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Maar-diatreme volcanoes: a review

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Abstract
Maar-diatreme volcanoes are produced by explosive eruptions that cut deeply into the country rock. A maar is the crater cut into the ground and surrounded by an ejecta ring, while the diatreme structure continues downward and encloses diatreme and root zone deposits. Here we attempt an evenhanded review of maar-diatreme volcanology that extends from mafic to kimberlitic varieties, and from historical maar eruptions to deeply eroded or mined diatreme structures. We conclude that maar-diatreme eruptions are episodic. Ejecta rings provide invaluable insight into eruption processes and sequence, but are incomplete records of diatreme formation. Deposits within the diatreme structure include, in varying proportions, lower unbedded deposits sometimes typified by subvertical contacts among domains of debris emplaced sequentially, and upper bedded deposits formed by sedimentation on surfaces open to the atmosphere. A basal root zone comprises the transition from coherent magmatic feeder dike to clastic deposits formed by fragmentation of magma and enclosing country rock; root zones are irregular in form, and the clastic deposits are typically intruded by contorted dikes. Irregular root zone-like chaotic breccias cut by contorted dikes are also present within diatreme deposits, where they represent intra-diatreme fragmentation zones and record changes in the location of the explosion locus during eruption.

1. Introduction

The number of studies of maar-diatreme volcanoes has been increasing rapidly in recent years. Two of the major drivers for this increase are the discovery and exploration of new kimberlite diatremes in Canada (Kjarsgaard, 2007), and recognition that many maar-diatreme structures have hosted long-lived and climate-sensitive lakes (e.g., Zolitschka et al., 2000). They are also research targets in studies of monogenetic volcanic fields, including those undergoing assessments of volcanic risk (Lorenz, 2007). Identification of maar-diatreme deposits precursory to flood-basalt eruptions has added larger-volume eruptions to the family, as well as information from superb Antarctic exposure at Coombs Hills and Allan Hills, sites which are part of the Jurassic Ferrar Large Igneous Province (White and McClintock, 2001; McClintock and White, 2006; Ross and White, 2006; Ross et al., 2008a).

Canadian kimberlites have proven, since their discovery in the late 1980's, to display a number of features differing from those of long-studied Southern African counterparts. A key feature at Fort à la Corne, for instance, is that many diatremes are small (e.g. Lefebvre and Kurszlaukis, 2008; Pittari et al., 2008), and are associated with significant volumes of volcaniclastic kimberlite that is not in a diatreme, and instead comprises remnants of tuff cones or tuff rings. The surficial deposits are intercalated with sedimentary deposits that provide evidence for repeated subaerial to submarine eruption over millions of years (e.g., Leckie et al., 1997; Zonneveld et al., 2004; Kjarsgaard et al., 2006, 2009a).

An important result of such findings was the rejuvenation of volcanological research into kimberlite eruptions. The long-established maar-diatreme model formulated and refined by Volker Lorenz has gained new nuances in explaining how diatremes can be excavated by a suite of processes driven by energy from magma-water interactions. Fluidization models initiated in the 1940's and 1950's (Cloos, 1941; Reynolds, 1954) then strengthened with analog modeling in the 1970's (McCallum et al., 1975; Woolsey et al., 1975) have been refined and adapted (Sparks et al., 2006; Walters et al., 2006). Accounts of kimberlite-specific emplacement styles have also continued to develop (e.g., Skinner and Marsh, 2004; Skinner, 2008; Scott Smith, 2008; Mitchell et al., 2009). In light of this diversity of approach and interpretation, several workers including the authors of this review are abandoning the habit of examining diatremes, including kimberlitic examples, with the view of simply choosing between a magmatic or phreatomagmatic interpretation. It is clear that not all kimberlite occurs as maar-diatreme volcanoes, just as not all small basaltic volcanoes are scoria cones.

In this current review of maar-diatreme volcanism, our attention is first focused on observed eruptions, then on surface features and deposits of maars, then progressively downward through diatremes to the root zones and dikes that feed them. A useful exhaustive review of early maar-diatreme literature by Lorenz et al. (1970) that covers a great deal of descriptive information in a similar order, is now available online at the NASA Technical Reports server. The rationale for our approach is that at the surface, there are observable features directly related to eruption energy and, particularly, eruption progression through time. There have also been historical observations of maar-forming eruptions, which provide additional insight into the processes that form them. Table 1 summarizes information that supports our approach of looking at kimberlite diatremes together with those formed by other magma types. Note that very few kimberlite diatreme sites preserve the syn-eruption surface.

1.1 Terminology

Many and conflicting terms have been used in description of diatremes, maars, and their deposits, with kimberlite terminology particularly contentious (contrast Cas et al., 2008a with Scott Smith et al., 2010). Table 2 and figure 1 present the terms we will use in descriptions and discussions of maar-diatreme eruptions and volcanoes in this paper. An attempt is made to distinguish clearly between transient, syn-eruptive features, formed and changed during an eruption, and features of the resulting deposit preserved as surface landforms and the infilled structure comprising the overall diatreme and in some cases, filled post-eruptive crater. The down-structure path of the paper follows headings indicating different parts of the overall maar-diatreme volcano and these add detail to the summary comments in the table. We do not use the term “facies”, as in “crater facies” or “diatreme facies”, because in current use they are not lithofacies, but a mixture of rock information and structural setting linked by an assumed common genesis. We do not use “pipe”, preferring "diatreme structure" to refer to the whole of the excavation into country rock that contains deposits of the root zone, diatreme fill, and any post-eruptive sediment.

1.2 Small volcanoes

Volcano names indicate morphology; stratovolcanoes are tall cones, calderas are big subsidence features, and shield volcanoes have a broad, low profile. Small volcanoes, whether cones, rings or maars, are often called "monogenetic", but many have been revealed by closer study to have had long histories of multiple eruption (Leckie et al., 1997; Freda et al., 2006; Harvey et al., 2009). For this reason we prefer the descriptive and encompassing term "small" for these volcanoes, while recognizing that most will also be the products of a single eruption.

Maar volcanoes are distinguished from other small volcanoes in having craters with their floor lying below the pre-eruptive surface (Ollier, 1967; Lorenz, 1973, Fisher and Schmincke, 1984; Table 3). Older papers included as maars phreatic explosion pits, which produced no juvenile ejecta (e.g. Ollier, 1967; table in Kienle et al., 1980); we follow Lorenz (1973) in restricting maar volcanoes to those formed by eruptions of magma, though juvenile fragments may form only a small proportion of total ejecta. Hydrothermal explosion pits have different physical origins (Muffler et al., 1971; Nairn and Wiradiradga, 1980; Nogami et al., 2000; Browne and Lawless, 2001), and are not discussed here. Maar volcanoes have formed in a great diversity of subaerial volcanic settings, from magmas with a wide range of compositions (Lorenz, 2007). Most are surrounded by a relatively thin ejecta ring extending around a kilometer from the crater rim, unless this feature has been eroded (for older volcanoes), although unusually thick ejecta piles are also possible for maars, as mentioned below. Tuff rings comprise similarly broad and relatively thin deposits of ejecta, but their craters do not cut below the original ground surface (Lorenz et al., 1970; Head et al., 1981; Wohletz and Sheridan, 1983; Table 3). Tuff cones and scoria cones, as the names imply, are volcanoes that have cone-like shapes and thick near-vent deposits that thin very rapidly outward (Head et al., 1981; Table 3). Various crater width vs. deposit height ratios have been used to categorize tuff ring vs. tuff cone or scoria cone deposits (e.g. Wood, 1979; Head et al., 1981), but the key difference between maars and other monogenetic edifices is the level of the crater floor. The morphology of the edifice beyond the central crater is controlled by depositional processes (Sohn, 1989), whereas the crater floor depth is controlled by excavation down the conduit to remove country rock, counterbalanced by the addition of new juvenile material. Because the controls are different, it is possible to have "tuff cones" that are also "maars" (Aranda-Gomez and Luhr, 1996), although this is unusual. Moore (1985; 1987) interprets Surtsey as a tuff cone built over an initial maar, i.e. underlain by a diatreme (but see Kokelaar, 1987).

It is also worth emphasizing that this classification is geomorphological. An eruption, following Fisher and Schmincke (1984), is a period of more or less persistent delivery of magmatic material to the earth's surface through one or a number of vents linked to the same magma source. For prehistoric volcanoes, it is commonly impossible to distinguish between products of a single eruption and an eruption episode comprising a series of eruptions, separated by distinct breaks in activity, over a short time, but geologically the distinction is not important. A single eruption or episode that initially produces a maar crater, but which then changes to a different type of eruption that fills and buries the maar crater produces a volcano that is named based on its final morphology, for example, a scoria cone (e.g., Lorenz, 1986; White, 1991a).

2. Historical maar eruptions

In 1977 the Ukinrek Maars formed in Alaska, and have become the most commonly cited example of an historic maar eruption (Kienle et al., 1980; Self et al., 1980; Büchel and Lorenz, 1993). An earlier eruption at Nilahue (Chile), in 1956, was also partially observed, and the still-steaming crater examined soon after the eruption (Muller and Veyl, 1956). The Ukinrek and Nilahue eruptions have been often commented on and will not be discussed further here.

Two other well-documented eruptions produced maar craters, but are not often cited in this context: the Rotomahana eruption of 1886, part of New Zealand's Tarawera eruption (Nairn, 1979; Rosseel et al., 2006), and the 1965 eruption of Volcano Island in Lake Taal (Philippines) (see below). The quality and extent of records for these four eruptions vary (Table 4), but shared features are common, with some features shared by all.

2.1 Taal 1965: a maar-forming eruption

The phreatomagmatic eruption of Taal in 1965 left behind a crater excavated below the pre-eruptive ground surface, surrounded by deposits of base surges (rich in non-juvenile fragments), so it should be recognized as a maar-forming eruption. The following summary is based entirely on the account of Moore et al. (1966), with additional interpretations from Moore (1967), Kokelaar (1986) and White and Houghton (2000).

Description. The September 1965 eruption of Volcano Island in Lake Taal lasted about 2.5 days. After a brief opening phase of magmatic-style explosive activity initiated at a vent about 1.2 km inland, the eruption became phreatomagmatic, for reasons discussed below. The most intense phase of the activity lasted about six hours during the first half-day and formed an elongate crater (NE-SW), now occupied by an arm of Lake Taal. During this phase, explosion sites migrated back and forward along a line parallel to the crater elongation, although "the most intense activity was near the NE end of the crater". Explosion columns reportedly reached 15-20 km in height, suggesting that subplinian intensities were attained, while at the same time base surges emanated from the column. The surges were "wet" (<100°C), as evidenced by thick mud coatings on trees and houses, and other observations. The coalesced phreatomagmatic crater is about 1.5 km long, 0.3 km wide and up to 200 m deep (cliff height above the water level, plus maximum water depth in the inlet), for an estimated crater volume of 0.04 km^2. The volume of ejecta is estimated at 0.09 km^2 and it comprises a mixture of "shattered and pulverized old lava, ash, and lake sediments" with juvenile material.
During the last 8 hours of the eruption, a "cinder cone" was built at the NE end of the phreatomagmatic crater. This edifice was 250 m-wide at the lake level, and rose 18.5 m above it. The base of the "cinder cone" was below the post-eruptive water level, and the rather large crater of the now-breached edifice is also occupied by water. On recent photographs (Fig. 2) the "cinder cone" looks more like a tuff cone, and the eruptions that formed it are described by Moore et al. (1966) as having "hurled up steam and black ejecta, much of which fell back in the vent area".

Discussion. Although Taal is sometimes described as a tuff ring (e.g., Sohn, 1996), the main phreatomagmatic phase of the 1965 activity was clearly a maar-forming eruption. Moore et al. (1966) state that the major explosive phase was "caused by lake water gaining access to the volcanic conduit", but as Kokelaar (1986) argued, water (not necessarily lake water) must have gained access to the magma "well below the vent" to induce deep crater excavation, and the invasion of Lake Taal into the crater was a consequence of the cratering and explosive eruption, not its primary cause. Violent eruptions occurred before and after the September 1965 events, and the resulting craters were not connected to the lake.

The Taal phreatomagmatic deposits do, however, contain some old lake sediment (Moore et al., 1966), confirming that when the fissure extended beyond the former shore, lake-bottom sediments were explosively excavated and lake water then invaded the fissure. The key to producing the coalesed maar along the fissure was the groundwater supply, feeding subsurface explosions that excavated the country rock. Subsequent influx of lake water into the coalescing craters is likely to have flooded the conduit(s) and changed the style of late eruption stages, but observations for this part of the eruption are inadequate to draw firm conclusions.

2.2 Rotomahana 1886, the maar-forming part of the Tarawera eruption

Description. In June 1886 an eruption began on Tarawera Mountain on the North Island of New Zealand, and subsequently the NE-SW eruptive fissure extended into the Rotomahana basin to the southwest, for a total length of ~5 km. On the mountain, the Tarawera eruption produced a fissure, fissure-rim cones and a widespread Plinian fall deposit (Thomas, 1888; Walker et al., 1984; Sable et al., 2006, 2009). Along the Rotomahana segment, deep maar craters, now filled by Lake Rotomahana to depths locally exceeding 100 m, were excavated, with country-rock-rich tephra ring deposits up to ~30 m thick emplaced around them, primarily by surges (Nairn, 1979). Casualties from roof collapses in a Maōri village resulting from thick Rotomahana co-surge fall deposits made this New Zealand's most deadly eruption (Keam, 1988). The Rotomahana craters were excavated by phreatomagmatic eruptions, which produced a range of relatively dense basaltic fragments that are ubiquitous throughout the deposit, but vary in proportion, abundance, and dominant grainsize.

Discussion. The increase in eruptive violence that accompanied extension of the fissure to water-rich sites away from the initial eruption centre mimicked at a larger scale the activity described above from the 1965 eruption at Taal. Basalt fragments originating from the

Rotomahana maars are distinct from strongly vesicular basaltic fragments, which were dispersed by the Tarawera eruption plume across the area of Rotomahana surge deposition and admixed into the surge clouds and with Rotomahana's own low plume. Composite bombs from the northeastern end of Rotomahana's fissure display complex histories of assembly and are preserved in a deposit only a few-meters thick, both suggesting low and unsteady rates of magma supply to the erupting fissure (Rosseel et al., 2006). The lag between the eruption's start on Tarawera and initiation of explosive eruptions along the Rotomahana segment could have resulted from slower rise of magma toward the latter, with small magma flux further reflected by the relatively high (commonly ~60-95%) proportion of wall-rock fragments in the deposits, particularly where groundwater occupied a hydrothermal system in thick volcaniclastic deposits cut by the fissure. Using methods of Pyle (1989), an unpublished study by J.-B. Rosseel at Otago University estimated the volume of fall deposits from the Rotomahana maars as ~0.3 km^3, larger than the tephra ring deposits (~0.25 km^3), but we are unaware of comparable volume estimates for fall deposits from other maar eruptions.

3. Ejecta rings and ash sheets: constructive deposits of maar volcanoes

Maar craters are typically if not inevitably underlain by diatremes and surrounded at the eruptive surface by a ring of ejecta (unless the latter has been eroded). Historic maar-forming eruptions provided key observational evidence on formation of maar craters and their ejecta rings (Thomas, 1888; Muller and Veyl, 1957; Moore et al., 1966; Kienle et al., 1980), and their study, supplemented by that of young maar volcanoes, introduced volcanologists to "base surges", dilute pyroclastic density currents that form conspicuous duneform deposits in the ejecta rings (Moore, 1967; Fisher and Waters, 1970; Waters and Fisher, 1971; Crowe and Fisher, 1973; Nairn, 1979; Self et al., 1980), with generally subordinate ballistic and pyroclastic fall deposits. Low plumes of ash were also produced in historical maar-forming eruptions, but the sheet-like deposits of these plumes, or of those from pre-historic eruptions, have yet to be well characterized. Kienle et al. (1980) indicate fall-deposit lobes from Ukinrek, and Giaccio et al. (2007) give qualitative information on fall-deposit distribution from Albano maar.

According to Lorenz (2007), “the radial width of a maar tephra ring may be 2-5 km when measured from the centre of the maar and depends on the size of the maar”. The country rock removed in the process of making the maar crater forms a part of the ejecta, and in many cases is the predominant constituent (Table 3). Where distinctly layered country rock has been excavated, fragments from the uppermost strata are most abundant near the base of the ejecta ring, with fragments from deeper strata appearing higher in the ring (Kienle et al., 1980; White, 1991a). A key feature of all described maar ejecta ring deposits is well-developed stratification (Fig. 4). This varies considerably in style, but indicates that multiple depositional events, each of relatively small volume, form the ejecta rings.

Diatreme excavation and maar crater growth are not necessarily directly linked to the depositional processes that control the geometry of ejecta deposits, so if late eruptions of a growing diatreme are largely captured within the deepening crater or in the diatreme itself, an "undersized" ring may surround it. Conversely, if juvenile supply is high relative to excavation, lateral transport may produce relatively broader rings. Eruptive plumes and currents from sub-vents that are off-center to the crater's center can produce sectoral ring deposits, as observed at Ukinrek (Kienle et al., 1980).

3.1 Pyroclastic density currents and their deposits

Much of the bedding in maar ejecta rings, as in tuff rings, forms by deposition from pyroclastic density currents. These produce a range of layering, typically changing from thick, structureless and commonly block-rich beds near the vent to well developed medial cross-bedding and duneform beds and thin distal planar beds (Yokoyama and Tokunaga, 1978; Sohn and Chough, 1989; White, 1991a; Vasquez and Ort, 2006).

The pyroclastic density currents from maar eruptions commonly contain large amounts of lithic material that is entrained at low temperatures, often from aquifers. Like low-temperature pyroclastic currents from tuff rings or cones, they may produce deposits containing accretionary lapilli, vesiculated tuff, or other indicators of free moisture or water droplets in the moving current (Heiken, 1971; Yokoyama and Tokunaga, 1978; Sheridan and Wohletz, 1981; Walker, 1984; Chough and Sohn, 1990; Vasquez and Ort, 2006). Walker (1985) showed that wet surge deposits are richer in fine-grained material, consistent with increased deposition of fines due to particle accretion, both into aggregates and, potentially, directly onto the bed (Heiken, 1971; Waters and Fisher, 1971). Allen (1984) suggested that accretion in response to current wetness determines whether bedforms of surge deposits show down-current or up-current crest migration during growth, but Cole (1991) and Chough and Sohn (1993) demonstrate that flow characteristics are a more fundamental control.

Maar ejecta rims, like tuff rings, thin away from the crater at rates that are low relative to those of proximal fall deposits that form tuff cones; Sohn (1996) analyzed this relationship and inferred that the low thinning rates result from efficient lateral transport of debris in pyroclastic density currents, whereas scoria cones and tuff cones form primarily from fall deposition from low columns and ballistic distribution. Both fall and current deposition can be significantly affected by syn-eruptive winds, which induce varying degrees of bed-by-bed deposit asymmetry. Directed jets, noted at Taal (Moore et al., 1966) and Ukinrek (Kienle et al., 1980) can also produce asymmetric deposition, and of course where deposition from currents takes place across significant pre-eruption topography there can be pronounced thickening and thinning of beds in response (e.g. Crowe and Fisher, 1973; Nairn, 1979; White and Schmincke, 2000).

Bedding sags are a common feature in beds of maar ejecta rims, with their abundance in part reflecting the excavation of blocks of country rock during diatreme formation (e.g. Kienle et al., 1980; Crowe and Fisher, 1983; White, 1991a; Auer et al., 2006). Sag structures can be difficult to identify in proximal deposits because ballistic showers into moving surges become too intertwined; with distance, currents organize and ballistics become more widely spaced, falling into and through currents (Sohn and Chough, 1989; White, 1991a, Vasquez and Ort, 2006). Often bedding sags are associated with current erosion or deposition around the bombs/blocks (e.g., Sohn and Chough, 1989). Bombs as well as blocks form bedding sags, and bombs are sometimes hot enough to deform on impact (Fig. 5).

3.2 Influence of the substrate on maar ejecta ring geometry and constituents

The degree to which maar-diatremes are excavated into strongly indurated rocks, fractured or weakly indurated rocks, and unconsolidated surficial sediment plays an important role in controlling the types of non-juvenile fragments in the maar ejecta rings, as well as the form of the diatreme structure itself (see Ross et al., 2010-this volume and references therein). Unconsolidated sediment yields grains and lumps (White, 1991a; Auer et al., 2006), poorly consolidated or friable rocks yield a mixture of grains and rounded fragments (Boxer et al., 1989; White, 1991a; Naidoo et al., 2004; Ross and White, 2006), and hard, tight brittle rocks yield angular fragments (Lorenz, 1986; Lorenz and Kurslaukis, 1997; Raue, 2004). Most have some mixture, derived from different depths intersected by the diatreme structure.

3.3 Palagonitization of mafic pyroclasts

Palagonite is an alteration product of sideromelane, and hence is typical of most old mafic maar ejecta rings and of diatreme infill (e.g. Fisher and Schmincke, 1984; White, 1991a,b; Aranda-Gomez and Luhr, 1996; Nemeth et al., 2003). At Surtsey, a tuff cone formed in Iceland's Westman Islands in 1963-64 (Thorarinsson et al., 1964), abundant palagonite was documented in a borehole drilled in 1979, in volcaniclastic deposits which had been affected by a dike-related hydrothermal system for 12-13 years (Jakobsson and Moore, 1986). This study showed that palagonite formed rapidly at relatively low temperatures – 40% of the glass was already replaced at about 60°C, and all of the glass was replaced at 120°C. Young tuff cones are characteristically altered to palagonite (Fisher and Schmincke, 1984), whereas young tuff rings and maar ejecta rims are less so. The most likely reason for this is that tuff cones are built above and around the vent and are more thoroughly permeated by post-eruptive hydrothermal/geothermal fluids, whereas tuff rings, and especially maar ejecta rings, are cut off hydrologically from the magma plumbing and hence geothermal fluid reservoirs. Bed-specific palagonite alteration has been cited for deposits at Pahvant Butte (Utah), with an inferred origin of alteration by water deposited with the particles from wet surges (Wohletz and Sheridan, 1983; Ferrand and Singer, 1991, 1992), but the relationship is only a local, bed-permeability controlled, one in an edifice dominated by cross-cutting palagonite zones (White, 2001). Sideromelane that is deposited in standing water and remains there, or is subsequently exposed in desert environments, can remain notably fresh for many thousands of years (e.g. Christenson and Gilbert, 1964; Oviatt and Nash, 1989) in sites unaffected by hydrothermal fluids.

4. Craters

The term "crater" applies to a geomorphological feature. Volcanic craters form during an eruption, and *may* persist afterwards as a landscape feature. Some craters form during an eruption but are subsequently filled or buried later in the eruption. We distinguish syn-eruptive craters, which are the open pits existing during part or all of an eruption, from post-eruptive craters, which remain after an eruption has ceased and which host lacustrine and other post-eruption sedimentary deposits (Table 2). Other names could be applied to syn-eruptive craters, such as open conduit, or vent; we specify syn-eruptive crater when emphasizing the role of an open pit in capturing ejecta or shaping the eruption's behavior. We use "diatreme walls" where it is either not known whether the walls at the point of discussion were pit walls open to the atmosphere, or where evidence is ambiguous or variably interpreted.

4.1 Post-eruptive craters

The post-eruptive maar crater is the hole in the ground that is present immediately at the end of the volcanic eruption (Table 2). Early wall retreat and sedimentation on the crater floor quickly modifies the original post-eruptive crater's width and depth, at rates that reflect wall-rock characteristics, crater-wall hydrology, and whether streams, marine or aeolian processes carry sediment into the crater. A compilation by Ross et al. (2010-this volume) indicates that typical diameters of post-eruptive craters in Quaternary maars are mostly in the range 200-1500 m, while depths are mostly in the range 10-300 m (this includes some post-eruptive modification in most cases, except for the historical eruptions). The freshest maar craters have diameter:depth ratios between 3:1 and 7:1.

Maar craters formed by historic eruptions at Ukinrek, Nilahue, Taal and Rotomahana filled rapidly with water after, and even during, eruption. For craters containing lakes, depth to crater floor depends on lacustrine sedimentation rates, and wall-retreat rates are mediated by the rim to lake surface height, and the ability of wind to produce erosive waves on the lake.
Maar craters commonly intersect the water table, and are separated from surrounding environments by their ejecta ring. This makes them particularly good repositories for high-resolution sedimentary records because of their low wave energy and often great water depth. The first part of the post-eruptive crater-filling deposits often consists of tens of meters of relatively

coarse-grained material formed by collapse of crater walls, formation of debris fans, rock falls, debris flows, etc. (Smith, 1986; Rayner et al., 1991; White, 1992; Pirrung et al., 2008). Later, quiet lacustrine sedimentation will develop in the deep centre of the lake, including the mud and diatomite layers that are of great use in studies of paleoclimate (Zolitschka, 1992; Ognjanova-Rumenova and Vass, 1998; Kaiser et al., 2006; Lindqvist and Lee, 2009). Lacustrine deposits are also very common in sedimentary deposits of ancient maars (Hack, 1942; Smith, 1986; Nemeth and Martin, 1999). Possible seismites have been identified in maar-lake deposits in Hungary (Nemeth et al., 2008).

Kimberlite craters. The post-eruptive crater fill of kimberlites is often termed "epiclastic kimberlite". In a few cases in Africa such deposits have been exposed by open pit mining. For example, studies of mine benches combined with that of drill cores has shown that the southern crater of the Orapa A/K1 kimberlite in Botswana (Cretaceous age) was filled by rock-fall avalanches, an externally derived pyroclastic flow, aqueous sheet floods, debris flows, and talus fans (Field et al., 1997; Gernon et al., 2009a). The post-eruptive crater of the Mwadui kimberlite in Tanzania (Tertiary age) was filled by (i) granite breccias formed by multiple crater wall collapse episodes; (ii) “resedimented volcaniclastic kimberlite deposits” formed mostly by slumping of portions of the tephra ring into the crater; (iii) kimberlitic turbidites; and (iv) shales (Stiefenhofer and Farrow, 2004). Stratigraphically below this sequence, Stiefenhofer and Farrow (2004) describe pyroclastic kimberlite which they attribute to a “lower crater” position, but we interpret these pyroclastic deposits as a bedded diatreme fill, as discussed below in the diatreme section. In Canada, sedimentation into open post-eruptive craters has produced deep-water shale and olivine-sand deposits in Diavik and Ekati kimberlites of the Lac de Gras field (Nowicki et al., 2004; Moss et al., 2008). For the Fort à la Corne field, Zonneveld et al. (2004) describe terrestrial shales with pedogenic alteration in the crater-filling deposits, while Kjarsgaard et al. (2009a) note marine shales among deposits infilling other craters in the field.

Hopi Buttes volcanic field. Maar crater lakes in the late Miocene – early Pliocene Hopi Buttes volcanic field of Arizona were depositional sites for reworked scoria from nearby scoria cones, as well as base surge deposits from other maars, lava flows, and tephra fall deposits from other vents, deposits of both deep and shallow isolated lakes, and aeolian dunes (Hack, 1942; White, 1989; 1990, 1991b). Such primary and secondary volcaniclastic deposits are intercalated with or overlie lacustrine deposits of crater lakes. Streams entering lakes built a variety of subaqueous sediment accumulations along the crater margins, such as small subaqueous fans and Gilbert-type deltas fed by streams (White, 1989; 1992). Volcanoes are closely spaced in the Hopi Buttes, and here as in similar volcanic fields it is not uncommon for post-eruptive craters to accumulate primary pyroclastic deposits from nearby eruptions, which overlie sedimentary deposits in the same crater (White, 1991b; Gernon et al., 2009b).

Effect of subsidence. Many maar crater deposits also show evidence for substantial and long-lived post-eruptive subsidence (Suhr et al., 2006; Lorenz, 2007), which further enhances their value as sensitive and long-lived recorders of environmental change and distal volcanism. Post-eruptive subsidence also increases the total thickness of sediments preserved in these craters, as well as their preservation potential (White, 1991b). Suhr et al. (2006) have shown that subsidence can also continue long after the crater itself ceases to be a morphological feature. At Kleinsaubernitz maar in eastern Saxony, Germany, a series of lignite beds initially deposited in bogs developed atop the infilled maar crater has been subsiding for millions of years, presumably as a result of diagenetic compaction of the underlying primary diatreme fill and lacustrine crater-filling sequence (Fig. 6).

4.2 Syn-eruptive craters

Formation of a volcanic crater is a necessary initial step in a maar-forming eruption, or phase of eruption, and is initially an explosion crater (Lorenz, 1970; Goto et al., 2001; Ohba et al., 2002). As the eruption progresses, the crater becomes wider and deeper, though different authors provide different explanations for this excavation, and infer substantially different rates of conduit excavation.

Excavating the crater. Rapid excavation has been inferred to result from intense rock-bursting from the diatreme wall into a conduit down which decompression waves move during eruptions driven by volatile decompression (Shoemaker et al., 1962, Hawthorne, 1975; Clement and Reid, 1982; Sparks et al., 2006; Wilson and Head, 2007a, b; Porritt and Cas, 2009), generally associated with wholesale fluidization within the diatreme structure (Reynolds, 1954). Inferred excavation times, where quantified, range from tens of seconds (Kieffer and Wu, 1998) through "at most a few tens of minutes" (Wilson and Head, 2007a) for these models, though Sparks et al. (2007) note the difficulty of accommodating such rates to the high degree of geological complexity noted in kimberlite diatremes. Somewhat more long-lived eruptions of days to years are implicit where diatremes are inferred to form by external modification of small mafic eruptions (cf. Lorenz, 1986; Wilson and Head 2007a, 2007b; Sparks et al. 2007). During historical maar-forming eruptions, which lasted many days, craters grew progressively wider (and possibly deeper) with time as the eruption progressed (e.g., Büchel and Lorenz, 1993).

Lateral collapse of crater walls is common during the eruption (White, 1991a, 1996; Kurszlaukis and Barnett, 2003; Barnett and Lorig, 2007; Gernon et al. 2009a), which contributes to enlargement of the syn-eruptive crater and to the overall diatreme generation.

Filling the crater. A significant portion of the volcanic debris expelled in eruption bursts and plumes will be deposited beyond the crater, but some will inevitably fall back in the syn-eruptive crater. There it accumulates, probably in beds, becoming part of the upper diatreme fill (Lorenz, 1986). Depositional mechanisms for pyroclastic deposits on the crater floor must be a combination of tephra fall and density currents, but this topic has not been addressed in detail. The answer lies in the upper diatreme beds of partly eroded maar-diatreme volcanoes.

Whether or not the syn-eruptive crater deepens with time, despite the accumulation of pyroclastic debris on its floor, depends on the balance between material permanently ejected versus new juvenile material added or porosity created by the eruption. Also, excavation of material at the base of the diatreme structure may drive syn-eruptive subsidence of material lying higher in the diatreme by undermining it. This undermining process involves transfer of material from the base upward into or through the diatreme structure, and has been considered to create a "mass deficiency" (Lorenz and Kurszlaukis, 1997; 2007) at depth, though the undermining acts through creation of open volumes and is insensitive to distribution of mass specifically. Near the end of eruptions, higher rates of magma production relative to diatreme excavation, commonly lead to aggradation of the crater floor, and even to complete infilling and burial of the crater beneath lavas or scoria cones (e.g. Shoemaker, 1962). In many cases, this evolution has been linked to a change from an explosive phreatomagmatic eruption to a cone-building magmatic one (Lorenz, 1986; White 1991a; Nemeth et al., 2001).

Cratering experiments. Goto et al. (2001) examined optimal depths for crater excavation with chemical explosives with volcanological implications in mind: with an 8900 kJ charge, crater diameter increased from 2.03 m for 0.4 m burial, to 2.3-2.7 m at 0.8 m burial, then decreased to only 0.2 m diameter as the charge burial increased to 2 meters; maximum scaled crater diameters were achieved when the scaled depth was about 4×10^{-3} m/J$^{1/3}$. Similar cratering experiments have been performed earlier by the military, although obviously not for volcanological reasons, with both chemical and nuclear explosives (e.g., Bening and Kurtz, 1967; Hansen, 1968). In Bening and Kurtz's (1967) results, the diameter of craters excavated by small chemical explosions in damp compacted sand were at a maximum when the charge was buried to about half the depth of the resulting crater's diameter; the deepest craters required slightly deeper charge burial, to about 2/3 crater diameter, with Hansen's (1968) review showing similar relationships for much larger nuclear explosion craters. Ohba et al. (2002), again using buried chemical explosives, found ejecta trajectories varying from funnel-shaped with shallow explosion depths, through elongate vertical jets for intermediate depths, then to short columns and eventually no ejecta as the scaled depth was increased.

Crater size. An interesting recent development has been the demonstration that crater size in maars of the 561-36 ka Albano field (Italy) scale with the energy of the largest explosions (eruptive bursts), as indicated by ballistic-block dispersal and other indices (Taddeucci et al., 2009). Although, like simple explosion craters, maar-diatremes formed by downward excavation accompanied by wall collapse (e.g. Lorenz, 1973) are inferred to have crater widths proportional to the depth of the deepest explosions, and hence to the depth of the diatremes (Lorenz, 1986), the depth of the post-eruption crater does not necessarily scale in concert. Valentine et al. (submitted) compare a "major-explosion dominated model" – "in which the dimensions of a maar crater are the result of the largest single explosion during the lifetime of a maar" – with an "incremental growth model" to explain the dimensions of Lunar Crater maar in Nevada. They conclude that the latter model, "where a crater grows due to subsidence and ejection of material over the course of many explosions, and the final size is an integrated result of multiple explosive events" is more suitable for Lunar Crater maar based on geologic evidence.

Deep and steep post-eruptive craters. Very deep (several 100s m), steep (e.g. 70°) syn-eruptive to post-eruptive craters or "empty pipes" are inferred for several kimberlites (Jwaneng, Botswana; Skinner and Marsh, 2004; Victor, Canada: Webb et al., 2004; Lac de Gras, Canada: Scott Smith, 2008; and kimberlites in Angola and Siberia, Skinner and Marsh, 2004). These deep open holes are inferred to have remained opened for "some time" (Scott Smith, 2008: perhaps millions of years for some Lac de Gras kimberlites?) following their original (explosive?) excavation. Subsequently they were filled by primary or resedimented volcaniclastic kimberlite – sometimes generated from neighboring volcanoes – or other clastic deposits. For example, the Webb et al. (2004) models for the Victor South and Victor North kimberlites (Jurassic, Ontario) features ≤600 m-deep open "craters" with 70° walls. Deep open holes are, however, mechanically unstable in both mines and volcanoes. In a recent review, Barnett (2008) summarizes this issue:

> *"Consider a shallow (<1 km), steep-sided volcanic crater or upper pipe that is partly empty … If such a pipe is situated in an adequately jointed rock mass the rock will simply unravel in tension with the sidewalls collapsing into the pipe. Volcanic explosions can reduce the overall rock mass strength by the formation of blasting fractures and the destruction of cohesion on joints, making country rock instability more likely."*

We do not consider that steep-walled open craters extending to great depths would be stable over significant post-eruptive time periods; other ways of explaining relationships in these volcanoes should be sought. For Lac de Gras, Moss et al. (2008) provide an alternate explanation, in which infilling took place within a decade or so after eruption, to that of Scott

Smith (2008) in which slow infilling followed a million years of non-sedimentation.

5. Diatremes

As indicated in Table 2, we use "diatreme" to describe the overall structure cut into the substrate (cf. "pipe" of Hawthorne, 1975; Kurslaukis and Lorenz, 2008). As this structure evolves during an eruption and afterwards, different processes operate and different physical environments exist within the structure. Deposits formed in the diatreme structure *during eruption* can be described as diatreme deposits, and include bedded diatreme fill, unbedded diatreme fill (including in zones that cut across bedded fill), and root zone deposits (Fig. 1). Not all diatremes have bedded diatreme fill, and in others there are no exposed zones in which unbedded deposits fill the width of the diatreme structure. Where both are present, however, bedded fill overlies unbedded fill; based on this relationship, we term bedded diatreme deposits *upper diatreme*, and unbedded ones *lower diatreme*, while recognizing that this neither implies a specific structural level, or the universal presence of both types of deposits. The reason for retaining the distinction is that the processes producing bedding in upper diatreme deposits are inferred to take place in syn-eruptive craters open to the atmosphere, whereas processes involved in producing the unbedded lower diatreme deposits may take place in an entirely subterranean setting (see below).

Examples of diatremes that lack any bedded upper diatreme deposits include the suite of kimberlite diatremes summarized and interpreted by Hawthorne (1975), which also lack any ring faults. White (1991a) found that some diatremes in the Hopi Buttes have lacustrine deposits (inferred to have formed in the post-eruptive maar crater; White, 1989; 1990; 1992), directly overlying either unbedded diatreme deposits or thin and disrupted, weakly bedded units. The great majority of the diatreme-complex deposits at Coombs Hills, though exposed at relatively shallow depths (no more than a few hundred meters below the pre-eruptive surface) are various types of unbedded deposits (White and McClintock, 2001; Ross and White, 2006); only one, relatively thin, bedded and in-situ deposit that might be considered a more typical upper diatreme deposit is known (McClintock and White, 2006).

After the eruption, primary deposits comprising the diatreme are enclosed within the diatreme structure, and underlie any deposits formed by sedimentation in a posteruptive crater. As a practical matter, the lowest demonstrably post-eruptive sediments mark the top of the diatreme-fill sequence. Any pyroclastic deposits (Table 2) that overlie early sedimentary deposits belong to the post-eruptive crater fill for the containing diatreme (see section 4.1), and would have been sourced from a different volcano (cf. White, 1991b; Moss et al., 2008; Gernon et al., 2009b).

Some evidence suggests that single diatremes may have been formed by more than one eruption or eruption episode (i.e. the volcanoes are not monogenetic but polygenetic). This could confound sensible description of diatreme elements if not recognized, because (1) magmatic products would not be co-genetic, (2) diatreme-filling deposits of the first eruption would potentially be lithified prior to subsequent eruption(s), and (3) sedimentary deposits formed over periods much longer than that of an eruption or eruption episode could accumulate in the post-eruptive crater prior to a subsequent eruption(s).

5.1 Upper diatreme

Among three highly influential summary diatreme descriptions published in the 20th century (Hearn, 1968; Lorenz, 1973; Hawthorne, 1975), a key difference is the presence or absence of bedded volcaniclastic deposits of inferred primary origin within the upper diatreme. Some diatremes preserve only unbedded pyroclastic deposits overlain by sediments deposited in the post-eruptive maar crater, whereas others contain great thicknesses of bedded pyroclastic deposits in their upper part. We distinguish two varieties of upper-diatreme deposits. One forms during diatreme excavation and subsides as the eruption progresses (type I); the other develops late in an eruption, subsequent to major excavation, and hence is not deeply subsided, commonly filling the syn-eruptive crater and grading upward into surficial tuff or scoria cones (type II).

Type I upper diatreme deposits. At the current levels of exposure, the Paleogene Missouri River Breaks diatremes (Montana) are filled mostly by deeply subsided tuff and lapilli tuff beds (Hearn, 1968; author's observations). Each bed was originally deposited on the syn-eruptive crater floor, yet the bedded sequence in these diatremes extends to more than one kilometer below the pre-eruptive ground surface. Hearn (1968) interprets the Missouri River Breaks diatreme fills, and Lorenz (1973) interpret these and other such sequences, as having subsided, during eruption, along ring faults at the diatreme margin. Evidence for such faults comprises slivers of country rock or poorly consolidated sediments brought down, in some instances with apparently minimal deformation, along the outer walls of the diatremes.

Direct evidence for subsidence of the stratified diatreme fill includes the already mentioned marginal slices of weak wallrock, which must have been buttressed during their descent into the diatreme, and centroclinal (saucer-shaped) bedding (Fig. 7), with dips of pyroclastic beds reaching up to 70° from horizontal near the diatreme walls, but becoming flat towards the middle. In addition, the otherwise subhorizonal bedded sedimentary strata containing the diatremes typically dip inward near the diatreme edges. The only plausible explanation for these features is subsidence of the diatreme fill, most of which must have taken place during the eruption. The bedded deposits in the Missouri River Breaks diatremes are cross-cut by unbedded volcaniclastic domains (Fig. 8) which are interpreted as manifestations of upward-moving debris jets (details below). Some of these would have reached the surface and dispersed tephra into the atmosphere, opening void space at depth that would have allowed continuous subsidence of the bedded diatreme fill.

The ubiquitous bedding of the Missouri River Breaks diatremes is typically of decimeter to meter scale, and weakly but pervasively disrupted internally, making details of individual contacts difficult to follow. At one site there is less disruption of the bedding so that surge dunes (Fig. 9) and bedding sags are locally identifiable. A feature of the ultramafic Missouri River Breaks deposits is the abundance of spherical juvenile pyroclasts that range in size from a few mm to several cm; these are the topic of current work.

Another sequence that we also interpret in part as filling an upper diatreme is the Mwadui kimberlite, described by Stiefenhofer and Farrow (2004). The kimberlite comprises layered deposits enclosed within a wall dipping 30 to 70°, though below a depth of about 230 m, the dip of the kimberlite-wall rock contact becomes a "constant" 75°. This depth corresponds to a major lithological change, of a "relatively sharp and sudden" nature, between the post-eruptive crater infill above, and layered pyroclastic deposits below (Stiefenhofer and Farrow, 2004). The pyroclastic deposits, which dominantly are lapilli tuffs, "closely resemble tuffisitic kimberlite breccia (TKB) in hand specimen" (see discussion on TKB below) and extend to at least 684 m vertical depth based on current drilling. Beds are centimeters to several meters in thickness and bedding ranges from "almost unobservable clast orientation, through diffusely bedded (most common), to very obvious, well bedded areas". Bedding is manifested by varying lapilli tuff compositions (juvenile-rich vs. lithic-rich) as well as varying granite clast sizes, and this allows several normally graded units to be defined. A single intersection of bedded ash was noted by Stiefenhofer and Farrow (2004). Perhaps the lack of recognition that the pyroclastic deposits are filling a diatreme at Mwadui was caused by their bedded nature, since that contradicts the classic South African kimberlite model in which the diatreme is filled by non-bedded TK/TKB (see below).

Type II upper diatreme deposits. A later study of diatremes, in the well-exposed Hopi Buttes volcanic field, led White (1991a) to identify a second class of bedded, "upper-diatreme", deposits (Fig. 10). These are effectively tuff rings, tuff cones or scoria cones that are initiated in open craters and accumulate to the fill the craters. They are diatreme deposits because they are primary pyroclastic units filling the upper part of the diatreme structure cut into country rock. They are distinctive in not necessarily being part of a *maar*-diatreme volcano; because the syn-eruptive crater may be entirely filled and buried late in the eruption, the resulting landform after eruption may not be a maar.

Little or no subsidence can have affected many Hopi Buttes upper diatreme deposits (though there was syn-eruptive subsidence during emplacement of underlying lower-diatreme deposits), because there are large blocks of almost in-place country rock adjacent to its source, and because the upper diatreme deposits extend to the pre-eruption surface (i.e., the syn-eruptive craters have been filled and no post-eruptive craters were left). Bedding styles differ from those in the Missouri River Breaks diatremes; bedded deposits are in broad lenses, with large country-rock clasts included near contacts with confining diatreme walls, and sequences commonly grade upward over tens of meters into agglomeratic deposits with weak bedding and layers of clastogenic lavas, which locally become loaded and injected into subsequently accumulated beds as dikes and sills (Shoemaker, 1962). These bedded deposits are observed at multiple sites to overlie, above irregular contacts locally punctuated with slump deposits, unbedded material of the lower diatreme (Fig. 11).

5.2 Lower diatreme deposits

Lower diatreme pyroclastic deposits characteristically lack bedding, and in both kimberlitic and non-kimberlitic systems, the volcaniclastic fill of the lower diatreme zone is often described as "homogenized" or "well mixed". They have a massive, poorly sorted appearance overall, but there are domains with differing particle populations, which in well-exposed examples are separated by steep sharp to gradational contacts (e.g., Hearn, 1968; Clement, 1982; White, 1991a; Ross and White, 2006). The particle populations may represent different sources, but overall the deposits display "a crude degree of textural and lithological consistency" (Clement and Reid, 1989). Differences in grain size, componentry, or depositional fabric define the different domains, which typically have steep contacts with adjacent domains. The domains are often of roughly columnar form, and are known both in basaltic systems at Hopi Buttes and in Antarctica (see below), and in kimberlites, where they have sometimes been called "feeder conduits" (Lorenz and Kurszlaukis, 2007; Kurszlaukis and Lorenz, 2008).

The lower-diatreme rocks we know best in the Hopi Buttes volcanic field and in Antarctica are exceptionally well-exposed, so we use them to illustrate key features. Antarctic exposures are highly complementary to those at Hopi Buttes with the former providing extensive outcrop of shallow-dipping surfaces that provide a map view of diatreme deposits, whereas Hopi Buttes diatreme rocks are characteristically cliff-forming, providing a cross-sectional view. Although it is often claimed that drill cores from kimberlite diatremes, or/and outcrop in kimberlite mine pits or tunnels provide exceptional 3-D coverage, our view is that the complexity of observed relationships in high-quality weathered natural outcrops would be difficult to capture in active pit mines, harder in underground exposures, and at best severely challenging in even closely (~10 m) spaced drill core.

Hopi Buttes volcanic field. Lower diatreme deposits are well exposed at a number of sites in the Hopi Buttes volcanic field. Two key ones are Standing Rocks West and Round Butte. The former site showcases steeply dipping cross-cutting bodies of pyroclastic rock, defined by differences in the proportions of juvenile fragments and of country rock from various stratigraphic units (Fig. 12). Around part of the diatreme, there is a distinctive unit subparallel to and lying along the walls of the diatreme structure, dominated by coarse blocks (clasts reach a meter in diameter) of country rock derived from both below and above their current outcrop levels.

Key features at Round Butte include the contact between unbedded lower diatreme deposits and overlying weakly bedded ones at the base of the upper diatreme. Along the contact there is a zone up to 5 m thick of once-liquefied mudrock and boulders of sandstone. At one point, this gently dipping contact abruptly becomes vertical, and unbedded diatreme deposits extend as a crosscutting body upward into the bedded deposits. Deformed mudstone blocks, and shear indicators along the contact, support upward motion of the tephra mass now forming the unbedded deposit (Fig. 13). Bedding in the upper deposits is not well defined, and beds are locally broken or weakly folded, which is inferred to reflect deformation that took place while the lower diatreme deposits were being emplaced and modified during eruption. There is an upward transition, with beds in the first few meters above lower diatreme deposits disrupted and not sharply defined, whereas those above this transition zone are distinct and well preserved. The relationship implies that upper diatreme deposits began to accumulate on top of lower diatreme deposits while they were still being formed and developed. Continued development of the lower diatreme during deposition of upper diatreme beds is characteristic of Type I upper diatreme deposits. Round Butte is interpreted to expose an incipient Type II upper diatreme deposit, but the transition from passage of eruptive jets through, and mixing of, lower diatreme material, to accumulation of bedded deposits that undergo no significant subsidence, is not instantaneous.

Coombs Hills. Mafic volcaniclastic deposits of the Mawson Formation at Coombs Hills show distinctive characteristics that are shared with unbedded lower diatreme deposits generally (Fig. 14). Large, tilted to moderately folded, rafts of bedded country rock (sub-tabular bodies, some >100 m along strike of raft bedding) are isolated within the pyroclastic deposits, and commonly progress from unbroken raft interiors to fragmented exteriors with clastic injections. Larger rafts are more common near the edges of the diatreme complex. In addition, "rafts" of variably oriented bedded tuff and lapilli tuff, including layers with surge cross-bedding and accretionary lapilli, are present at many sites within the otherwise unbedded pyroclastic infill. A key feature is that there are numerous steep-sided and irregularly shaped zones of pyroclastic rock, defined by differences in country rock type and proportion relative to juvenile clasts, that make up the deposit. Individual zones are roughly subequant, with simple to complex shapes and typical areas of several hundred m^2 (Ross and White, 2006). Also present are irregular intrusions of basalt with glassy margins and development of fluidal peperite, which typically have areas of tens to a few hundred m^2.

Clasts in the Coombs Hills deposit include, in proportions that vary greatly from zone to zone, coal, siltstone and sandstone from the host country rock; blocks of dike and sill rocks; uncommon granitic basement boulders; fragments of volcaniclastic rocks such as tuff; fragments of peperite (composite clasts); irregular, fluidally shaped glassy basalt fragments from cms to dms in long dimension; quartz sand grains; and dense, angular formerly glassy juvenile clasts commonly having modal grainsize in the coarse ash to fine lapilli range, and only minor (<10%) relicts of fine ash (Ross, 2005). Country rock debris comprises, conservatively, at least ¼ of the volume of the primary volcaniclastic rocks averaged across the Coombs Hills deposit. In-place zones of peperite formed adjacent to coherent intrusions, also represented by composite clasts dispersed throughout the deposit, comprise glassy basalt fluidally mingled, on scales from meters to less than a millimeter, with clastic debris (White et al., 2000) consisting of both tephra and country-rock grains.

The pyroclastic deposits filling the diatreme complex are cross-cut by volcaniclastic dikes up to 75 m across, which are described by Ross and White (2005). These consist of material identical in grainsize distribution and componentry to the average ash-grade matrix of the complex. The dikes are inferred to have formed by extracting the matrix by elutriating it from the host lower diatreme deposits in zones of fluidization below thick sills, then accumulating it until it could be injected *en masse* upward into the host deposits (Ross and White, 2005).

Maegok diatreme, Korea. Kwon and Sohn (2008) provided a detailed field and petrographic study of a freshly exposed outcrop of unbedded diatreme deposits and the adjacent Miocene sedimentary strata. The studied outcrop, at Maegok in southeastern Korea, suggests a <50 m-wide cylindrical mafic diatreme exposed about 100 m below the pre-eruption surface. Since the volcaniclastic deposits are not bedded and the diatreme is narrow, the authors propose that they occupy the lower part of a shallow diatreme. Volcaniclastic rocks are very poorly sorted and comprise a mixture of mafic volcanic fragments up to 1.3 m across and material (mostly gravel) from the enclosing sedimentary strata. Although no bedding is observed in the volcaniclastic rocks, grain size varies and in places, larger clasts “are concentrated in irregular patches with very indistinct and diffuse margins”. These large-clast-rich domains can have sub-vertical boundaries and are inferred to reflect “incomplete homogenization of debris that entered the conduit by different fallback or collapse episodes”.

Also interesting is that Kwon and Sohn (2008) describe the enclosing country rock as having been unconsolidated, porous, permeable and probably water-saturated at the time of eruption, being dominated by rounded pebbles and cobbles (with sandy lenses). The diatreme wall has an irregular, undulating shape at a meter and decimeter scale, probably due to regular inward collapse of the diatreme walls, in a soft substrate emplacement setting. The sedimentary strata are deformed near the diatreme walls: they are either steeply inclined inwards or severely deformed (including folding) over several meters laterally from the diatreme walls. In addition, the enclosing sedimentary deposits contain "irregular patches or dike-like structures that are rich in mafic tuffaceous materials" which cross-cut the bedding and have diffuse and gradational boundaries. These injected bodies of volcaniclastic material demonstrate that the sediments were unconsolidated and that the diatreme-filling material could be fluidized or otherwise mobilized in granular mass flows under some circumstances.

Kwon and Sohn (2008) infer that at higher levels relative to the described outcrop, the diatreme flared upwards rapidly due to the soft-substrate setting, with a tuff ring at the surface. These higher levels unfortunately have been eroded and cannot be examined.

5.3 Subsidence in diatremes

Subsidence of materials in diatremes takes place during eruption, as well as after eruptions end, as noted above. We distinguish among (a) subsidence of slices of intact to semi-intact wall rock along diatreme walls, such as the faults slices in the Missouri River Breaks diatremes (Hearn, 1968); (b) subsidence of large wallrock-derived megablocks, rafts, or "floating reefs" (Wagner, 1914; Williams, 1932; Cloos, 1941; Dawson, 1971; Lorenz 1975), away from the diatreme walls, such as observed at Coombs Hills (White and McClintock 2001, Ross et al., 2008), at the Sterkspruit volcanic complex (McClintock et al., 2008), at Costa Giardini in Sicily (Suiting and Schmincke, 2009; Calvari and Tanner, in press) and in many kimberlites (Mitchell, 1986; Clement and Reid, 1989, Kurszlaukis and Barnett, 2003; Brown et al., 2009), and; (c) bulk subsidence of the overall diatreme-filling material. Both wallrock slices and floating reefs can comprise or include country rock breccias (e.g., Naidoo et al., 2004; Kurszlaukis et al., 2009). The key distinction between them is that wallrock slices must have subsided more or less in contact with the diatreme wall, whereas rafts or floating reefs are separated from the wall, suggesting some lateral transport, and may be rotated. Marginal slices have a special significance, in that they need to be buttressed during subsidence to keep them from falling or sliding away from the wall, whereas rafts and megablocks are separated from the wall, indicating inward movement. Such movement may take place into a syn-eruptive crater (e.g. Brown et al., 2009).

5.4 Kimberlitic diatremes

For kimberlites, as for other diatremes (Cloos, 1941; Reynolds, 1954; McCallum et al., 1975), a key feature is an internal architecture of cross-cutting, steep-sided pyroclastic deposits characterized by different contents of both country-rock xenoliths and olivine and diamond populations (Wagner, 1914; Clement, 1982; Naidoo et al., 2004; Hetman, 2008). We consider that this shared feature of kimberlites and other lower diatremes is a key indicator of processes active in diatreme-forming eruptions.

Tuffisitic kimberlite. The characteristic unbedded diatreme-filling rocks of kimberlite volcanoes have traditionally been called "tuffisitic kimberlite" (TK) or "tuffisitic kimberlite breccia" (TKB) (Clement and Reid, 1989; Field and Scott Smith, 1999; Hetman, 2008; Scott Smith, 2008), depending on the proportion of country rock fragments larger than 4 mm (TKB: >15%). According to Kurszlaukis and Lorenz (2008), who reviewed the topic thoroughly, "the key features of [tuffisitic kimberlite] are its fragmental nature, its massive, well-mixed appearance and a specific matrix mineralogy characterized by the presence of serpentine and microlitic clinopyroxene and the absence or scarcity of carbonate". Other features commonly observed are low abundance of identifiable fine particles and presence of "pelletal" juvenile pyroclasts, often of lapilli size, "which typically consist of an altered olivine kernel surrounded by a thin and often incomplete rim of highly clay-altered material" (Kurszlaukis and Lorenz, 2008). The word "tuffitisitic" originally had an intrusive connotation (Cloos, 1941). Deposits with TK features can also occur in the root zone (see below).

Gas-solids fluidization during emplacement (Sparks et al., 2006; Walters et al., 2006), has often been proposed to explain the lack of fine particles in TK/TKB, and sometimes other features such as their "well-mixed" aspect. In such models, the whole diatreme fill, or at least a significant part of it, becomes fluidized due to vigorous streaming of magmatic volatiles though solid debris; this elutriates the fine particles. As stated by Sparks et al. (2006), "the interstitial matrix (…) might be alteration products of fine ash, but alternatively could be secondary minerals filling the original pores of the primary deposit". Although Walters et al. (2006) infer that fluidization elutriated fines, fluidized medium-sized olivine-cored pelletal lapilli, and coarser lithic clasts with the lapilli by "sinking", a simpler explanation is that fines either were not abundant to begin with, or are not resolvable in the altered kimberlite, and that the mixture of 2 to 3 phi (1/8 - 1/4 mm) modal size olivine lapilli with -3 to -7 phi (8 – 128 mm) polymodal lithic fragments reflects a lack of sorting. This raises the recurring issue of how fluidization, which is effectively a sorting process, can be a major process in forming poorly sorted diatreme deposits.

A very restrictive use of TK remains favored by a number of kimberlite specialists, such as Hetman (2008), who stresses that "several features must be identified within multiple drill cores and thin sections before a rock can be classified as a TK as defined by Field and Scott Smith (1999)". Sparks et al. (2006) and Cas et al. (2008a, 2008b) present spirited critiques of this approach, and we find it surprising that observations at the scale of centimeters and smaller define a single type of rock that is taken to be a defining feature of kimberlite diatremes (according to the traditional model, without TK/TKB there is no diatreme, e.g., Clement and Reid, 1989; Skinner and Marsh, 2004; Webb et al., 2004; Field and Stiefenhofer, 2006; Hetman, 2008). This restriction also rules out, by definition, the existence of any non-kimberlite diatreme, which we view as a major conceptual problem. And in practice, even for kimberlites, such a restriction is deeply flawed because one of these "defining" characteristics, the microlitic diopside, is proposed by Stripp et al. (2006) to be the product of secondary aqueous alteration at temperatures below 370ºC.

Taking a different view, we stress that it is the structure cut into country rock that bounds a diatreme, and that all primary volcaniclastic rocks occurring within the structure comprise elements of the eruptive diatreme infill, regardless of their hand-sample and petrographic textures. Characteristic rock types can occur in different environments, and several workers have noted rocks with the characteristics of TK and TKB in kimberlite deposits that are not in the diatreme itself, for example at Orapa (Field and Stiefenhofer, 2006), or in bedded volcaniclastic kimberlite at Venetia (Kurszlaukis and Barnett, 2003) or Mwadui (Stiefenhofer and Farow, 2004). Kurszlaukis and Lorenz (2008) further argue that the style of hydrous alteration proposed by Stripp et al. (2006) to be involved in formation of TK and TKB textures arises as a specific product of diatreme excavation by way of a "phreatomagmatic process chain" in which groundwater involved in phreatomagmatic fragmentation, along with heat in the kimberlite magma system, create the requisite diagenetic environment for TK formation. However, an alternate explanation is that the fluids which produced serpentine and other secondary minerals are deuteric in origin (e.g., Wilson et al., 2007).

6. Root zone: dike-lower diatreme transition

In a simple diatreme, the pipe structure narrows fairly regularly with depth, and eventually terminates in the dike, or part of a dike, that fed the eruption. The transition from diatreme to dike takes place in a "root zone" (Clement, 1982; Mitchell, 1986; Lorenz and Kurszlaukis 1997, 2007) that is the lowest part of the diatreme structure, immediately above the dike itself, which comprises coherent igneous rock. Features recognized in such root zones include marginal monolithic contact breccias of broken country rock, chaotic mixtures of country rock-derived debris with intruded coherent igneous rock (Clement 1982; Clement and Reid, 1989; White, 1991a), and bodies of coherent igneous rock apparently incorporated as partially molten fragments into the mixture (Lorenz and Kurszlaukis, 1997) plus mixtures including components derived from the overlying diatreme. The key observed vertical change is from pyroclastic rocks with juvenile grains but little or no coherent igneous rock (i.e. dikes or lavas), and bounded by sharp contacts with pipe-wall country rock in the diatreme, to chaotic rocks having complex crosscutting relationships with coherent rock, and bounded by brecciated country rock in the root zone.

The typical description of a root zone given above has two components; a suite of rock properties related to the processes that formed them, and a specific site in the overall structure of the maar-diatreme volcano (e.g. Clement, 1982; Mitchell, 1986). Root zones are incompletely studied, and unfortunately are poorly understood. Most root zones are characterized from drill core as irregular bodies of hypabyssal intrusive rock having various forms of gradational transitions through breccia with volcanic components, to brecciated country rock and finally to undisturbed country rock (e.g., Hetman et al., 2004). One reason for the limited detail available is that although dikes with associated breccias inferred to represent incipient root zones are not uncommon, we know of no natural exposures clearly revealing the upper transition from root zone to unbedded or bedded diatreme fill. For this reason, many limited exposures in mines (Clement, 1982; Kurslaukis and Barnett, 2003), and in outcrop where overlying rocks have been removed or were not developed (White, 1991a; Lorenz and Kurslaukis, 1997), have been integrated to characterize root zone properties.

Complexities arise when rocks sharing many of the same characteristics are present in sites other than the base of the pipe. Such "out of place" root zones are common in some settings (e.g., White and McClintock, 2001), but in this review we take a restrictive view of diatreme "root zones" in which they must lie at the base of the overall diatreme structure. "Intra-diatreme fragmentation zones" are treated separately below in section 7; only true root zones are discussed here. Lorenz and Kurslaukis (2007) provide a comprehensive review and update of root zones, and we summarize features of only a few well-exposed root-zone assemblages in this section.

Ship Rock, New Mexico. The formation of breccias in very simple root-zone style fragmentation zones along dikes associated with the Ship Rock diatreme, formed by eruption of lamprophyric (minette) magma through the stable continental Colorado Plateau, is addressed in detail by Delaney and Pollard (1981). The base of Ship Rock diatreme, and a group of elongate dikes that converge at the diatreme, lie ~750 m below the eruptive surface, in undisturbed flat-lying sedimentary rock. The dikes are weakly to non-vesicular, average 2 m in width and are segmented in plan view (Fig. 15).

Widened zones of the dikes are termed "buds"*, and they, along with bodies of country-rock bearing

* Buds are called "blows" in kimberlites.

breccias, are found both within dike segments, and at segment offsets. Delaney and Pollard (1981) recognize both *monobreccias*, which are zones of coarse angular fragments, partly interlocking, of country rock that lack any dike-derived juvenile fragments, and *heterobreccias*, which have a wider range of fragment sizes and contain juvenile material. Breccias are best developed on thicker dike segments, and those that are most irregular. Although most breccia bodies of both types are inferred to have formed *in situ*, others occur as lenticular screens along dikes and are inferred to have been transported. *Buds* are inferred to represent incipient conduits along which country rock has been entrained and carried away. The *monobreccias* of Delaney and Pollard (1981) have the characteristics of *contact breccias* commonly identified with root zones of kimberlites and other diatremes (e.g. Clement, 1982; Lorenz and Kurszlaukis, 2007), while *heterobreccias* correspond with other varieties of country-rock breccia (e.g. Kurszlaukis and Barnett, 2003) or lithic-rich breccia (Brown et al., 2009). The transported breccia masses correspond to the subsidence breccias of Lorenz and Kurslaukis (2007; and references therein).

Delaney and Pollard (1981) evaluate the potential for brecciation by explosive boring, but note the absence of significant dike vesiculation. They consider a model by McBirney (1959) for boring of plugs by magma churning and convection, but conclude that breccia distribution at Ship Rock is inconsistent with the model, and that such churning and convection faces a large problem in maintaining magma mobility while losing so much heat to entrained and settled country rock. Finally, they consider and rule out brecciation due solely to thermoelastic stresses associated with dike or bud intrusion. They conclude that fragmentation resulted from pore-pressure increases in the low-permeability country rocks, though possibilities not addressed are hydraulic fracturing as suggested by Brown et al. (2007) in a kindred analysis of kimberlite dikes and buds, or phreatomagmatic fragmentation by localized explosions.

Gross Brukkaros, Namibia. A study of a carbonatitic root zone well exposed ~550 m below the eruptive surface in the Upper Cretaceous Gross Brukkaros Volcanic Field, Namibia, is presented by Lorenz and Kurszlaukis (1997), and focuses particularly on country rock fragmentation, concluding that phreatomagmatic explosions were responsible. A key feature of the rock assemblage is preservation of zonation from the country rock stratigraphy (Fig. 16), which shows that fragmentation even to produce breccias in which some mixing has taken place was almost *in situ*. Coarse angular fragments of carbonatite indicate destruction of a precursor dike, whereas intrusive carbonatite demonstrates that magma continued, briefly, to enter the zone of brecciated rock after it was formed. The reconstructed root-zone shape, particularly its depth of only several meters below exposure, together with the rock-fragment zonation, supports the inference that fragmentation must have been more or less in place. This is inferred to rule out breakup by rockfall or other gravity-driven processes and leads to the interpretation that fragmentation resulted from an explosion, the last explosion, of this diatreme.

Hopi Buttes. At Hopi Buttes, White (1991a) described features of a root zone (or rather, intra-diatreme fragmentation zone) at Standing Rocks East, which will be addressed in more detail in section 7 below. Numerous other transitions from dikes to fragmental deposits with associated country-rock breccias are exposed in the field and constitute small root zones, though the root zones of the largest maar-diatremes in the field remain unexposed several hundred meters below the deepest erosional level. One of the small root zones, at Castle Butte Trading Post (Fig. 17), shows a transition, in sandstone, from *in situ* bedded sandstone, to fractured, locally rotated, sandstone blocks set in matrices varying from granulated, apparently liquefied, sandstone to juvenile-rich ones. Juvenile-dominant volcaniclastic rocks of indeterminate origin, with dispersed fragments of sandstone, lie above and inward from the contact with sandstone. Layering of apparently depositional origin increases away from the country rock. This dike terminates northward at the margins of sedimentary deposits that filled a post-eruptive maar crater, mimicking the relationship of buds and breccias on the dikes intersecting the Ship Rock diatreme deposits.

7. Intra-diatreme fragmentation zones

Models for diatreme formation rarely mention primary volcaniclastic deposits formed where dikes have risen into diatremes before being disrupted in higher-level fragmentation zones. We have encountered these deposits in two large-scale diatreme complexes associated with Large Igneous Provinces (White and McClintock, 2001; Ross and White, 2006; McClintock et al., 2008; Ross et al., 2008), and they are now recognized in moretypical, individual, diatremes of the Hopi Buttes volcanic field.

The root zone-like assemblages found at sites higher within the overall diatreme are products of fragmentation sites within the diatreme infill; though they were "root zones" to small diatremes within large diatremes, and similarly linked to a feeder dike from below and the full range of syn-eruptive crater and diatreme processes and structures above, the "country rock" to these zones was unconsolidated diatreme infill. Here we assign such deposits to *intra-diatreme fragmentation zones*, recognizing that though they share many characteristics with root zones, they lack zones in which original country rock is brecciated or otherwise damaged, and are not involved in shaping the overall diatreme structure.

Coombs Hills. Forming an important element of an assemblage of deposits inferred to represent a family of coalesced diatremes at Coombs Hills are non-bedded zones of juvenile-rich tuff breccia, which appear to invade and locally grade into a dominant non-bedded heterolithic lapilli tuff. Juvenile-rich tuff breccia zones with sharp contacts are inferred to have formed late in the history of the Coombs Hills complex, and were studied in detail (Ross, 2005). Among those, one end-member (type 2 TB_j zones of Ross and White, 2006) has highly irregular map-view outlines (Fig. 18), and contains abundant fluidal juvenile clasts, few blocky juvenile clasts (which are abundant in the enclosing deposits), and few lithic fragments except for some largely unbroken quartz grains (Ross and White, 2006). This volcaniclastic material can grade inward into irregular bodies ("pods") of coherent basalt and has some features of fluidal peperites, suggesting non-explosive or mildly explosive fragmentation processes. In this scenario the type 2 TB_j zones are created mostly by peperitic interaction between the newly invading coherent magma and the wet enclosing diatreme-filling volcaniclastic debris, and do not represent sites of violent explosions important in forming the overall diatreme structure, fragmenting significant country rock, and producing eruption clouds and tephra ring deposits. Stellate outlines of these TB_j bodies, are, however, indicative of rapid crack-like injection of particles within the existing diatreme fill, and is inferred to require high-rate injections and hence some form of explosive expansion. Further, the other end-member type of TB_j zones – which are more regular (e.g., elliptical) in plan-view shape, and richer in blocky juvenile clasts and lithics (although still often containing fluidal juvenile clasts) – are considered to have formed due to the passage of debris jets as a result of violent phreatomagmatic interactions (Ross and White, 2006). So there appears to be an incompletely understood range of juvenile clast-forming and deposit-emplacing processes active at Coombs Hills and elsewhere, ranging from violently explosive, through abruptly expansive / weakly explosive, to non-explosive mingling. The locations of the most intense explosions (most effective fragmentation and expansion events) are difficult to pinpoint in diatreme deposits because they disperse their products. Instead, the peripheral damage zones to such explosions, the failed premixes, and the least explosive to non-explosive interactions (such as peperite-generating events and late dikes) are what is preserved in deposits of the intra-diatreme fragmentation zones and in the true root zones.

At Coombs Hills, to explain vertical changes in loci of phreatomagmatic explosions, a complex hydrology within the diatreme-complex fill was inferred, resulting in part from recycling of water involved in phreatomagmatic explosions (White and McClintock, 2001). The presence of abundant composite clasts (White and Houghton, 2006) in the dominant diatreme fill, with characteristics matching those of domains in the peperite suggest an intimate link between intrusion and multi-stage fragmentation (McClintock and White, 2006).

Hopi Buttes. Standing Rocks East was presented as an exemplary representative of a diatreme root zone by White (1991a), based on superb exposure of a chaotic assemblage including irregular dikes, juvenile-rich pyroclastic rocks, lithic breccias (Fig. 19), and associated dikes extending away from the massif. The massif itself contains "jutting fingers of hypabyssal nephelinite intruding (and intruded by) pyroclastic and accidental lithic debris (…)". Within the volcaniclastic rocks, blocks "show abundant evidence of flow alignment and tend to be subvertically oriented" (White, 1991a).

Recent detailed remapping (Lefebvre and White, 2009) reveals that this outcropping massif is enclosed for most of its periphery in a zone of country-rock breccia which has minor admixed juvenile material, and includes blocks of country rock that have dropped >100 m from their source, and others that have been carried upward by a similar amount (Fig. 20). Such mixing is a feature of diatreme-filling rocks (see above), but not of root zones. Moreover, these rocks lie at the edge of a much larger structure that adjoins the area detailed in Fig. 20; this larger structure is roughly 1.5 km in long dimension, cut into intact country rock and filled with a mixture of rotated and subsided country rock blocks up to tens of meters in length and thickness, small pyroclastic complexes, and chaotic breccias. In other words, the "exemplary" root zone lies within a larger structure, the "true" root zone for which presumably lies at some depth, at least 100 m based on clusters of large blocks of country rock lifted from those depths, below the current erosion level. This larger structure itself is one piece of a much larger complex.

8. Synthesis

In this section we synthesize information presented above, discuss the relationship of kimberlites to other maar-diatremes, and provide our overarching interpretations.

8.1 Key indicators

Table 5 lists a number of the previously described features of maar-diatreme volcanoes that are both widely recognized and in our view particularly informative. Perhaps the characteristics of deposits most useful in interpreting eruption processes and development of the characteristic diatreme structure are those that tell us about styles of fragmentation plus associated and subsequent thermal histories.

Other features, and particularly absences of features, are less diagnostic. For example, an absence of documented kimberlite lava flows is unsurprising given the almost universal absence of preserved syn-volcanic surface exposures. At Igwisi Hills, Tanzania, kimberlite lava flows have been identified (Reid et al., 1975; Dawson, 1994), but not all researchers agree with a kimberlite classification for these rocks (see Mitchell, 2008).

8.2 Eruptive processes of maar-diatreme volcanoes

Taking into account the indicators of eruption process listed previously, most diatremes show evidence of cool deposition (absence of welding in lower diatreme and type I upper diatreme deposits), episodic processes (multiple ejecta-ring layers, multiple beds in the upper diatreme, and cross-cutting structures in the lower diatreme and upper diatreme), strong or buttressed diatreme walls (steep pipe walls), in-ground energy release (broken-up country rock), and where preservation extends to the pre-eruptive ground surface, particle distribution by ballistic trajectories and pyroclastic density currents (ejecta rings).

Particle mixing in diatremes. Both the homogenized aspect of many diatremes, and the generation of steep internal contacts, have been attributed to whole-pipe fluidization by some recent workers (e.g. Walters et al., 2006) and in earlier publications (Reynolds, 1954; McCallum et al., 1975). We present alternative views below.

Other recent kimberlite emplacement models call for Plinian to subplinian-scale eruptions (Sparks et al., 2006; Wilson and Head, 2007a; Cas et al., 2008a; Porritt and Cas, 2009). A key feature of these various models is that massive volcaniclastic kimberlite is emplaced during a single major eruptive event. Such scenarios are not compatible with the deposits of maar ejecta rims, where it is clear that small episodic bursts, not sustained Plinian plumes, must explain the genesis of the hundreds of relatively thin beds. Maar-diatreme volcanoes are not known to have generated large ignimbrites or thick widespread pyroclastic fall layers. One example of a diatreme for which a "plinian" model has been proposed is the Fox kimberlite in the Lac de Gras field, North-West Territories (Porritt and Cas, 2009). In this field, there were probably ~150 kimberlites that existed with incompletely filled (post-eruptive) craters, when the Fox eruption occurred; yet the Fox eruptive products are not seen as extra-vent deposits in other craters (B. Kjarsgaard, pers. commun., 2010). Bedding within kimberlites (e.g. Lorenz, 1973; Boxer et al., 1984) is difficult to reconcile with plinian eruption styles, as are all bedded upper diatreme deposits. Further discussion of the in-vent column collapse idea is provided by Brown et al. (2008).

We seek an explanation for diatreme particle mixing that is consistent with evidence that eruptions involve repeated small-volume events over a typical lifetime of a small volcano. In the underground part of the maar-diatreme, an explosion at deep levels in the diatreme structure (e.g., in the root zone) will generate enough gas to mobilize newly fragmented magma and existing debris upward into a "debris jet", typically much narrower than the width of the diatreme, for large diatremes (McClintock and White, 2006; Ross and White, 2006; Lorenz and Kurszlaukis, 2007). Debris jets propagate within the existing diatreme fill and may or may not reach the surface. Experimental studies can be used to illustrate the processes at work (Ross et al., 2008b, 2008c). With time and repetition, debris jets due to episodic bursts will not only create columnar bodies, but will also contribute to homogenizing the vent fill and destroying any bedding features in the lower diatreme. Only the last debris jets leave a record of well contrasted columnar bodies, since older ones would likely have been smeared out or destroyed by subsequent explosions. Post-depositional shaking or movement of the tephra would also have further diffused and blurred the boundaries between these older columnar bodies.

Magma fluidization and kimberlite-specific fragmentation processes. A special type of "fluidization" sometimes invoked in the kimberlite literature is "magma fluidization" due to degassing of magmatic volatiles (e.g., Reynolds, 1954 (for granite); Field and Scott Smith, 1999). This process is thought to generate the pelletal lapilli in kimberlite, with Clement and Skinner (1985) envisaging that this happens before the eruption breaches the surface, though Field and Scott Smith (1999) propose that it occurs afterwards. As discussed by Cas et al. (2008a), the concept of fluidization does not apply to coherent magma. Skinner and Marsh (2004) nevertheless elaborate on this "magma fluidization" model by describing rocks that are petrographically "transitional" between hypabyssal kimberlite (HK) and diatreme-filling rocks, and interpreting fragmentation of kimberlitic magma in the following manner:

> *"Due to the unusually high volatile content of the kimberlitic magma exsolution does not produce the characteristic vesiculation textures (i.e. gas bubbles enclosed in a continuum of silicate liquid) of common volatile-poor magmas. Rather, the silicate magma becomes substantially fragmented in a manner, which is the antithesis of normal magma vesiculation."*

It is unclear why Skinner and Marsh (2004) consider that passage from bubbly magma to a dispersion of fragments of magma in gas is unusual, as this is a fundamental process of all explosive volcanism; the only unusual thing is the inference of the authors that the eruption is not a normal explosive one.

Mitchell et al. (2009), who studied TKs/TKBs from the Wesselton Mine (Kimberley area, South Africa), prefer the terms "subvolcanic magmaclastic kimberlite" to TK/TKB. They state that TKs/TKBs "have no known counterparts with rocks formed from common magma types, suggesting that a specific subvolcanic process is involved in their genesis". They envisage the formation of TK/TKB entirely in the subsurface, as follows:

> *"At depths from 2 km upwards, the proto-tuffisitic kimberlite represents subsurface kimberlite magma which, as a consequence of the introduction of xenoliths and degassing, undergoes a different crystallization history to that which generates typical hypabyssal kimberlite (…). The magma disrupts to form sub-spherical magmaclasts, which are essentially segregations of partially-crystallized magma. (…) The magma crystallizes as an assemblage of diopside, phlogopite, apatite, perovskite and spinel (…). The water-rich remnants of the magma occupy the intermagmaclast areas. These initially crystallize diopside-phlogopite-rich mantles on all pre-existing magmaclasts and*

xenoliths. The remaining fluids are K-Al-rich and crystallize en masse as interclast minor diopside and dominant phlogopite, which is subsequently replaced by chlorite–smectite from late-stage juvenile hydrothermal-like fluids."

In other words, except for chlorite and smectite, there is no post-crystallization alteration of volcaniclastic kimberlite in the Mitchell et al. (2009) model. We are unconvinced by this model for development of tuffisitic kimberlite and tuffisitic kimberlite breccia, and prefer a more conventional fragmentation style. The origin of these textures needs to be interpreted in light of the clear evidence for strong and pervasive alteration (e.g., Stripp et al., 2006; Cas et al., 2008a; Kurszlaukis and Lorenz, 2008).

8.3 Role of magmatic volatiles in the generation of diatremes

As citations above indicate, diatremes are formed by a range of generally but not necessarily silica-poor magmas (e.g. Kött et al., 1995 report a rhyolitic maar), including kimberlite, various ultramafic melts (Hopi Buttes volcanic field – Williams, 1936; Missouri River Breaks – Hearn, 1968), carbonatite (Gross Brukkaros volcanic field: Kurszlaukis and Lorenz, 1997; Lorenz and Kurszlaukis, 1997), through alkali basalts (Lorenz, 1970; 1973; Leys, 1983; Nemeth et al., 2001; 2003), to at least tholeiitic basaltic andesite (Coombs Hills: Ross et al., 2008a). These magmas have a wide range of viscosities and inferred volatile contents, but there are few systematic differences in diatreme deposits or structures beyond those of the juvenile particles themselves (e.g., juvenile fragment shape, vesicularity, crystallinity). We previously listed features observed in diatremes formed by all these magma types in Table 5 (see also McClintock et al., 2009), and reiterate here that despite the unusual petrography and xenocryst abundance of kimberlite, the diatreme structures scarcely vary across this range of magma compositions.

Despite this similarity, kimberlite eruption processes have been widely inferred to differ not only quantitatively, but qualitatively from other diatreme-forming eruptions (e.g., Wilson and Head, 2007a; Cas et al., 2008a; Mitchell et al., 2009). A key reason for inference of this qualitative difference is that kimberlite magmas, formed fairly deep in the mantle by small-degree partial melts, are inferred to have been exceptionally volatile-rich (Mitchell, 1986; Sparks et al., 2006; Wilson and Head, 2007a; Wilson et al., 2007; Kjarsgaard et al., 2009b).

This has some petrological support (e.g., Sparks et al., 2006, and references therein), but is not directly informative of volatile contents of magma at near-surface levels where magma may fragment, since most of this gas may have been lost during kimberlite ascent. On the other hand, present-day elevated H_2O and CO_2 contents in whole-rock samples may result at least in part from post-depositional hydrothermal alteration.

Many kimberlite dykes, not connected to the surface in any way, are non-vesicular (Clement, 1982; Tompkins and Haggerty, 1985; Thy et al., 1987; Kurszlaukis and Lorenz, 1998; Kurszlaukis and Barnett, 2003), indicating that the magmas that formed them were either degassed, volatile-poor to start with, or were volatile rich but that volatile solubility was high enough to prevent exsolution of fluids into vesicles (this would explain, for example, magmatic carbonate in hypabyssal kimberlites). If the magmas were degassed, or volatile-poor to start with, or if volatile solubility was high, then magmatic fragmentation is very unlikely to cause the generation of kimberlitic diatremes. Extensive devolatilization of kimberlite magma en route to shallow crustal levels would mean that volatile contents inferred from glass inclusions or source-area melting calculations are not indicative of magma properties at eruptive levels. In either case, the presence of vesicle-poor intrusions at shallow levels suggests that emplacement is neither necessarily rapid, nor necessarily explosive.

8.4 Role of external water

One of the "battleground" issues for diatreme formation has long been the significance of external water in producing explosive eruptions. In 1963 Alexander McBirney wrote an extensive comment on a paper that described diatreme deposits in Arizona and interpreted them in terms of the then-current fluidization model. His comment made an important point about the difficulty of excavating diatremes with magmatic gases. It regards the source for sufficient volumes of magmatic gas to produce the modest ~30 m wide diatreme; for basalt with 4% initial water, McBirney calculated that 40,000 to 150,000 m^3 of magma needed to be quickly crystallized to yield a reservoir of gas capable of being abruptly released to lift the diatreme's clastic fill ~900 m vertically (he neglected friction, but more importantly the additional energy needed to cut the diatreme or fragment magma or country rock). Such calculations are rarely offered in support of gas-driven fluidization models, nor are the means of accumulating volatiles prior to eruption, but perhaps they should be.

The case for phreatomagmatic explosivity driven by interaction of magma with external volatiles has, since McBirney's early comment, been argued for four decades by Volker Lorenz and collaborators in many papers, and the mechanisms of strong magma-water explosions are well understood (e.g., Zimanowski et al., 1997). Questions remain about how magma-water explosions in messy geological environments may differ from those dissected in the lab (e.g. Kokelaar, 1986; White, 1996), but there is ample evidence from observed eruptions that water can make otherwise quiet eruptions explosive, and result in crater excavation. We regard magma-water explosions as being the best way to generate the repeated small explosions represented in maar ejecta ring deposits and type I upper diatreme beds, the characteristics of the ejecta, and the energy needed to break apart and remove the country rock that is evacuated during excavation of the diatreme structure.

Repeated explosions are observed in historical eruptions, and outcrop and analog models show that they are capable of producing the well-mixed deposits and cross-cutting pyroclastic domains found in many lower diatreme deposits.

Paleo-hydrology has not been given much consideration in most kimberlite studies due to the widespread belief that fragmentation of kimberlitic magma is not due to phreatomagmatic activity – why look at paleo-hydrology if external water plays no role in the eruption? But paleo-hydrology has been addressed in studies of other maar-diatremes (e.g. Németh et al., 2001; Carrasco-Núñez et al., 2007), and the details of magma-aquifer interactions are fertile topics for future studies of maar-diatreme volcanoes involving kimberlite as well as other magma types. The biggest challenge for high-resolution studies is a general one – how to assess the dynamics of transport systems for an un-preserved constituent (the water). The challenge is increased when climate changes, topography-changing erosion or deposition, permeability-changing cementation or dissolution, or tectonic tilting, uplift or deformation have changed the hydrological framework. Simple steps, however, such as determining ranges of likely wall-rock hydrological transmissivity and storativity (~permeability and porosity) should be routinely undertaken as part of any study examining how deep vent structures develop.

9. Concluding statement

Maar-diatremes are small volcanoes, but historical eruptions have produced loss of life, and they are the most immediately dangerous eruptions in monogenetic volcanic fields. Their craters accumulate sedimentary successions ideal for high-resolution studies of past environments and climates. Kimberlitic diatremes are the most important economically, but despite decades of research, numerous open pit and underground mines, and hundreds of kilometers of diamond drilling, they remain poorly understood in volcanological terms, with multiple and strongly conflicting models in place. We are convinced that increased understanding of other maar-diatremes will drive advances in kimberlite volcanology, and is best accomplished by integrating information from all parts of all types of maar-diatreme volcanoes, and from both subsurface and surface observations. High-quality outcrop of diatreme deposits, where it exists, reveals complex assemblages that would be difficult to characterize from isolated exposures or cores. Petrographic analysis is a fundamental research tool, but cannot replace outcrop geology or geochemical analysis of fresh material.

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Table 1: Comparison of kimberlitic diatremes with non-kimberlitic diatremes

Kimberlite diatremes	Other diatremes
Small-volume eruptions	Small-volume eruptions *
No evidence for large dispersal	Small dispersal areas *
Low-viscosity magma	Typically low-viscosity magmas (ultramafic to mafic)
Small feeder dikes	Small feeder dikes
Diatremes up to ~2 km deep, similar crater width	Diatremes up to ~2 km deep; similar crater width
Diatreme character varies with host rock	Diatreme character varies with host rock
Diatremes contain blocks of bedded volcaniclastic rock or rafts of country rocks	Diatremes contain blocks of bedded volcaniclastic rock or rafts of country rocks
Occur in fields of limited geographic extent with eruptions over millions of years	Occur in fields of limited geographic extent with eruptions over millions of years

** for individual diatremes; volume larger in diatreme complexes from Large Igneous Provinces*

Table 2: Definition of some terms employed in this paper.

Term	Description	During eruption (transient), or after?	Related terms (not necessarily equivalent)
Volcaniclastic	"Core term for the family of particles and deposits formed by volcanic eruptions" (White and Houghton, 2006)	During and after	Pyroclastic, epiclastic
Pyroclastic	"A pyroclastic deposit forms from volcanic plumes and jets or pyroclastic density currents, as particles first come to rest. Deposition occurs by suspension settling, from traction, by en masse freezing, or any combination of these" (White and Houghton, 2006). A pyroclastic deposit is one of four types of primary volcaniclastic deposits	During and after	Volcaniclastic, autoclastic, peperite, hyaloclastite. We refer to all in-diatreme volcaniclastic rocks as pyroclastic unless demonstrably reworked by surface processes, or formed as peperite, hyaloclastite, or autoclastic rocks*
Maar, maar volcano	A volcano characterized by a central crater cut into the pre-eruptive ground, surrounded by an ejecta ring, and underlain by a diatreme	After. Also during the eruption of a volcano that ends up producing a maar (but not, e.g., a scoria cone)	Tuff ring, explosion crater, tuff cone, scoria cone
Maar ejecta ring	Ejecta deposited on the pre-eruptive ground around the maar crater	Formed during eruption, persists afterward unless eroded or resedimented partly into the post-eruptive crater	Tuff ring, tuff cone, scoria cone
Scoria cone, tuff cone	Positive cone-shaped landform	Applied to edifice as and after it is formed	Tuff ring, maar (see table 3)
Vent	Open mouth of a conduit through which material is emitted from the ground during an eruption or phase of eruption. In large syn-eruptive maar craters, the vent will likely be smaller than the crater	During eruption; an inactive vent structure may become a debris-filled vent	Feeder vent, feeder channel, conduit
Conduit	Transient passageway through which material can travel upward toward the vent	During eruption; conduits vary in form and scale	Feeder channel, feeder vent, vent, dike, blow
Crater	Pit open to the sky, open hole	During eruption: syn-eruptive crater After eruption: post-eruptive crater	During: open vent, upper conduit, upper diatreme, primary pyroclastic deposits (eruption-fed granular flows, cf. Brown, Gernon, are density currents) After: depositional basin, maar lake, reworked volcaniclastic deposits
Crater-fill deposits	Deposits of any type and origin filling the post-eruptive crater. Excludes pyroclastic rocks formed during eruption since these become part of the upper diatreme	After	Diatreme deposits

Table 2 (cont.)

Term	Description	During eruption (transient), or after?	Related terms (not necessarily equivalent)
Diatreme structure	Relatively steep-sided (typically cone-shaped) and regular structure within enclosing country rocks or deposits. Extends from syn-eruptive crater floor to feeder dike at base	During eruption; structure evolves by deepening, wall erosion or collapse, and may be filled or temporarily empty of deposits at different times in eruption	Vent, conduit, blow, crater (can be temporarily equivalent if no deposits in diatreme)
Diatreme deposits	Primary volcaniclastic infill of the diatreme structure, mostly pyroclastic but can contain rafts of country rock, slumped tephra ring, and peperite.	During and after	Many related terms; see next 5 rows.
Upper-diatreme deposits	Bedded pyroclastic deposits enclosed within a diatreme structure, and overlying any lower-diatreme deposits present. The bedded deposits can be cross-cut by non-bedded zones of volcaniclastic material	During and after; produced by sedimentation on syn-eruptive crater floor during eruption	Centroclinally dipping beds, bedded diatreme fill
Lower-diatreme deposits	Unbedded pyroclastic deposits low within diatreme	During and after. See text for discussion	Diatreme infill, diatreme facies, tuffisitic kimberlite, tuffisitic kimberlite breccia, etc.
Debris jet	An upward moving mixture of pyroclastic debris and gasses, ±liquid water, enclosed by existing non-fluidized non-consolidated diatreme fill, generated by an explosion within the diatreme or root zone (Ross and White, 2006; Ross et al., 2008b, 2008c)	During. Leaves a body, cylindrical, conical or irregular, filled by non-bedded pyroclastic debris, typically with moderate to steep contacts relative to the enclosing diatreme fill. Recognized by different grain size and/or composition relative to the enclosing material. Contacts can be sharp or gradational. Will not be recognized if there is no componentry or granulometric contrast	Feeder channel, temporary conduit, fluidization channel [not our interpretation]
Root zone	Chaotic zone of irregular form and including coherent and clastic rock, marginal breccias, at base of diatreme, above feeder dike	During: active root zone After: root zone (deposits)	Blow, fragmentation , zone
Intra-diatreme fragmentation zone	Chaotic zone of irregular form and including coherent and clastic rock, but not in contact with country rock	During: site at which magma fragmentation occurs After: intra-diatreme fragmentation zone deposits	Root zone, peperite complex

* This is because all proposed explosive diatreme eruptive mechanisms deposit particles from jets, plumes, or density currents.

Table 3. Comparison of the different types of small basaltic volcanoes [1, 2]

	Scoria cones	Tuff cones	Tuff rings	Maars
Crater above/below original ground surface	**Above**	**Above**	**Above**	**Below**
Volume of erupted material (m^3)	10^5-10^9	10^5-10^9	10^5-10^9	?
Basal diameter of cone (W_{co}, km); averages from Head et al. (1981)	0.25-2.5	-	1.6	1.4
Diameter of crater compared to basal diameter	Small	Medium	Large	Large
Height of cone or rim; averages from Head et al. (1981)	0.18 W_{co}	Often >100, but <300 m	<50 m [3]	<30 m
Crater or rim diameter (Head et al., 1981)	0.4 W_{co}	<0.1-1.5 km	0.2-3.0 km	0.2-3.0 km
Height/rim diameter ratio (Head et al., 1981)	-	0.5-0.2	0.13-0.05	0.13-0.05
Associated lava flows	Very abundant; lavas typically comprise great majority of total eruptive mass	Minor	Minor	Minor
Primary slope of construct/dip of beds	25-38°	10-30°	Sub-horiz to 20°	Sub-horiz to 20°
Stratification	Crude to coarsely stratif.	Massive or weakly stratif.	Well-stratif.	Well-stratif.
Dominant grain-size	Blocks & lapilli	Ash & lapilli	Ash & lapilli	Ash & lapilli
Proportion of non-juvenile fragments	Generally << 1% (but see Carey and Houghton, 2010)	Small (a few %, or less)	Variable [4]	Up to 90%
Vesicular juvenile fragments (scoria and spatter)	Very abundant	Possible	Possible	Possible
Dense to poorly vesicular juvenile fragments	Rare	Typically abundant	Typically abundant	Typically abundant
Transport mode of pyroclasts	Fallout, grain avalanches	Base surges [5, 6], fallout and remobiliz.	Base surges & fallout	Base surges & fallout
Vesiculated tuffs [7]	Absent	Possible	Possible	Possible
Accretionary lapilli	Absent	Common	Common	Common
External water	Opening phase?	Abundant	Abundant to limited	Limited
Underlain by diatreme	No	No	Shallow diatreme	Deep diatreme

(1) Note that volcanoes with mixed characteristics are common, and record variations from the predominant eruptive process forming the volcanic edifice (e.g. Houghton and Schmincke, 1989).

(2) Table after Sheridan and Wohletz (1983), Fisher and Schmincke (1984), Lorenz (1986), Sohn and Chough (1989, 1992), White (1991b), Vespermann and Schmicke (2000), and the authors' observations in a number of sites. Table modified from Ross (2005).

(3) The Capelinhos tuff ring is about 200 m thick (Waters and Fisher, 1971)

(4) Phreatomagmatic beds in the Ohakune tuff ring (NZ) are characterized by a "very low abundance" of non-juvenile material, attributed by Houghton and Hackett (1984) to very shallow explosions. A commonly cited figure is 1-5% non-juvenile fragments (Lorenz, 1986), but two Korean tuff rings contain over 10% accidental material (Sohn, 1996) and the phreatomagmatic phases of the Rothenberg scoria cone contain up to 55% non-juvenile material (Houghton and Schmincke, 1989, their Fig. 24b).

(5) Sohn and Chough (1992) and Sohn (1996) consider that pyroclastic surges have only a minor role in the formation of tuff cones.

(6) Base surges are overpressured flows (Valentine, 1998), but the term has been used much more broadly to refer to dilute pyroclastic density currents, especially those associated with phreatomagmatic eruptions.

(7) Care is needed in interpretation of vesicular ("vesiculated") tuff because vesicles can be produced by soil-forming processes (Slate et al., 1991).

Table 4: Summary of historical observations of four maar-forming eruptions.

Observation	Nilahue 1956	Rotomahana 1886	Taal 1965	Ukinrek 1977	Interpretation
Activity from multiple vents on fissure	Early	Extension from non-maar eruption	Mid-late	Mid-late	Maar eruptions are fed by dikes, and need not be formed during the first activity of an eruption
Short-lived eruption	5 days continuous, with low-level activity for 3 months	~5 hrs (linked magmatic Tarawera eruption had begun ~ 1 hr before), and hydrothermal eruptions at Waimangu for days after	62 hours	10 days	Maar eruptions are not long-lived or large-volume eruptions; large eruptions would bury initial maar-crater structures with later eruptives
Episodic explosivity	20-30 min bursts, every 30 minutes in main phase	No reports, but deposits consist of many separate beds	Almost continuous from different sites along fissure, waning to 5-10 minutes explosive phases	Yes (Kienle et al., 1980)	Observed maar eruptions have been both small-volume and episodic
Minor strombolian phases	-	-	-	35 scoria fall layers intercalated with the phreatomagmatic deposits (Büchel and Lorenz, 1993) at Ukinrek East	Minor strombolian activity occurred more or less simultaneously with the phreatomagmatic activity, typically from a different vent, often at the same time [(1)]
Other…		One "end" of eruption that began as a plinian fissure eruption, which continued throughout Rotomahana eruption			Excavation of a maar-diatreme volcano can follow initiation of a different type of eruption

[(1)] cf. 2010 Eyjafjallajokull video clips showing incandescent bursts simultaneous with steamy ash plume generation

Table 5: Features indicative of eruptive processes in maar-diatreme volcanoes

Feature	Site	Significance
Primary current bedding, many layers [(1)]	Ejecta ring, mega(?)blocks	Multiple, small pyroclastic currents produced during the maar-forming eruption, can continue after morphological crater has stabilized near end of eruption
Country rock fragments[(2)]	Ejecta ring, diatreme, root zone	Pyroclastic deposits of maar-diatreme volcanoes have variable but generally substantial contents of country rock fragments, particularly enriched in the root zone; one or more processes of fragmenting the volcanoes' host rock are required. Shallow entrainment of loose conduit-wall material is not restricted to maar-diatremes [(3)]
Fresh country rock fragments; low-temperature alteration	Diatreme	Country rock fragments in diatremes are not commonly thermally metamorphosed (except where contained in juvenile fragments); alteration is generally hydrous. "Structural and contact/metasomatic effects associated with diatreme formation are remarkably few."[(4)]
Unwelded pyroclastic deposits	Ejecta ring, diatreme	Unwelded deposits indicate that juvenile fragments were cooled during fragmentation, or during transport to depositional sites; rare large juvenile bombs may land while deformable.
Welded deposits or clastogenic coherent rocks	Root zone, Intra-diatreme fragmentation zones[(5)]	Welded deposits indicate local emplacement of still-hot fragments, indicating heat retention and minimal pre-deposition cooling
Variably, but generally weakly, vesicular juvenile pyroclasts	Ejecta ring, diatreme, +/- root zone	Pyroclast vesicularity is typically high for products of magmatic eruptions unless the magma has been degassed. [(6)] Carbonatite pyroclasts are typically weakly to non-vesicular[(7)], though highly vesicular carbonatitic melilitic lapilli are known[(8)].
Blocky to fluidal pyroclast shapes	Ejecta ring, diatreme, +/- root zone	Shape of juvenile pyroclasts indicates fragmentation of magmas that are generally not highly vesicular, either because volatiles were lost prior to fragmentation, or because fragmentation quenched the magma prior to substantial exsolution of volatiles from the melt (Houghton and Wilson, 1989). Fluidal ash-grade / fine lapilli clasts are described predominantly and abundantly from kimberlite and carbonatite[(7)].

(1) Typical of all ejecta rings.

(2) Typical of all maar-diatreme volcanoes.

(3) Carey and Houghton, 2010

(4) Mitchell (1986) (p. 76).

(5) Lefebvre and White (2009).

(6) Cashman and Mangan (1994); Lautze and Houghton (2007)

(7) Keller (1981)

(8) Stoppa (1996)

Figures:

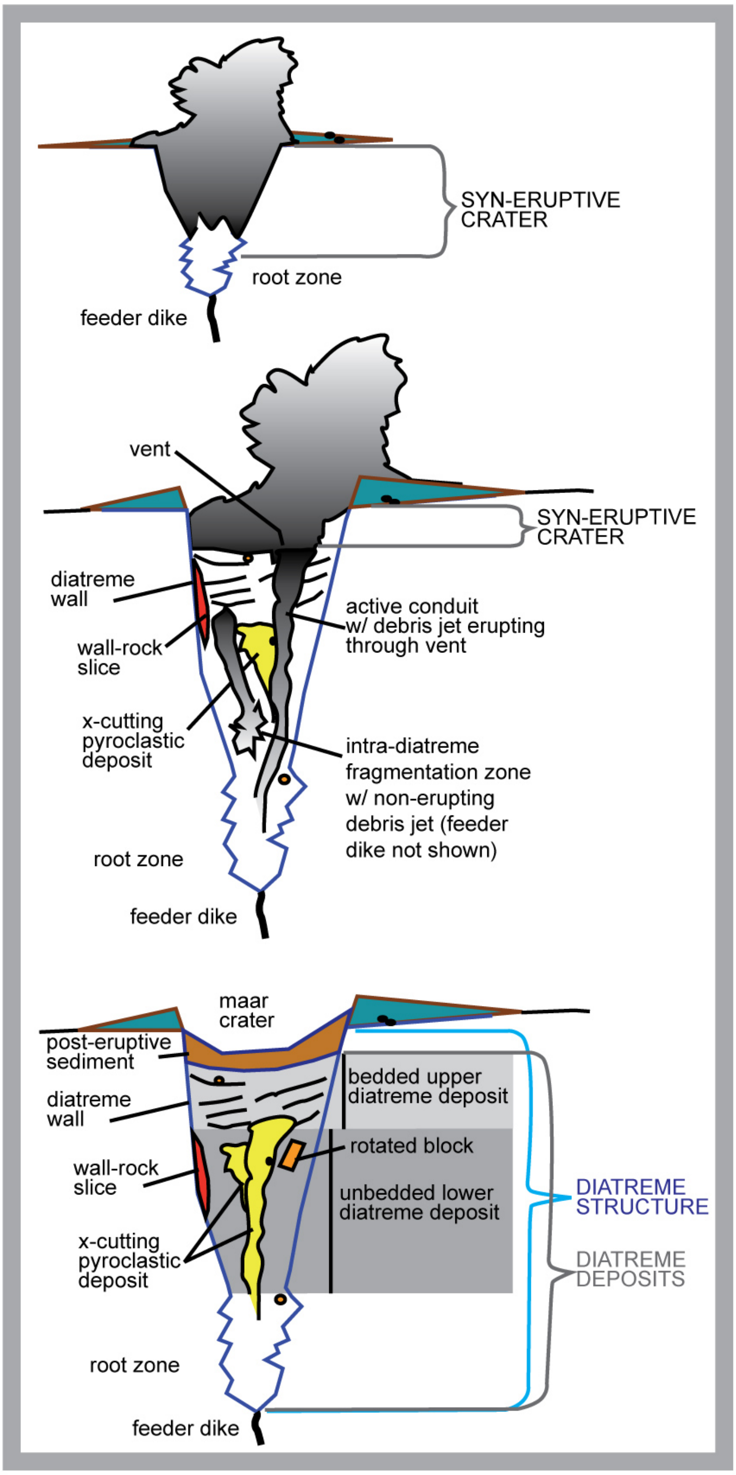

Figure 1. Schematic cross-sections of an erupting maar-diatreme volcano (bottom), and of the volcano after crater has been partly filled with post-eruptive sediments (top) with labels indicating terms used in this review. For definition of terms, see Table 2.

Figure 2. Remnant of the late tuff cone within coalesced maar craters of the Taal 1965 eruption, from http://www.uhh.hawaii.edu/~csav/gallery/decker/philippines_taal.php. (Photo copyright R.W. Decker, website accessed December 9, 2010.) A 1965 photo by J.G. Moore shows the tuff cone was originally complete.

Figure 3. Rotomahana craters soon after formation; view from top of Tarawera Mountain. Photo from Burton Brothers. Source: *http://collections.tepapa.govt.nz*

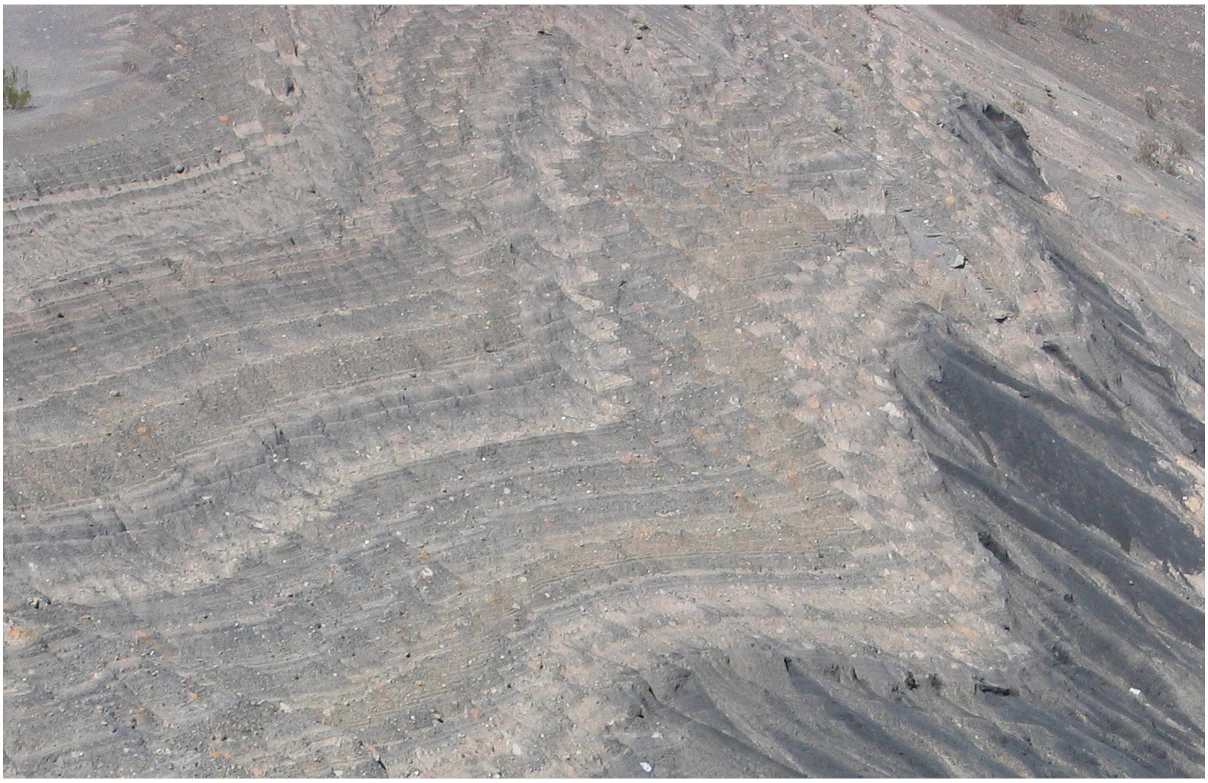

Figure 4. Bedded tephra in an ejecta ring from Ubehebe, Death Valley, California. Individual beds are centimeters to a couple of decimeters thick.

Figure 5. Large bombs with impact sags in maar-rim deposits in the Crazy Waters area, Hopi Buttes volcanic field. Bomb at left appears to have flattened upon impact, and is ~0.7 m long.

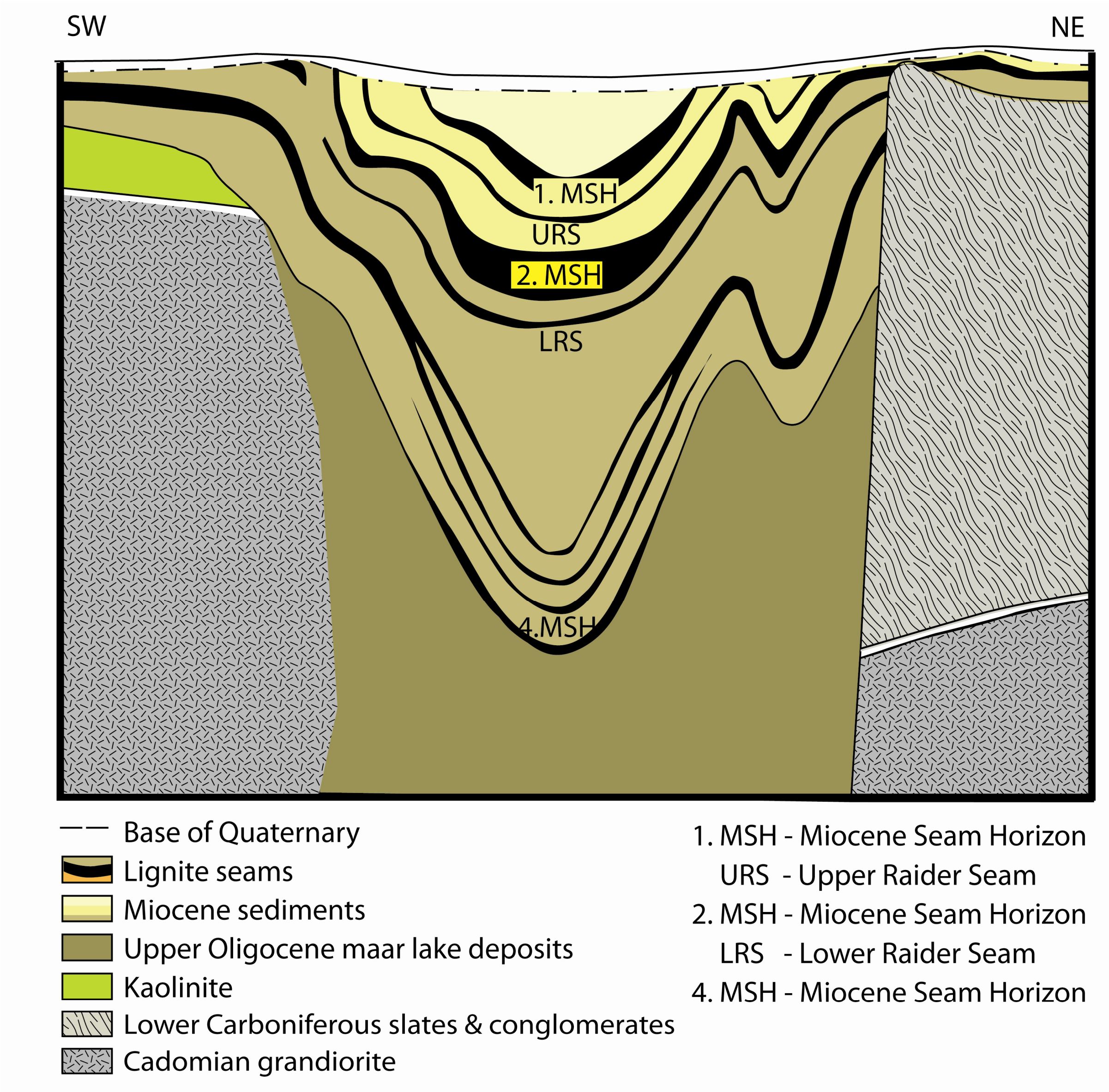

Figure 6. Subsidence of originally subhorizontal lignite seams in Kleinsaubernitz maar records subsidence exceeding 200 m, and continuing today. Diagram after Suhr et al. (2006); 10x vertical exaggeration.

Figure 7. Bedded diatreme fill at Hay Coulee diatreme, Missouri River Breaks, Montana. The vertical dimension of the photo is about 40 m.

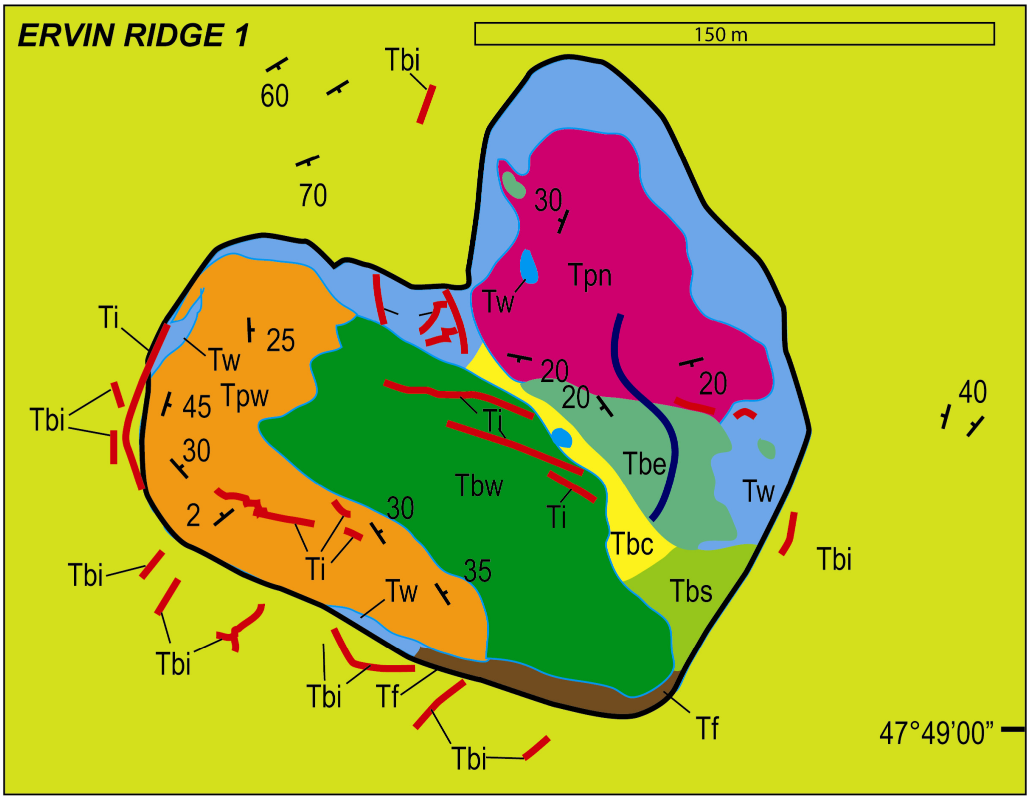

Figure 8. Geological map of the Erwin Ridge 1 diatreme, Missouri River Breaks, after Hearn (2009). Country rock is Bearpaw Shale; Tbw, Tbs, Tbe and Tbc are tuff breccia units; Tpw and Tpn are bedded pyroclastic deposits; Tf and Tw are marginal slices of different country rocks downdropped in the structure; Tbi=tuff-breccia dikes; Ti=coherent dikes. Units within diatreme structure (bold outline) have mostly subvertical contacts with one another.

Figure 9. Photograph of cross-bedding in the diatreme fill at the Lone Tree Ridge diatreme, Missouri River Breaks. Swiss army knife for scale.

Figure 10. “Upper diatreme” beds at Hoskietso Claim, Hopi Buttes volcanic field, Arizona. The exposed sequence is ~100 m high, and the uppermost deposits extend to the elevation of the pre-eruption ground surface.

Figure 11. Transition (white line) between weakly bedded “upper diatreme” and unbedded lower diatreme at Round Butte, Hopi Buttes volcanic field (see White, 1991a). The dashed line identifies a domain richer in country rock fragments. The vertical dimension of the photograph is ~8 m high.

Figure 12. Photograph from Standing Rocks West, Hopi Buttes volcanic field, illustrating a steeply dipping contact between two bodies of unbedded pyroclastic rock in the lower diatreme deposits. Field of view ~ 4 m high.

Figure 13. Round Butte diatreme in the Hopi Buttes volcanic field. The exposed sequence is ~25-30 m high.

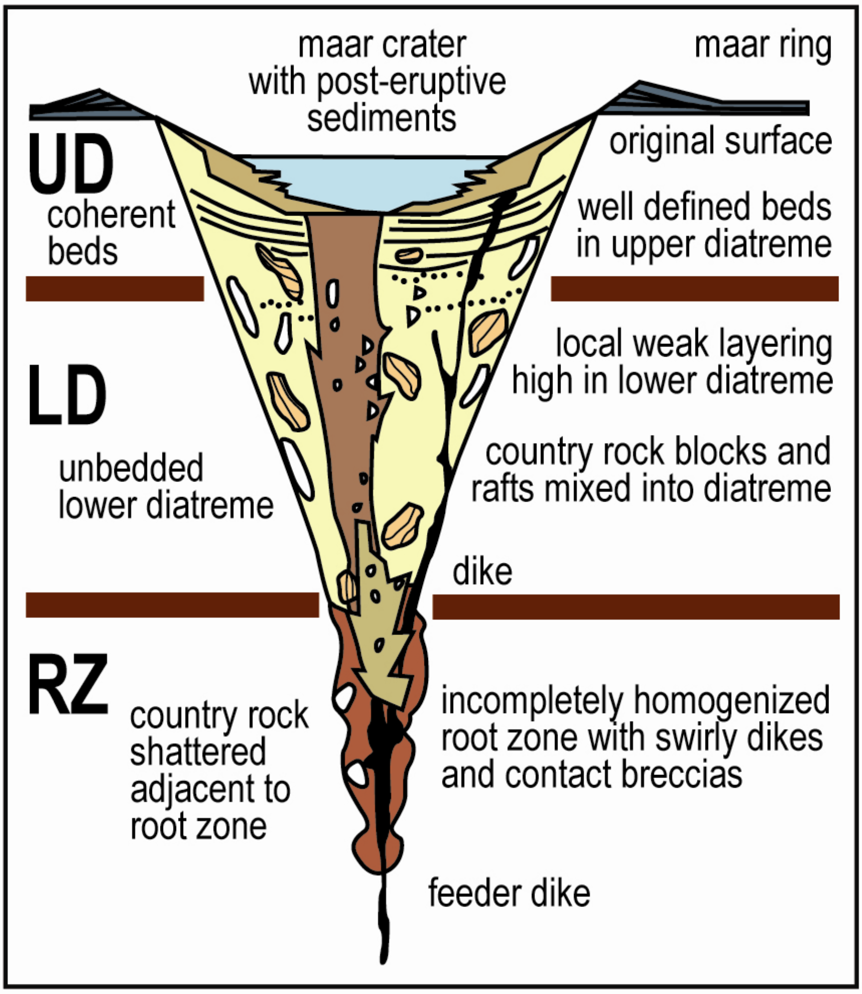

Figure 14. Schematic diagram representing the Coombs Hills diatreme complex as a single diatreme to illustrate the typical features, slightly modified from White and McClintock (2001).

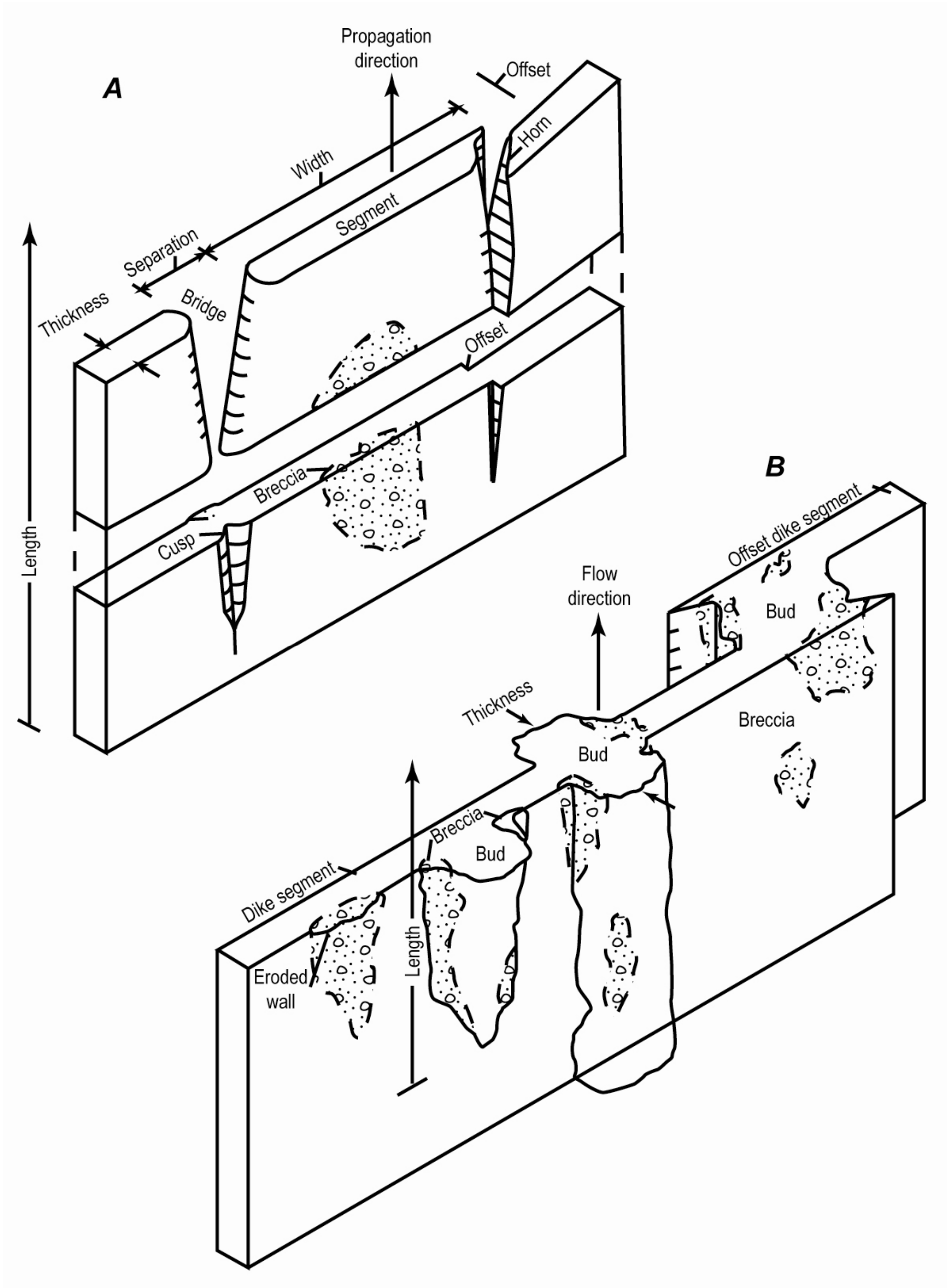

Figure 15. Incipient root zone features at Ship Rock diatreme, New Mexico: (A) Near the tip of an upward propagating dike, segmentation produces a series of dike segments, often offset, separated by bridges of country rock. (B) Continued flow through the dike allows coalescence of segments; thickened dike "buds" occur both where offset segments rejoin, and in segment midzones, and are commonly associated with monobreccias and sometimes heterobreccias. Diagram redrawn after Delaney and Pollard (1981).

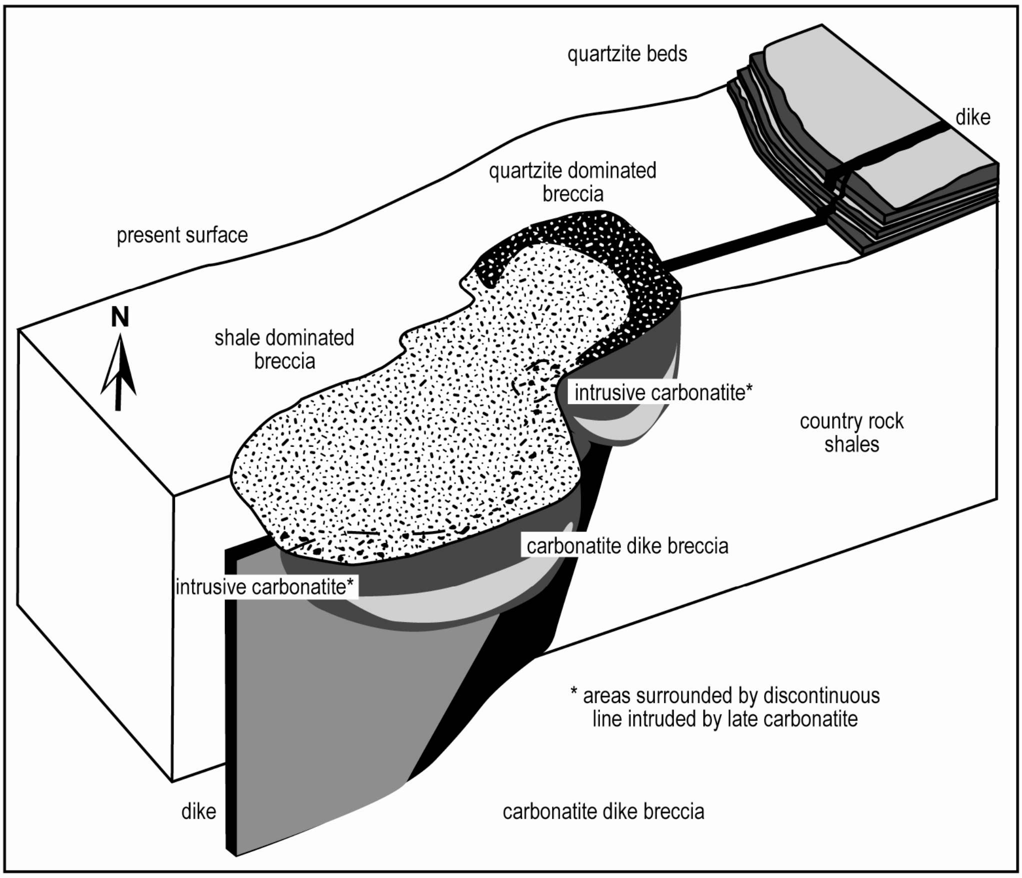

Figure 16. Root zone comprising monobreccia and heterobreccia with associated carbonatite feeder dike, Gros Brukkaros, Namibia; long-axis of root zone ~20 m. Redrawn after Lorenz and Kurszlaukis (1997).

Figure 17. Transition to fragmental deposit in an elongate, dike-like root zone at Castle Butte Trading Post, Hopi Buttes volcanic field. Note preserved bedding in sandstone host, which was locally disaggregated and liquefied to produce peperites with the intruding dike. There is also well-developed layering of uncertain origin in the pyroclastic root zone deposits, which incorporate varying amounts of country rock fragments. Vertical dimension of feature shown is ~15 m.

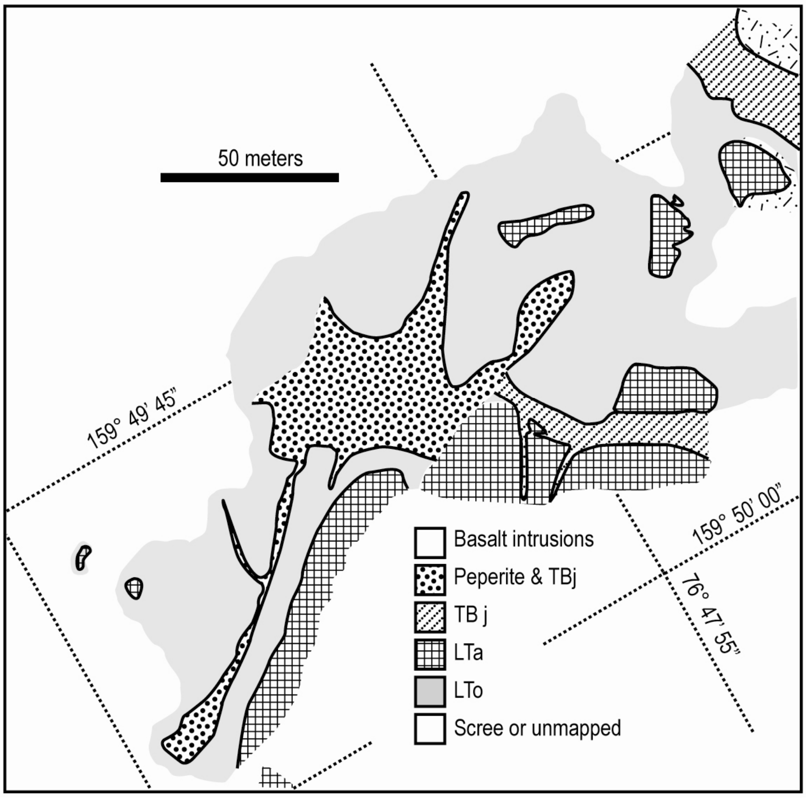

Figure 18. Detailed map of a small portion of the diatreme complex at Coombs Hills showing a juvenile-rich tuff breccia domain with stellate form (after Ross and White, 2006). See text for discussion.

Figure 19. Standing Rocks East, Hopi Buttes volcanic field. The feature shown is ~20 m tall, and comprises deposits of an intra-diatreme fragmentation zone. Dark rock is coherent to transitional magmatic (melanephelinite) rock in irregular and crosscutting geometries. The transitional rocks include both peperitic contacts and clastogenic coherent bodies. These crosscut volcaniclastic rock having orange shading from incorporation of sandstone fragments from ~200 m higher in the stratigraphy.

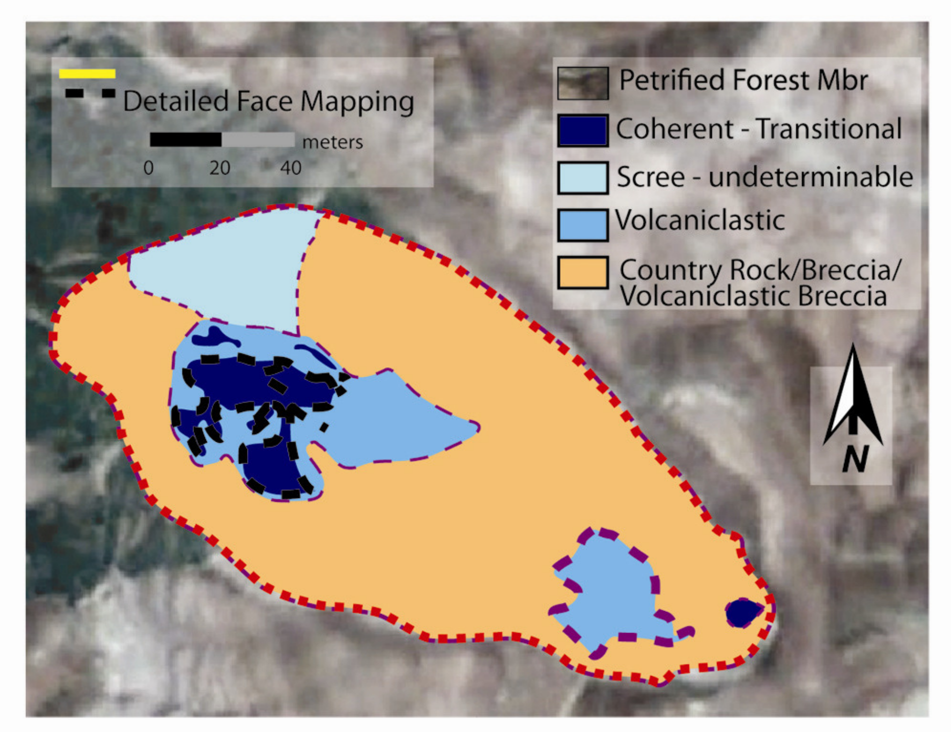

Figure 20. Geological map, on an aerial photo background, of Standing Rocks East in the Hopi Buttes volcanic field, showing assemblage of coherent and pyroclastic rock exposed 300 m below the eruption surface. The Petrified Forest Member is part of the Chinle Formation and consists of variegated mudrock. The country rock blocks inside the diatreme comprise blocks extracted from units spanning at least 500 m of Mesozoic and Cenozoic strata.