within aa flows. The linear grooves and ridges of spiny pahoehoe are parallel to the local flow direction, and are believed to reflect the roughness of the boccas’ cross-sections. Tumuli are not unexpected in toothpaste aa; in fact, Walker (1991) writes that some of the best examples of tumuli are found in toothpaste aa. Rowland and Walker (1987) also illustrate imbricate stacking of surface plates or slabs on a toothpaste aa flow tongue (their Fig. 8), within a broader aa flow field, much as in the Barren Island flow.

We therefore consider this flow to represent a form of toothpaste lava, with which it shares many characteristics (Rowland and Walker 1987; Guest and Stofan 2005). This flow forms a part of the present-day western coastline as well as the edge of the lava delta. Its location is significant, as it provides indirect information about its likely age, a matter of some confusion in the literature. Shanker et al. (2001, Fig. 1.24) ascribed this toothpaste aa flow (whose corrugated surfaces they described as “ropy pahoehoe”) to the historic (1787–1832) eruptions. But if this is indeed historic, how does it form the edge of the lava delta which is overwhelmingly made up of the recent (1991, 1994–95 and 2005–06) flows? Note that this toothpaste aa flow cannot represent a long and isolated promontory of an historic aa flow, jutting out into the sea much more west than other historic flows and which the recent lava flows might have failed to cover. No such promontory is visible in the photograph of March 1990 (Fig. 3b in Luhr and Haldar 2006) which shows the extent of the historic flows. In fact we find no grounds for allocating this flow to the historic eruptions (~150–250 years old), and also note its freshness and lack of alteration despite the moist, tropical climate, and, most important, the absolutely total absence of vegetation on it. In comparable settings, such as Hawaii, new vegetation begins growing on lava flows only a few years after their emplacement (Macdonald et al. 1983).

The confusion is further accentuated by a January 2009 report by a GSI geologists' team, referred to earlier. This report (http://www.gsi.gov.in/news.htm) contains a photograph (their Fig. 4) of the lava flow on the northern side of their landing site (our Landing Site 1). This is exactly the toothpaste aa flow we have illustrated in our Fig. 10c, and the said report considers this flow as having erupted between July 2005 and March 2006. But this toothpaste aa flow cannot be a 2005–06 flow, because photographs of it appear in Shanker et al. (2001), and eyewitness accounts have described the 2005–06 flows as having flowed mainly along the far northern edge of the recent flow field.

Therefore the most likely date for this toothpaste aa flow is either 1991 or 1994–95. Noting that the 1994–95 aa flows cover a major portion of the recent aa flow field (Fig. 3, 10a), the latter is the more probable of the two.

The cinder cone and recent ash cover

The central cinder cone on Barren Island has existed for the past 220 years at least, and must be described as polygenetic, in contrast to the much more common monogenetic cinder cones that are typically active for a few months to a few years (Paricutin 1943–1952; Luhr and Simkin 1993), and which form fields of up to hundreds of individual cones worldwide, both in continental and oceanic areas (e.g., Baker et al. 1974; White 1991; Carn 2000; Walker 2000; Németh et al. 2003; Valentine and Gregg 2008). The interplay of primary eruptive processes and erosion in the development of cinder cones has been addressed by several workers (Ollier 1959; Ollier and Brown 1971; Wood 1980a, b; Hooper and Sheridan 1998; Németh 2004). In the time between the historic eruptions, which ended in 1832, and the first of the recent eruptions, in 1991, Barren Island's cinder cone managed to survive erosion, testifying to the general rule that cinder cones are well sorted and highly permeable, which means slow erosion because of little surface runoff (Segerstrom 1950; White et al. 1997). This cinder cone lost half its original height during the 1991 eruption (225 m above sea level at the end of the eruption), but grew higher and steeper again during the 1994–95 and 2005–06 eruptions, so it now rises well over 400 m above the sea. It is already active in a new eruption—so far only of tephra—that began in 2008. If activity persists, new lava flows are not unexpected.

Fine ash from the 1991 eruption, after the associated aa flows were erupted, reportedly covered the flows, and the rest of the volcano around the cinder cone was under ash up to 3 m thick. This was quickly removed by rain from over the 1991 flow as reported by Shanker et al. (2001). The possibly syn-caldera ash beds as well as the ash blanket deposited during the historic and recent eruptions

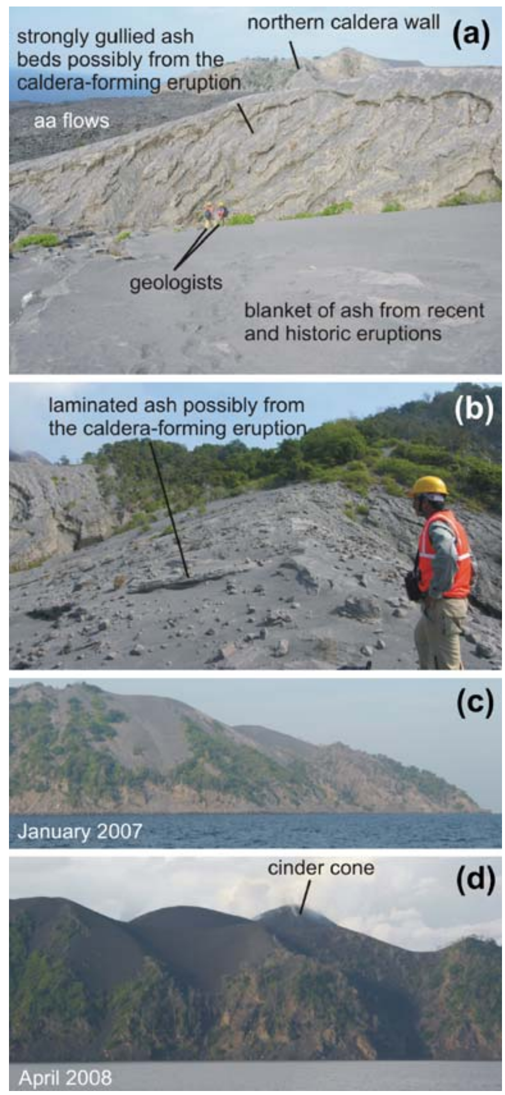


Fig. 11 **a**, **b** Recent ash blanket covering the tilt block on the south of the recent lava field. Well-stratified ash beds from the caldera-forming eruption show severe gullying (**a**), and are being exhumed from the most recent ash deposits (**b**). Recent (1990's) ash blanket on the southeastern side of the volcano, as photographed during our two trips, is shown in (**c**) and (**d**)

are now deeply gullied (Figs. 9e, 11a, b). Fall of fine ash can also cause gullying of older, previously stable, landscapes (Segerstrom 1950, 1966). However, the recent eruptions have also deposited a large amount of dark grey ash and cinders on the southern and southeastern side that when incandescent burned part of the thick forest (Fig. 11c, d).

Summary

Barren Island, a young and growing, mafic, island arc volcano in the Andaman Sea has been known since 1787, when its first historically recorded eruption began, continuing until 1832. Three recent eruptions in the past two decades (1991, 1994–95, 2005–06) have produced aa and blocky aa lava flows of basalt and basaltic andesite, along with tephra, and a newer eruption—so far only of tephra—has begun in 2008. The prehistoric activity of the volcano is represented subaerially by lava flows of basalt and basaltic andesite (and rare andesite) and volcaniclastic deposits (including deposits left by lahars, and pyroclastic fall and surge deposits).

Understanding current and past eruptive styles of active volcanoes—and particularly those as little known as Barren Island—is important because it helps us with clues to the future eruptive activity of the volcano as well as potential volcanic hazards. The volcano is only some 500 km from the Myanmar coast, 135 km from Port Blair, and a mere 70 km from the nearest inhabitated island of the Andamans. Barren Island's pre-caldera volcanism occurred well before its historic records, perhaps in the earliest Holocene, late Pleistocene, or even earlier (Luhr and Haldar 2006), and three recent eruptions in rapid succession and a fourth ongoing one suggest considerable excitement in store for the volcanological community.

Whereas we have provided here the first volcanological synthesis of Barren Island volcano, we acknowledge that there is ample scope for detailed work on the volcano and for refining the broad picture presented here, when current accessibility and logistic issues become less daunting. Besides geochemical and isotopic work on Barren Island volcanics that we have currently in progress, future endeavours for research in this region should include (i) more-continuous monitoring of the volcano by government agencies than hitherto carried out, (ii) dredging and possibly drilling its undersea bulk, (iii) precise age-dating of its eruptive products with appropriate radioisotopic techniques (such as U-Th and Ar-Ar), and (iv) similar studies of Narcondam. The combined results of these endeavours will be important for understanding the volcanic and tectonic evolution of the Andaman subduction-accretion zone.

Note added in proof: We would like to report a confirmed, ongoing, lava eruption on Barren Island. The first four authors of this paper and Neeraj Awasthi visited Barren Island yet again, on 30th March 2009, on board the Indian Coast Guard vessel the *ICGS Bhikaiji Cama* (Captain: Commandant M. Bhatia). The volcano's central cinder cone was continuously emitting dark ash clouds every few seconds from its central crater (reminiscent of the activity in 2007 illustrated in Fig. 2c of this paper), and these clouds were expanding and getting deflected towards the south. The pre-existing valley between the cinder cone and the northern caldera wall has been filled up by deposition of new ash in the past year, which has enabled the new, active lava flow to completely abandon the westerly route (taken by all historic and recent lava flows) and to reach the sea over the northern caldera wall. This lava flow is not ensuing from the summit crater, but apparently from an intermediate elevation on the cinder cone, though details were hard to distinguish given the distance of the ship from the island. The new, channelized lava flow is currently descending at a steep angle over the northern caldera wall's outer cliff face, and into the sea. Incandescent lava is seen at a few places in it, particularly in the dark. A sizeable steam plume is currently rising from the sea where this new lava flow is entering the sea. The new lava flow has built a structure resembling an alluvial fan along the shore. We were able to reach this fan by using a Gemini (inflatable rubber boat) from the ship, carefully circumventing the steam plume and through seawater which was very hot (an estimated ~60–70°C). We could also collect lava samples from the southern edge of this "fan", which are typical clinkery ~~as~~ basalt in hand specimen. A full account of this eruption with photographs and petrological study of the rock will be attempted separately. An interesting possibility is that, were this activity to continue, the new embryonic lava delta will grow thicker and laterally, and merge with the existing lava delta on the western side of the volcano shown in Fig. 3. In summary, Barren Island is evidently a *very* active volcano, meriting close study.

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