Cretaceous basaltic phreatomagmatic volcanism in West Texas: Maar complex at Peña Mountain, Big Bend National Park

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Received 23 March 2007; accepted 25 January 2008
Available online 10 March 2008

Abstract

A structurally complex succession of basaltic pyroclastic deposits produced from overlapping phreatomagmatic volcanoes occurs within Upper Cretaceous floodplain deposits in the Aguja Formation in Big Bend National Park, West Texas. Together with similar basaltic deposits recently documented elsewhere in the Aguja Formation, these rocks provide evidence for an episode of phreatomagmatic volcanism that predates onset of arc magmatism in the region in the Paleogene. At Peña Mountain, the pyroclastic deposits are ≥70 m thick and consist dominantly of tabular beds of lapillistone and lapilli tuff containing angular to fluidal pyroclasts of altered sideromelane intermixed with abundant accidental terrigenous detritus derived from underlying Aguja sediments. Tephra characteristics indicate derivation from phreatomagmatic explosions involving fine-scale interaction between magma and sediment in the shallow subsurface. Deposition occurred by pyroclastic fall and base-surge processes in near-vent settings; most base-surge deposits lack tractional sedimentary structures and are inferred to have formed by suspension sedimentation from rapidly decelerating surges. Complexly deformed pyroclastic strata beneath a distinct truncation surface within the succession record construction and collapse of an initial volcano, followed by a shift in the location of the conduit and excavation of another maar crater into Aguja strata nearby. Preserved portions of the margin of this second crater are defined by a zone of intense soft-sediment disruption of pyroclastic and nonvolcanic strata. U–Pb isotopic analyses of zircon grains from three basaltic bombs in the succession reveal the presence of abundant xenocrysts, in some cases with ages >1.0 Ga. The youngest concordant analyses for all three samples yield a weighted mean age of 76.9±1.2 Ma, consistent with the presence of Late Campanian vertebrate fossils in the upper Aguja Formation. We infer that the volcanism is related to the intraplate Balcones igneous province, which was emplaced in the same time frame in a marine setting farther east.

Keywords: phreatomagmatism; Aguja Formation; Trans-Pecos Texas; Balcones igneous province; Cretaceous volcanism

1. Introduction

The Trans-Pecos province of Texas is well known in the literature for the extensive Cenozoic igneous rocks present in the region (Fig. 1), which were emplaced during arc magmatism and subsequent Basin-and-Range extension (Price et al., 1987; Henry et al., 1991; White et al., 2006). Recently, Breyer et al. (2007) documented restricted outcrops of near-vent basaltic phreatomagmatic pyroclastic deposits within Upper Cretaceous strata in the eastern part of the Trans-Pecos province. These strata significantly predate the onset of arc volcanism in the region, leading Breyer et al. (2007) to suggest that the near-vent basaltic deposits represent a westward extension of the intraplate Balcones igneous province, which comprises a large number of igneous centers active in the Late Cretaceous farther to the east (Fig. 2). Here we report a second, better exposed and more complex succession of proximal Upper Cretaceous basaltic pyroclastic deposits in Trans-Pecos Texas, at Peña Mountain in Big Bend National Park. The succession provides the opportunity to examine in detail the style of eruptions and the nature of source vents associated with a hitherto unknown Cretaceous volcanic episode in the Trans-Pecos region.
2. Geological framework

Voluminous magmatism occurred in the Trans-Pecos province primarily at 48–16 Ma (Henry et al., 1991), although one plutonic complex in the western part of the province has yielded isotopic ages as old as 64 Ma (Gilmer et al., 2003). Igneous rocks emplaced during the early part of this time frame define the easternmost extent of the Cordilleran magmatic arc, which swept eastward in the Cretaceous and Paleogene. Arc volcanism was widespread in Arizona and western New Mexico in the Cretaceous (Fig. 2), but the arc front did not reach the Trans-Pecos region until the Paleogene (Damon et al., 1981; Henry et al., 1991). Prior to the work of Breyer et al. (2007), the only known Cretaceous volcanic rocks in the region were distal felsic tuffs presumably derived from arc volcanoes farther west (Lehman et al., 2006).

Both the basaltic pyroclastic deposits reported by Breyer et al. (2007) and those described herein occur in the Upper Cretaceous Aguja Formation, within the upper shale member (Fig. 3), an informally named unit at the top of the formation that is exposed near and within Big Bend National Park (Lehman, 1985, 1989, 1991). The Aguja Formation has a maximum thickness of ∼ 300 m. Lower parts of the unit consist dominantly of marine to paralic sandstone and mudstone derived from source regions farther west. The upper shale member was deposited at the end of a major regression associated with final progradation of the strandline eastward across the region. The lower part of the member accumulated in deltaic coastal marsh and coastal flood-plain environments, whereas the upper part was deposited in an inland floodplain (Lehman, 1985, 1986). Abundant paleosol horizons within the upper part of the member record alternating humid and semi-arid conditions (Lehman, 1989; Atchley et al., 2004). In situ fossil tree trunks suggest that the floodplains at times supported tropical evergreen forests (Lehman and Wheeler, 2001).

The basaltic pyroclastic deposits described by Breyer et al. (2007) occur in a small, fault-bounded block ∼ 250 m across within a structurally complicated area on the privately owned Pitcock Rosillos Ranch, adjacent to the northern part of Big Bend National Park. Approximately 15 m of pyroclastic strata are exposed in the fault block and comprise interbedded pyroclastic fall and base-surge deposits containing a high proportion of terrigenous detritus intermixed with juvenile basaltic tephra. Impact sags occur beneath coarse basaltic pyroclasts and beneath blocks of sedimentary rock up to 1 m across derived from underlying strata. The pyroclastic deposits are inferred to have accumulated in a near-vent setting on the flanks of a small maar or tuff ring; the source vent is not exposed. The deposits are gradationally overlain by fossiliferous strata that appear to have accumulated in a small lake or pond and contain abundant turtle fossils. The assemblage of turtle genera present is found elsewhere in the Big Bend region only in the upper shale member of the Aguja Formation (Tomlinson, 1997).

3. Basaltic phreatomagmatic deposits at Peña Mountain

3.1. General features

Peña Mountain is a prominent ridge in the southwestern part of Big Bend National Park (Fig. 4), ∼ 40 km southwest of the...
pyroclastic deposits reported by Breyer et al. (2007). The ridge is capped by an Oligocene syenodiorite intrusion inferred by Carman (1994) to be a thick sill. The western part of the sill intrudes terrigenous strata within the upper shale member of the Aguja Formation. The eastern part of the sill intrudes previously undescribed basaltic pyroclastic deposits ≥ 70 m thick that are exposed on the northern and southern flanks of the mountain. Contact metamorphism is only noticeable within a few meters of the intrusive contact, and there is no evidence that emplacement of the sill caused significant deformation of the country rocks. The pyroclastic deposits conformably overlie typical inland floodplain facies in the upper part of the upper shale member.

These nonvolcanic fluvial strata include banded purple and grey overbank mudstones exhibiting paleosol horizons, with less common channel or sheet sandstones up to several meters thick. Lag deposits within the channels contain abundant pebbles of reworked pedogenic carbonate derived from erosion of paleosol horizons. On the eastern side of the mountain, the pyroclastic beds form an eastward-thinning tongue underlain and overlain by fluvial strata of the upper shale member.

The lower part of the pyroclastic succession on the northern side of Peña Mountain shows abnormally steep dips (up to 80°) and is separated from overlying, more gently dipping pyroclastic strata by a distinct truncation surface (Figs. 4–6).
3.2. Tephra composition and characteristics

Grain-size designations and terminology of pyroclastic deposits used here in general follow White and Houghton (2006), although we retain use of the term “lapillistone” for deposits consisting dominantly of lapilli (Fisher, 1961). We also follow White and Houghton (2006) in applying pyroclastic terminology to sedimentary detritus ejected explosively during phreatomagmatic eruptions. Additional descriptive details to those given here may be found in Befus (2006).

Pyroclastic deposits at Peña Mountain comprise interbedded lapilli tuff, lapillistone, and tuff. Most lapilli are ≤3 cm in length, but outsized basaltic and sedimentary clasts (described more fully below) make up to 15% of some beds. Alteration of the tephra generally imparts a light yellow-grey to grey-brown color to the deposits. Armored lapilli are present, in which basaltic or sedimentary pyroclasts are coated with a layer of ash several millimeters thick (Fig. 7A), but accretionary lapilli have not been observed. Fluidal basalt bombs and angular, subsequent juvenile clasts of basalt are up to 1 m in length (Fig. 7B). Bombs contain ≤30% vesicles and in some cases show cauliflower-type margins, typical of bombs produced from phreatomagmatic eruptions (Lorenz, 1973).

In thin section, the basalt contains euhedral olivine phenocrysts 1–2 mm in length, which are almost completely replaced by carbonate and yellow-brown (smectitic?) clay. Coarser basaltic clasts generally have an interstitial or pilotaxitic groundmass, with plagioclase microlites partly replaced by clay ± carbonate. Basaltic ash and finer lapilli (≤3 cm) typically contain abundant glass, which is altered to extremely fine-grained clay and zeolites (?) but retains the light-brown, transparent appearance of sideromelane. Coarser grained carbonate also replaces the glass in places. Palagonite, a typical initial alteration product of sideromelane (e.g., Hay and Iijima, 1968), has been recognized only in minor amounts but may have been obscured by subsequent alteration.

Basaltic ash and finer lapilli containing altered sideromelane have fluidal to blocky, angular shapes (Fig. 8), with margins of angular particles being defined by curvilinear fractures. Vesicularity is typically in the range 0–30%, but sparse scoriaceous clasts are intermixed with poorly vesicular pyroclasts in some beds. Angular particles of poorly vesicular tachylite, in which plagioclase microlites are contained in a mesostasis of nearly opaque glass (Fig. 8D), are present in minor amounts (<2–3%).

The basaltic tephra is intermixed with abundant accidental sedimentary detritus, including angular to subrounded, medium sand- to silt-sized monocrystalline quartz, plagioclase, and orthoclase grains (Fig. 8) and medium to fine sand-sized lithic grains of felsic and intermediate volcanic rock; grains of polycrystalline quartz and mica-quartz schist are present in minor amounts. Lapillistones typically contain ≤15% of these accidental grains. Lapilli tuffs and tuffs contain up to ~80% accidental grains, and in many of these beds the coarser pyroclasts are set within an altered matrix comprising fine to very fine basaltic ash intermixed with a high proportion of terrigenous mud (Fig. 8A). We have not attempted to estimate
the percentage of sediment in this finer fraction because of the difficulty in discerning terrigenous mud from altered basaltic ash in thin section. Sand grains of quartz and felsic volcanic rock also commonly occur as xenocrysts and xenoliths within the basaltic pyroclasts (Fig. 8B).

Angular to subrounded clasts ≤1 m across of light-grey mudstone (Fig. 9A, B) and grey to brown sandstone and conglomerate are common within the succession. The conglomerate contains reworked pedogenic carbonate nodules, and some distinctive sandstone blocks contain abundant oyster shells. Some mudstone clasts were only partly lithified when they were erupted and were flattened and disrupted upon impact, with partly detached portions being connected by thin, plastically deformed necks. One unusually large slab of mudstone 3 m long contains abundant subangular to fluidal bodies of basalt ≤10 cm in length. The fluidal basalt bodies have highly irregular margins and show quench fragmentation and fine-scale peperitic intermixing with the mud in the ejected slab (Fig. 9C).

Chemical analyses of three basalt bombs are given in Table 1. Sample B4 was collected from the lower pyroclastic sequence below the truncation surface on the northern side of Peña Mountain. Samples C29 and C30 come from a single bomb-rich layer on the southern side of the mountain (Fig. 4). A single available analysis for the Pitcock Rosillos Ranch basaltic deposits is also shown for comparison. The major-element analyses are difficult to interpret because of the high degree of alteration, which is reflected in the high CaO and LOI values. The samples show significant scatter on a total alkalis versus silica diagram (Fig. 10), but the three Peña Mountain samples cluster tightly in the field for subalkaline basalt in the Zr/TiO₂ versus Nb/Y diagram. The latter result is considered more reliable because it is based on trace elements that are resistant to disturbance during low-temperature alteration (Winchester and Floyd, 1977). The sample from the Pitcock Rosillos Ranch, in contrast, plots in the field for alkaline basalt.

3.2.1. Interpretation

The general low vesicularity of the basaltic tephra indicates that explosive release of magmatic volatiles played only a limited role in eruptive behavior and that much of the impetus for explosions came from conversion of external water to steam. Abundant sideromelane in the juvenile pyroclasts is consistent with rapid quenching of magma in contact with water during phreatomagmatic interactions (e.g., Fisher and Schmincke, 1984)

Fig. 6. View looking southwest at truncation surface (dashed) separating pyroclastic sequences on the northern side of Peña Mountain.

Fig. 7. A. Lapilli-tuff and lapillistone layers. AL = armored lapillus in lapilli-tuff bed. Pencil points to erosional scour at base of bed, which cuts into underlying tuff and is filled with coarse lag deposit. Openwork lapillistone is visible in lower part of view. B. Irregular basaltic bomb, which fractured upon impact. Arrow points to pencil for scale. Bomb is exposed on bedding plane within lapilli tuff; loose pieces weathered from bomb are visible in foreground.
Sparse tachylite grains are inferred to have been derived from magma batches that had cooled more slowly in the conduit prior to explosive disruption. Intermixture of sideromelane and tachylite clasts, as well as the wide range in vesicularity shown by pyroclasts within a single deposit, probably results from explosive disruption of heterogeneous batches of magma in the conduit, coupled with recycling of clasts during repeated eruptions (Houghton and Smith, 1993).

Abundant sand- and mud-sized terrigenous detritus intermixed with the juvenile basaltic tephra, as well as the large number of coarser accidental sedimentary clasts, suggest that phreatomagmatic explosions were generated when rising magma encountered groundwater-rich strata. Many of the mudstone clasts show evidence of being poorly consolidated at the time of eruption. Xenocrysts of sand grains within the basaltic pyroclasts imply intimate, fine-scale magma-sediment interaction prior to or during eruption. Similar xenocrysts within basaltic pyroclasts are known from other phreatomagmatic settings where magma has interacted extensively with unlithified sands (e.g., Hanson and Elliot, 1996; Lorenz et al., 2002; Ross and White, 2006). Peperitic textures within one large mudstone block are inferred to record precursory stages in magma/wet-sediment interaction in the subsurface that led to explosive phreatomagmatism (cf. White, 1991, 1996).

3.3. Depositional features

Pyroclastic deposits at Peña Mountain consist dominantly of tabular beds of lapilli tuff and lapillistone ≤30 cm thick that are laterally continuous on the scale of individual outcrops (Fig. 11A). Representative measured sections are shown in Figs. 5, 12, and 13. Lapillistones are relatively well sorted, with openwork texture (Fig. 7A), and are cemented by carbonate. Lapilli tuffs either lack finer ash particles and show openwork texture or contain abundant fine ash and are matrix-supported. Out-sized clasts in some cases show well-defined impact sags (Figs. 9B, 11B), indicating emplacement along ballistic trajectories.

Lapillistone and lapilli-tuff beds are generally either massive and structureless (Fig. 9B) or show normal or inverse grading. Some beds exhibit internal planar bedding defined by diffuse changes in particle size or proportion of matrix. Tabular lapilli occur in otherwise massive parts of beds. Some beds have basal erosional scour features up to 10 cm deep (Figs. 7A, 12). Low-angle cross-stratification or undulatory or wavy bedding are present in a small percentage (<10%) of beds (Figs. 12 and 14), particularly in their upper parts, and record development of broad, low-amplitude bedforms. Dips on stoss and lee sides of these
dune-like structures are typically ≤15°. Crests are smooth and rounded, and are draped by overlying layers. Individual bedforms show progressive upward growth from planar bases.

Interbeds of tuff 1–25 cm thick commonly separate the coarser layers (Fig. 9B) and are massive or show diffuse planar bedding and lamination, with less common low-angle cross-stratification. Some tuff interbeds show swirled or contorted laminae recording soft-sediment deformation of cohesive ash not obviously related to impact sags.

The lower part of the succession on the southern side of Peña Mountain contains unusually thick (4–9 m), massive to diffusely planar-bedded lapillistone and lapilli-tuff layers that exhibit basal erosional scour features (Fig. 13). Basaltic bombs and blocks and accidental blocks of sandstone and mudstone are abundant in these layers and include the 3-m-long clast of mudstone shown in Fig. 9C.

A fossil trunk of an araucarioid conifer 1.2 m tall (Fig. 15) is enclosed by the stratigraphically lowest pyroclastic beds on the northwestern side of the mountain (locality PM 1.2 in Wheeler and Lehman, 2005). The trunk shows excellent preservation of microscopic textures (see Plate I-D in Wheeler and Lehman, 2005), with no evidence of carbonization. Its roots are not visible, but the lowest exposed portion of the trunk is ~15 cm above the contact with Aguja floodplain mudstone. When the enclosing beds are rotated back to horizontal, the trunk is vertical. A fragment of wood lies on a bedding plane within the enclosing pyroclastic strata (Fig. 15), but it is unclear whether this fragment was derived from the larger trunk. Sparse, noncarbonized fossil wood fragments and parts of logs ≤30 cm long are also present in other parts of the pyroclastic succession on both the northern and southern sides of the mountain.

3.3.1. Interpretation

The range of depositional features is consistent with emplacement of tephra from base surges and by direct fallout from eruption columns during pulsatory phreatomagmatic eruptions. Erosional scour features, undulatory bedding, and/or low-angle cross-stratification in some lapillistone and lapilli-tuff beds indicate lateral transport of tephra by turbulent, decelerating base surges (Moore, 1967; Fisher and Waters, 1970; Waters and Fisher, 1971; Crowe and Fisher, 1973; Druitt, 1998; Wohletz, 1998). Zones of inverse grading within some beds suggest development of high-particle-concentration traction carpets in the basal parts of surges (Sohn, 1997). Beds exhibiting typical features of base-surge deposits include those enclosing the upright fossil conifer trunk shown in Fig. 15. We infer that the tree was killed by relatively weak, low-temperature surges that
were unable to knock it down. There is no evidence that ash was plastered against the tree, but such a feature might be difficult to recognize in ancient, lithified deposits.

Little or no reworking of tephra appears to have occurred by wind or running water after initial deposition. The geometry of preserved bedforms is consistent with tractional sedimentation under conditions of high lateral bed shear stress at the base of rapidly moving pyroclastic density currents, and contrasts markedly with tractional structures found in typical eolian or fluvial deposits (e.g., Waters and Fisher, 1971; Smith and Katzman, 1991; Valentine and Fisher, 2000).

Tabular, laterally continuous, massive to diffusely planar-bedded lapillistone and lapilli-tuff beds lacking basal scour features or tractional bedforms either are of pyroclastic fall origin or were deposited from high-concentration, deflating base surges in which rapid suspension sedimentation prevented tractional reworking of particles. Base-surge deposits of this type are well known in phreatomagmatic settings (Sohn and Chough, 1989; Chough and Sohn, 1990; Lajoie et al., 1992; Allen et al., 1996; Colella and Hiscott, 1997). The unusually thick, massive to diffusely bedded, bomb- and block-rich units in the lower part of the section shown in Fig. 13 must have been emplaced, at least in part, from turbulent pyroclastic density currents, as shown by the presence of erosional scour features at the bases of the units. These beds are inferred to have been deposited from heavily overladen base surges close to the vent, probably combined with ballistic fallout of tephra. The high proportion of blocks and bombs makes these beds similar to the explosion breccias described by Wohletz and Sheridan (1983), which are interpreted to record initial conduit-clearing explosive eruptions during construction of phreatomagmatic volcanoes. Analogous deposits have also been described in the literature from other near-vent phreatomagmatic sequences (e.g., Sohn and Chough, 1989; White, 1991; Németh and White, 2003).

Sorting in phreatomagmatic deposits is controlled in part by the wetness of the tephra during eruption and deposition (e.g., Self et al., 1980; Ross, 1986; Sohn and Chough, 1989; Valentine and Fisher, 2000). Fines-depleted beds with openwork textures in the Peña Mountain succession are inferred to have been produced from high-temperature parts of eruption columns containing dry, superheated steam, which allowed efficient winnowing of finer particles during deposition by either pyroclastic fall or base-surge processes (e.g., Druitt, 1992; Colella and Hiscott, 1997). Cohesion of wet tephra emplaced at temperatures below the boiling point of water prevents efficient sorting and likely explains formation of ash-rich, matrix-supported lapilli-tuff beds within the succession (cf. Allen et al., 1996). That some of the tephra was moist during or after deposition is proven by the occurrence of armored lapilli, soft-sediment deformation of fine-grained layers, and numerous impact sags within the succession. The presence of noncarbonized wood is also consistent with low emplacement temperatures for some deposits. These inferred variations in temperature and wetness between different units within the succession imply fluctuations in the proportions of magma and water (or wet

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### Table 1
Chemical analyses of basaltic bombs from Peña Mountain (B4, C29, and C30) and basaltic block from Pitcock Rosillos Ranch (A19c)

<table>
<thead>
<tr>
<th>Sample</th>
<th>B4</th>
<th>C29</th>
<th>C30</th>
<th>A19c</th>
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<tr>
<td>SiO2</td>
<td>46.61</td>
<td>48.65</td>
<td>54.77</td>
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<td>TiO2</td>
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<td>1.55</td>
<td>1.61</td>
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<td>18.15</td>
<td>20.93</td>
<td>17.66</td>
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<tr>
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<td>7.30</td>
<td>2.59</td>
<td>3.84</td>
<td>14.29</td>
</tr>
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<td>MnO</td>
<td>0.13</td>
<td>0.10</td>
<td>0.10</td>
<td>0.24</td>
</tr>
<tr>
<td>MgO</td>
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<td>1.82</td>
<td>3.33</td>
<td>4.83</td>
</tr>
<tr>
<td>CaO</td>
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<td>19.29</td>
<td>7.53</td>
<td>10.16</td>
</tr>
<tr>
<td>Na2O</td>
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<td>5.71</td>
<td>5.12</td>
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<td>0.06</td>
</tr>
<tr>
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<td>1.50</td>
<td>0.64</td>
<td>0.64</td>
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<td>LOI</td>
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<td>12.50</td>
<td>5.93</td>
<td>9.76</td>
</tr>
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<td>Sum*</td>
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<td>92.31</td>
<td>86.64</td>
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<td>79</td>
<td>107</td>
<td>178</td>
</tr>
<tr>
<td>Nb</td>
<td>10</td>
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<td>11</td>
<td>35</td>
</tr>
<tr>
<td>Y</td>
<td>56</td>
<td>36</td>
<td>36</td>
<td>24</td>
</tr>
</tbody>
</table>

Major elements in wt.% (normalized to 100% on volatile-free basis), trace elements in ppm. Major elements analyzed by XRF on fused glass beads with lithium tetraborate flux. Trace elements analyzed by ICP-MS. Analyses were carried out at the Washington State University Geoanalytical Laboratory, Pullman, Washington.

* Total before normalization.

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* Fig. 10. Analyses of basaltic pyroclasts plotted on standard diagrams. A. Total alkalis versus silica diagram of Le Bas et al. (1986); Bta=basaltic trachyandesite; TB=tachyrite and basanite; Trb=trachybasalt. B. Zr/TiO2 versus Nb/Y diagram of Winchester and Floyd (1977), revised by Pearce (1996).
sediment) undergoing explosive interactions in the conduit (Sheridan and Wohletz, 1983; Sohn, 1996; White, 1996).

Accretionary lapilli, which are common in many phreatomagmatic deposits (Schumacher and Schmincke, 1991), are lacking in the present case. Reasons for their absence are unclear. In cases where the tephra is inferred to have been moist during transport and deposition, excess water may have prevented formation or preservation of well-defined accretionary lapilli (e.g., Walker, 1981; Ross, 1986; Schumacher and Schmincke, 1995; Martin and Németh, 2005). Leat and Thompson (1988) have suggested that a high proportion of fine-grained accidental sedimentary detritus ejected during phreatomagmatic eruptions may also inhibit formation of accretionary lapilli.

3.4. Structural relations

The well-defined truncation surface on the northern side of Peña Mountain separates the pyroclastic strata into two different sequences with similar depositional features but different structural geometry. Bedding attitudes beneath the truncation surface vary markedly over distances of tens to hundreds of meters, with dips ranging up to 80° (Fig. 4). The structural complexity is partly caused by large folds with irregular trends, and partly by variable degrees of offset along high-angle syndepositional normal faults. In places, these faults define small grabens a few meters across (Fig. 5). The amount of offset decreases upward along individual faults, some of which are filled by zones of disaggregated pyroclastic material that lacks evidence of brittle cataclasis and is inferred to have been injected in a soft-sediment state. All of these faults terminate against the truncation surface (Fig. 16). A longer fault cutting the western part of the succession on the northern side of the mountain (Fig. 4) displaces the truncation surface and may be unrelated to the Cretaceous volcanism.

The base of the upper sequence above the truncation surface is partly defined by the laterally discontinuous lacustrine unit shown in Fig. 5. Where that unit is lacking, the lowest pyroclastic beds in the upper sequence lie on the tilted lower sequence with marked angular discordance (Fig. 6); there is no evidence for faulting along the contact. Strata within the upper sequence dip relatively gently to the southeast, consistent with regional attitudes of Upper Cretaceous strata in the area.

On the southern side of Peña Mountain, a zone up to 60 m across of intense structural disruption of original bedding separates less disturbed pyroclastic beds from typical non-volcanic Aguja strata to the east and south. The disrupted zone has an overall arcuate trend (Fig. 17A), but its margins are locally highly irregular (Fig. 17B). The outer part of the disrupted zone affects nonvolcanic Aguja sandstones and mudstones, in which individual beds generally can only be traced a few meters and are deformed by numerous soft-sediment faults and small-scale folds. Locally, more resistant sandstone units can be traced for tens of meters within the disrupted zone and help to define its overall structure. Bedding in one such unit progressively steepens and then becomes overturned toward the contact with the pyroclastic rocks (Fig. 17A).

The inner part of the disrupted zone consists of pyroclastic strata that show comparable depositional features to other parts of the pyroclastic succession but are intensely deformed in the same manner as the adjacent nonvolcanic Aguja strata. The contact between the nonvolcanic strata and the disrupted pyroclastic beds is defined in places by a subvertical, soft-sediment fault zone that abruptly truncates bedding on both sides. Sandstone beds are dragged down into the fault zone and form detached, meter-scale masses strung out along it. Masses of nonvolcanic Aguja strata are also enclosed by disrupted pyroclastic strata inward from the fault. In some cases, steeply tilted pyroclastic layers rest depositionally on more strongly deformed nonvolcanic strata, indicating that some disruption of Aguja strata occurred prior to deposition of the pyroclastic beds.

The zone of disrupted pyroclastic beds is in abrupt contact with a sequence ≥65 m thick of less disturbed pyroclastic strata (Fig. 13). Bedding trends within the less disturbed pyroclastic sequence generally are subparallel to the contact with the disrupted zone but in places are truncated by it (Fig. 17A, B). Locally near the contact, the less disturbed strata dip into the disrupted zone, but they typically dip 25–60° to the northwest, with dips tending to increase upward in the sequence.
3.4.1. Interpretation

An interpretative cross section of the pyroclastic deposits exposed at Peña Mountain is given in Fig. 18. Detailed interpretation is rendered difficult by the fact that only small parts of the succession are preserved, whereas other parts are obscured by the younger syenodiorite sill. We have therefore not attempted to calculate original volumes of the deposits or source vents.

An early stage in the development of the succession is recorded by the sequence beneath the truncation surface on the northern side of Peña Mountain. The structural complexity within this sequence is not the result of tectonic deformation because neither the pyroclastic strata above the truncation surface nor laterally equivalent, nonvolcanic parts of the Aguja Formation show comparable effects. We infer that the sequence underwent tilting and soft-sediment deformation during slumping of unstable rim deposits into a developing crater to the south. Similar truncation surfaces resulting from collapse of parts of small volcanoes have been described by Sohn and Park (2005) in Pleistocene tuff rings and tuff cones in Korea.
On the southern side of the mountain, the disrupted zone separating the pyroclastic deposits from typical nonvolcanic Aguja strata to the southeast and south (Fig. 17A, B) is inferred to represent a partly preserved crater margin. Collapse of the crater margin caused disruption of poorly consolidated Aguja strata and adjacent pyroclastic beds. Piecemeal collapse is recorded by highly irregular, subvertical contacts within parts of the disrupted zone (Fig. 17B). Less disturbed pyroclastic strata in contact with the disrupted zone accumulated at higher levels and underwent tilting during downward subsidence as the crater expanded laterally.

4. Age of the pyroclastic deposits

The lower part of the upper shale member of the Aguja Formation contains a diverse assemblage of fossil mammals and reptiles (including dinosaurs) that indicate a Late Campanian age (Rowe et al., 1992; Sankey, 2001). Fossils are generally less abundant and diagnostic in the upper part of the upper shale member (Lehman, 1985), where the pyroclastic deposits described herein occur, and that part of the member could extend into the Early Maastrichtian. The overlying Javelina Formation contains Maastrichtian vertebrate fossils (Lehman, 1985; Lehman and Coulson, 2002), and a distal felsic tuff present within the middle of that unit and inferred to be derived from the Cordilleran arc farther west has yielded a U–Pb monazite age of 69.0±0.9 Ma (Lehman et al., 2006).

In order to complement the existing paleontological constraints on the age of the upper shale member, U–Pb isotopic analyses were conducted on zircon separated from the same three basaltic bombs within the Peña Mountain pyroclastic succession that yielded the chemical analyses shown in Table 1. Zircon was extracted from the samples using standard crushing and mineral-separation techniques at the University of Texas, Dallas. Hand-picked zircon grains were analyzed using the sensitive high mass resolution ion microprobe with reverse geometry (SHRIMP-RG) operated by the U.S. Geological Survey and Stanford University. Analytical techniques are given in Appendix A, and results are shown in Table 2.

Analyzed zircon grains in all three samples are typically euhedral, with dipymidal terminations and concentric zonation typical of igneous zircon; inherited cores were not observed (Fig. 19). Most grains yielded concordant or nearly concordant analyses representing a wide range in crystallization ages, in some cases >1.0 Ga (Fig. 20A, C, E). These old grains are clearly xenocrysts. The two youngest concordant analyses for sample B4 (Fig. 20B) have overlapping 2σ error ellipses and yield a weighted mean age of 77.9±1.4 Ma, within error of the age for the youngest concordant analysis (76.8±2.0 Ma) for sample C29 (Fig. 20D). The four youngest concordant analyses for sample C30 also have overlapping 2σ error ellipses and yield a weighted mean age of 76.3±2.1 Ma. An older population of concordant, overlapping analyses from the same sample (Fig. 20F) yields a weighted mean age of 96.7±1.3 Ma.

Interpretation of the age results for the analyzed zircon grains is rendered problematic by the abundance of xenocrystic grains of various ages in the samples, which have similar Th/U ratios and appear similar under cathodoluminescence to the grains yielding the youngest concordant analyses. The sources of these xenocrystic zircon grains are unclear. Older components in the zircon population may have been incorporated into the basaltic magma during its transit through Precambrian and Paleozoic crust inferred to underlie the region at depth (e.g., James and Henry, 1993). An alternate possibility is that at least some of the xenocrystic grains were incorporated during interaction between basaltic magma and Cretaceous terrigenous sediment that preceded or accompanied explosive phreatomagmatic eruptions.

![Fig. 14. Lapillistone, lapilli-tuff, and tuff layers near location of fossil conifer trunk on northwestern side of Peña Mountain. Massive lapillistone beds in upper part of view are separated by thin tuff layers. Lapilli tuff and tuff in middle part of view show low-angle cross-stratification and wavy bedding. Low-angle cross-stratification (dashed) is also visible in lapillistone at base of view. Pencil (arrowed) for scale.](image1)

![Fig. 15. Fossil conifer trunk (labeled) within pyroclastic strata on northwestern side of Peña Mountain (Fig. 4). Head of hammer to right of trunk is parallel to bedding. A fragment of wood lying on a bedding plane is also labeled.](image2)
The youngest concordant analyses for all three Peña Mountain samples have overlapping $2\sigma$ error limits and form a distinct population separate from the older analyses. When considered together, the youngest concordant analyses yield a weighted mean age of 76.9±1.2 Ma (Fig. 21). One interpretation is that these youngest grains are also xenocrysts, which would require the Peña Mountain basaltic pyroclastic deposits to be younger than ~77 Ma. Another interpretation, which we prefer, is that the weighted mean age for the distinct population of youngest zircon grains represents the crystallization age of the basalt.

U–Pb isotopic analyses obtained in the same laboratory were reported by Breyer et al. (2007) for zircon extracted from a juvenile basaltic block in the Cretaceous pyroclastic sequence on the Pitcock Rosillos Ranch. These data also revealed the presence of numerous xenocrysts. Accidental blocks of fine-grained felsic crystal-vitric tuff contained within the basaltic deposits are one possible source for the zircon xenocrysts. The youngest concordant analyses from the sample yielded a weighted mean age of 72.6±1.5 Ma, which was inferred to represent the crystallization age of the basalt. Both this result and our new data for the Peña Mountain deposits are consistent with the available palaeontological constraints and with the time scale of Ogg et al. (2004), who place the upper boundary of the Campanian at 70.6±0.6 Ma. Considered together, the palaeontological and geochronological evidence indicates that basaltic volcanism at Peña Mountain and farther to the northeast in the area of the Pitcock Rosillos Ranch occurred in a relatively narrow time frame within the Late Cretaceous.

5. Discussion

5.1. Role of subsurface phreatomagmatic explosive activity

Textural characteristics of the tephra in the Peña Mountain pyroclastic deposits indicate interaction between basaltic magma and external water (or water-rich sediments) during explosive volcanism. There is no sedimentological evidence for large bodies of standing water in this part of the Big Bend region at the onset of Cretaceous phreatomagmatic activity. As in many other documented examples of explosive phreatomagmatic volcanism in nonmarine settings, the most likely source for the required volumes of water appears to be groundwater within underlying units. The high proportion of accidental sedimentary detritus in the form of loose sand grains and abundant mud intermixed with the Peña Mountain pyroclastic deposits points to generation of explosions within unlithified, groundwater-rich sediments at shallow levels in the subsurface (e.g., Leat and Thompson, 1988; White, 1991; Hanson and Elliot, 1996; Lorenz, 2003; McClintock and White, 2006; Ross and White, 2006). Types and relative proportions of these accidental grains are similar to those found in sandstones in underlying parts of the Aguja Formation, which contain a high proportion of detritus derived from the Cretaceous volcanic arc farther west (Lehman, 1985, 1991; Bohanan, 1987). Most coarser accidental sedimentary clasts in the Peña Mountain pyroclastic deposits can be matched with lithologies within floodplain deposits of the upper shale member, and it is likely that subsurface phreatomagmatic explosions were initiated at shallow levels in that unit. The upper shale member is composed dominantly of impermeable mudstones, with a subordinate amount of sandstones that could have provided adequate flow of groundwater. Following White (1991, 1996), we infer that water-rich mud rather than free water acted as a major coolant during explosive subsurface fuel-coolant interactions. Peperitic textures in the erupted mudstone slab shown in Fig. 9C document initial stages in these processes.

The base of the upper shale member is exposed ~1.5 km west of Peña Mountain, and stratigraphic sections given in Lehman (1985) indicate that the nonvolcanic part of the member beneath the pyroclastic succession is 100–140 m thick.
in the area of concern here. This estimate provides some indication of possible depths at which subsurface phreatomagmatic explosions occurred, although a more rigorous estimate would require correcting present stratigraphic thicknesses for compaction. Distinctive sandstone blocks rich in oyster fossils must have been derived from the marine Rattlesnake Mountain Sandstone Member of the Aguja Formation, which is the only unit within the Cretaceous stratigraphic sequence in the region that contains large numbers of oyster fossils (Lehman, 1985). This unit is estimated to occur ∼20 m beneath the base of the upper shale member in the Peña Mountain area, based on projection of thickness trends in the Aguja Formation. The locus of explosive activity may have migrated downward as groundwater was exhausted at shallow depths, as is inferred for many subaerial phreatomagmatic volcanoes (Lorenz, 1986). However, we have not observed consistent vertical variations in composition of accidental sedimentary debris within the Peña Mountain pyroclastic succession that would provide direct support for such a model.

5.2. Nature and evolution of source vents

Abundant coarse ballistic clasts within the Peña Mountain pyroclastic deposits indicate accumulation in near-vent settings, which is consistent with the general scarcity of tractional current structures in base-surge deposits within the succession. Several studies of lateral facies relations in outflow surge deposits from phreatomagmatic volcanoes have shown that massive beds are common in proximal settings (typically within a few hundred meters of the vent) and transform downcurrent to beds showing increasing evidence for tractional sedimentation, recording progressive dilution of the surges and decrease in rate of fallout of particles from suspension (e.g., Sohn and Chough, 1989; Chough and Sohn, 1990; Lajoie et al., 1992; Vazquez and Ort, 2006). Preservation of the upright fossil conifer trunk within surge deposits on the northwestern side of Peña Mountain is somewhat surprising given the proximal setting and suggests rapid lateral variations in surge energy during pulsatory phreatomagmatic explosions.

The gross morphology and internal depositional features of the pyroclastic sequences above and below the truncation surface on the northern side of Peña Mountain are consistent with deposition in overlapping tephra aprons rimming tuff rings or maars, the most common types of subaerial monogenetic phreatomagmatic volcanoes (Wohletz and Sheridan, 1983; Cas and Wright, 1987). Field relations on the southern side of the mountain (Fig. 17A, B) reveal the presence of a typical maar volcano, in which the crater floor was below the ground surface and was excavated into pre-existing strata (Lorenz, 1973, 1986), with downward subsidence of rim deposits into the developing crater.

The Peña Mountain pyroclastic deposits provide a case study for recognition of ancient, partly preserved maar sequences in the stratigraphic record and illustrate the structural complications that may result from synvolcanic collapse or subsidence of parts of the volcanoes. Similar collapse of small phreatomagmatic volcanoes has been well documented in other studies (e.g., Gutmann, 1976; Sohn and Park, 2005) and may result from instability of the substrate, coupled with removal of support due to lateral expansion of crater margins. Sohn and
Park (2005) have pointed out that such collapse processes are likely to be common where the substrate consists of poorly consolidated, mechanically weak sediments. In the present case, volcano collapse appears to have resulted from construction of Late Cretaceous maar-type vent complexes within and upon unconsolidated floodplain sediments in the upper shale member of the Aguja Formation.

One possible scenario for the evolution of the Peña Mountain pyroclastic succession is shown in Fig. 22, in which initial formation of a maar or tuff ring was followed by collapse of the crater margins and large-scale slumping to generate the deformed sequence beneath the truncation surface on the northern side of Peña Mountain. Erosion of the slumped beds was followed by deposition of the overlying lacustrine unit during a period of volcanic quiescence. The exposed succession to the south represents a higher structural level than the truncation surface to the north and is inferred to record lateral migration of the conduit during an episode of renewed explosive volcanism. Subsidence of pyroclastic deposits formed during this second episode juxtaposed them with disrupted strata along the new crater margin. Pyroclastic beds overlying the truncation surface on the northern side of the mountain may equate with part of the succession exposed to the south, as implied in Fig. 22, but could also have originated from another nearby vent.

5.3. Regional considerations

Upper Cretaceous near-vent basaltic successions erupted from small phreatomagmatic volcanoes have now been documented from two separate localities in the Big Bend region of West Texas. Although U–Pb zircon geochronological data for the deposits can be interpreted in various ways, it should be stressed that the available paleontological evidence indicates a Late Campanian (to possibly Early Maastrichtian) age for the pyroclastic deposits at both localities. Monogenetic volcanoes of the type described here commonly occur in clusters (Connor and Conway, 2000), making it likely that additional examples of the same age remain to be discovered in the region.

Our new data from Peña Mountain are consistent with the suggestion of Breyer et al. (2007) that Cretaceous basaltic pyroclastic deposits in the Big Bend region represent a western extension of the intraplate Balcones igneous province. As presently defined in the literature (Barker et al., 1987; Byerly, 1991; Ewing, 2004), the Balcones province comprises several hundred individual small volcanoes and/or associated intrusions, which occur both in outcrop and in the subsurface and are aligned along the frontal zone of the largely buried, Late Paleozoic Ouachita orogenic belt (Fig. 2). Rock compositions are dominantly melilite nephelinite and nephelinite, with less abundant alkali basalt, phonolite, and basanite (Spencer, 1969; Barker et al., 1987). Different volcanic centers or intrusions have yielded $^{40}$Ar/$^{39}$Ar ages of $\sim$82–72 Ma (Miggins et al., 2004; Griffin et al., 2005), as well as a more limited number of U–Pb zircon ages of $\sim$86–77 Ma (Griffin et al., 2005). The overall age range of the province is best constrained by stratigraphic relations between the volcanoes and adjacent fossiliferous Upper Cretaceous carbonate units. This biostratigraphic evidence shows that igneous activity spanned the Cenomanian through Maastrichtian (i.e., much of the Late Cretaceous), with the largest percentage of the magmatism occurring in the Campanian (Byerly, 1991; Ewing, 2004). The similarity in timing between the peak in Balcones igneous activity and basaltic volcanism in the Big Bend region farther to the west suggests a direct relation.

In the main part of the Balcones province, volcanism occurred in marine environments and constructed subaqueous to emergent Surtseyan tuff cones, in which explosive phreatomagmatism was driven by interaction of ultramafic and mafic
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*Error correlation coefficient.
magnas with sea water (Ewing and Caran, 1982; Ewing, 2004). The style of volcanism is different in the inferred extension of the province in West Texas, where the vents were located in low-relief floodplain settings far west of the Late Cretaceous shoreline (Fig. 2). This environment favored formation of maars-type volcanoes fed by explosive interactions between uprising basaltic magma and groundwater-rich sediment in the shallow subsurface. The resulting pyroclastic deposits are characterized by intermixture of juvenile basaltic pyroclasts with accidental terrigenous debris derived from underlying parts of the Cretaceous sequence.

Based on limited trace-element data presented herein, the Peña Mountain subalkaline basalts are compositionally distinct from the alkaline basaltic deposits present on the Pitcock Rosillos Ranch. They also contrast markedly with the main part of the Balcones province to the east, where only alkaline compositions are known and melilite nephelinites are abundant (Spencer, 1969; Barker et al., 1987). More detailed geochemical studies are needed on the West Texas Cretaceous basalts, which could provide useful insights into along-strike variations in magma compositions and petrogenesis within the Balcones province.

Acknowledgments

We thank Roy Pitcock for allowing the first two authors to stay on his ranch during part of the field work. We are indebted to Joe Wooden for assistance with acquisition and interpretation of the U–Pb zircon geochronological data. We also thank Trey Hargrove for his help in working with the zircon images, and John Breyer for assistance with parts of the field work. Art Busbey gave valuable help in figure preparation. Permission to carry out field work in Big Bend National Park was granted to the first author by the U.S. Department of the Interior. The manuscript benefited from thorough reviews by Ian Skilling and an anonymous reviewer.

Appendix A

U–Pb zircon isotopic analyses were performed in 2005 by W.R. Griffin at the SHRIMP-RG facility co-operated by the U.S. Geological Survey and Stanford University. Zircon grains were mounted in epoxy, ground and polished to a 1 µm finish to reveal quasi-equatorial sections through the grains, and imaged via cathodoluminescence on a JEOL 5600LLV scanning electron microscope and via transmitted light on an optical microscope prior to and following analysis. Grain mounts were washed with a saturated EDTA solution, rinsed in distilled water, dried in a vacuum oven, and coated with gold. The primary ion beam of the SHRIMP-RG instrument was first rastered for 120 s over the analytical spot to remove a small region of the gold coat and any surface contamination. The beam was then focused for an analysis time of 12 min, which typically produced an ablation spot 20–40 µm wide and 1–2 µm deep. Secondary ions were generated from the target spot with an O²⁻ primary ion beam at 4–6 nA. Nine peaks were measured sequentially for each zircon spot. Concentration data for unknown zircon grains were calibrated against the R33 zircon standard [419 Ma, quartz diorite of Braintree Complex, Vermont (Black et al., 2004)].

Data reduction followed the methods described by Williams (1997) and Ireland and Williams (2003) and was performed by J. Wooden at the USGS/Stanford lab using SQUID 1.02 software (Ludwig, 2001). Correction for common lead was performed by fitting a model age for the sample, based on the uncorrected data, to the lead evolution model of Stacey and Kramers (1975) and applying that to the uncorrected data. Analysis points were plotted and ages calculated using Isoplot/Ex 3.0 (Ludwig, 2000).

References


Fig. 20. U–Pb zircon isotopic analyses plotted on standard concordia diagrams (A, C, E) and Tera-Wasserburg diagrams (B, D, F). Analyses are shown as 2σ error ellipses or as solid dots where ellipse is too small to show at scale of diagram. Labels for individual analyses correspond to those in Table 2. Ages on concordia are shown in Ma. A. Analyses yielding ages >300 Ma for sample B4. B. Analyses yielding youngest ages for B4. C. Analyses yielding ages >400 Ma for sample C29. D. Analyses yielding youngest ages for C29. E. Analyses yielding ages >140 Ma for sample C30. F. Analyses yielding youngest ages for C30. MSWD = mean square of weighted deviates.


