Field excursions to the Jemez Mountains, New Mexico

by Stephen Self, Grant Heiken, Martha L. Sykes, Kenneth Wohletz, Richard V. Fisher, and David P. Dethier

BULLETIN 134 New Mexico Bureau of Mines and Mineral Resources 1996

A DIVISION OF

NEW MEXICO INSTITUTE OF MINING & TECHNOLOGY

Part II: Field guide to the maar volcanoes of White Rock Canyon

Acknowledgments

This study was supported by the University of California Institute of Geophysics and Planetary Physics, Los Alamos Branch. Information on the hydrology of the region came from hydrologist and geologist Bill Purtymun, Los Alamos National Laboratory, who also helped us with access to the lower reaches of White Rock Canyon.

Grant Heiken and Kenneth Wohletz
Los Alamos, New Mexico 87545

Richard V.Fisher
Santa Barbara, California 93106

David P. Dethier
Williamstown, Massachusetts 01267

Introduction

Rift volcanism is often characterized by the complex interplay between volcanic and geomorphic processes. By their very nature, rifts are the loci for basaltic volcanism and often host the continental depressions that serve also as groundwater basins, major rivers, and lakes. Eruption processes are controlled by the depths at which rising magma intercepts either ground or surface water; in some instances, modern or ancient aquifers can be mapped by identifying the degree of phreatomagmatic activity that characterizes cones or tuff rings within a rift-bound volcanic field.

The Cerros del Rio volcanic field, located in the Española Basin, immediately west of Santa Fe (Fig. W1), contains typical examples of the products of rift processes that have occurred within the Rio Grande rift over the last 3 m.y. The best exposures of sedimentary and volcanic rocks of the rift and of proximal pyroclastic rocks from the adjacent Jemez volcanic field are along White Rock Canyon. This canyon of the Rio Grande, with a maximum depth of 300 m, begins at Otowi, along the road between Santa Fe and Los Alamos, and ends at Cochiti Pueblo (Fig. W2); much of the lower half of the canyon has been partly flooded by Cochiti Reservoir, a flood control and irrigation dam constructed upstream from the pueblo.

There are no roads and only difficult trails along White Rock Canyon; for this reason it is easiest to use a boat to gain access to the locations described in this guide. The following descriptions along the canyon are keyed to prominent geographic features in a manner similar to a traditional geological field guide.

Geologic setting and history of the Rio Grande rift

The Rio Grande rift is a chain of grabens over 1000 km long, which trend south from Leadville, Colorado, to the U.S.–Mexico border near Presidio Texas (Baldridge et al., 1984). The northern rift, from central Colorados to central New Mexico is well-defined, but breaks into a broad zone of grabens and horsts south of Socorro, New Mexico, where the actual rift is better defined by geophysical characteristics than by geomorphic features. Based upon petrologic, gravity, heat flow, and magnetotelluric surveys, the crust under the rift thins to as little as 30 km, in contrast with crustal thicknesses of 40 to 45 km east and west of the rift (summarized in Baldridge et al., 1984).

Extension, leading to rift formation, began between 32 and 26 m.y. ago, along a N–S-trending zone of weakness that had developed during late Paleozoic time (Chapin, 1979). Within the central segment, from Alamosa, Colorado, to Socorro, New Mexico, the rift consists of a NNE-trending line of en echelon basins.

The Cerros del Rio volcanic field, the subject of this field trip, is located in one of these en echelon basins, the Española Basin (Fig. W3); this basin is 40 km by 65 km, separated from adjacent basins by basement ridges (Manley, 1979). The predominantly coarse-grained clastic sedimentary rocks (partly consolidated arkosic sandstones, siltstones, and conglomerates) of the basin are overlain by or intertongued with volcanic and volcaniclastic rocks of the Jemez Mountains to the west, basaltic lava flows of the Taos Plateau to the north, and the Cerros del Rio volcanic field along its southern margin. Many of these sedimentary rocks represent alluvial fan deposits from the Precambrian and Paleozoic highlands to the east, north, and west (underlying the volcanic rocks of the Jemez Mountains) (Cavazza, 1986).

Most of these clastic sedimentary rocks are included in the Santa Fe Group, which ranges in age from ~21 Ma to 1.0 Ma (Hawley, 1978; Manley, 1979). Interbedded with these Santa Fe Group rocks are fallout tuffs, which serve well as marker beds and can be observed in outcrops along the road from Santa Fe to Española. These tuffs have ages of 9–14 Ma (fission track ages) and are probably from the Jemez volcanic field. Total thickness of these sedimentary rocks is between 1.5 and 2.5 km, depending upon the model used to interpret the gravity data or by extending slopes measured on the older rocks exposed at the basin margins (Manley, 1979). Detailed stratigraphic and paleontologic studies are covered in a monograph by Galusha and Blick (1971); later basin studies have built upon this work.

How old is the Rio Grande and how long has it been in its present location along the central rift? Originating in central Colorado, this river presently connects the rift basins until the rift is no longer visible in west Texas (where the river forms the border between the U.S. and Mexico it is known as the Rio Bravo del Norte). Fluvial gravels that include lithologies derived from outside the basin to the north, unconformably overlie sedimentary rocks dated at between 5.6 m.y. and 2.9 m.y. (Manley, 1979). The Rio Grande was established as a through-going river during early Pliocene time (Machette, 1978; Waresback, 1986).

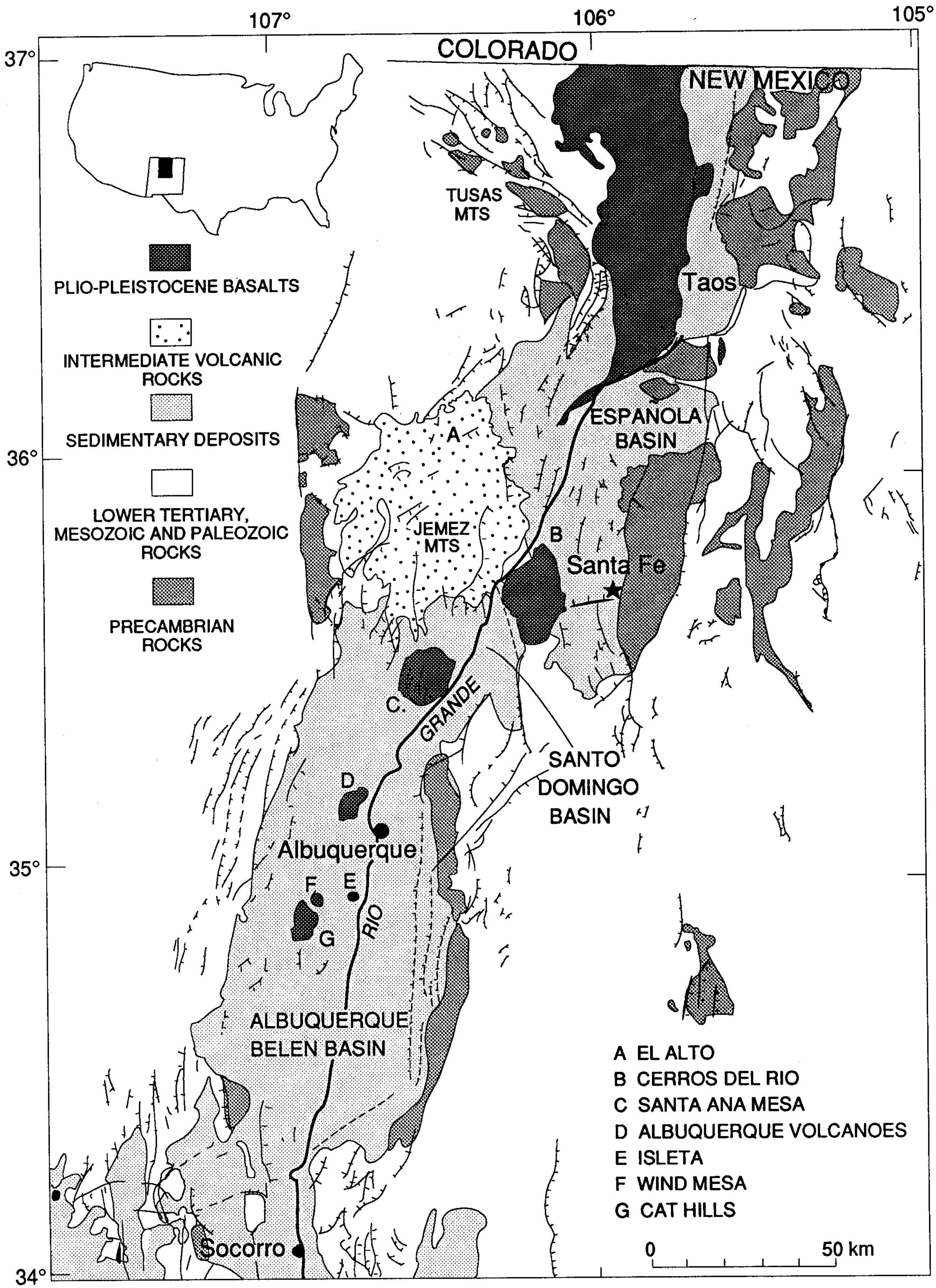

FIGURE W1—Generalized geologic map of the central Rio Grande rift (from Baldridge, 1979). Letters A to G indicate major Plio–Pleistocene volcanic fields. White Rock Canyon begins at the letter "B", on the north side of the Cerros del Rio volcanic field (see Fig. W6).

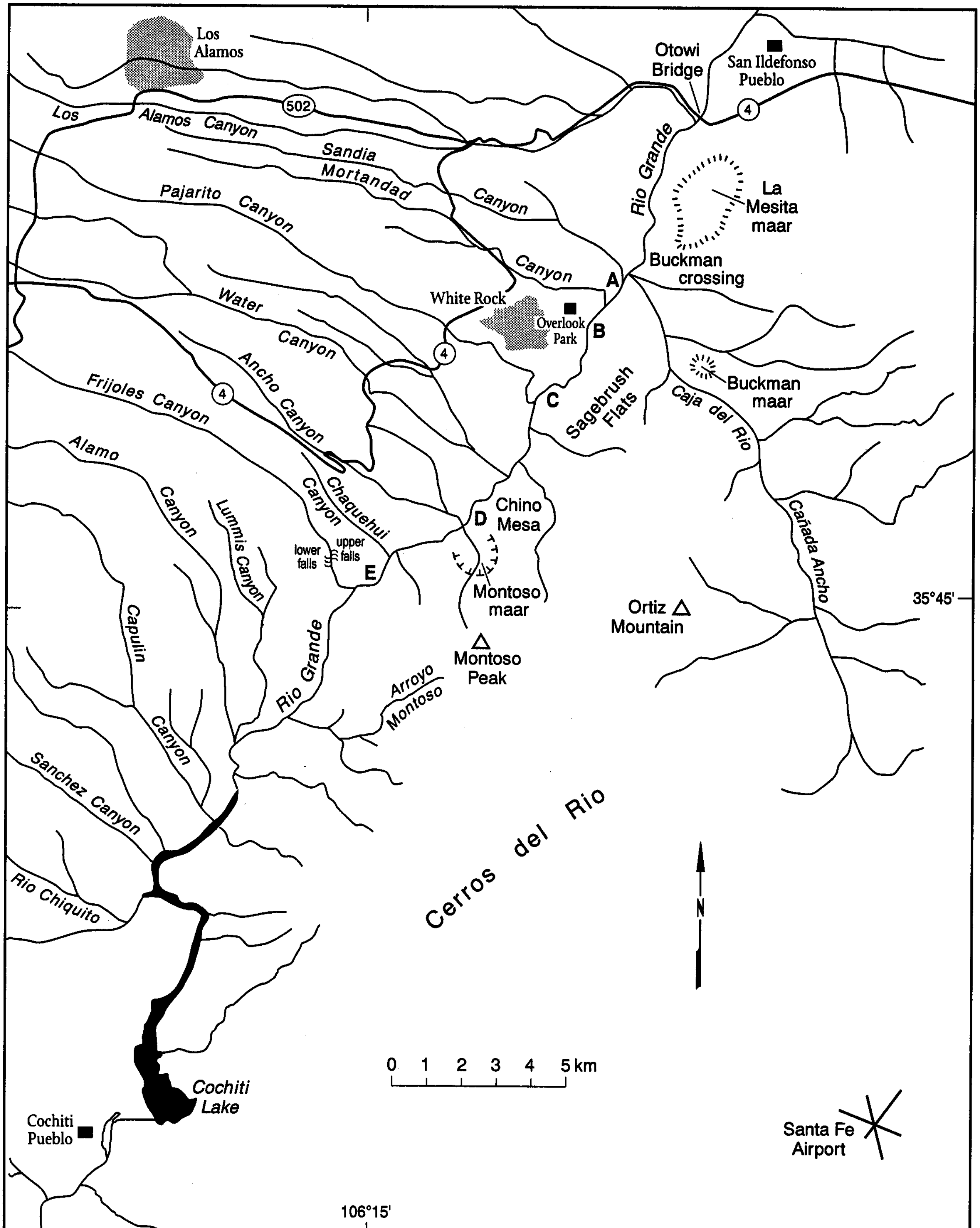

FIGURE W2—Map of White Rock Canyon, from Otowi Bridge, where NM-502 crosses the Rio Grande, to Cochiti Lake. Sites discussed in the text are keyed to geographic names on the map.

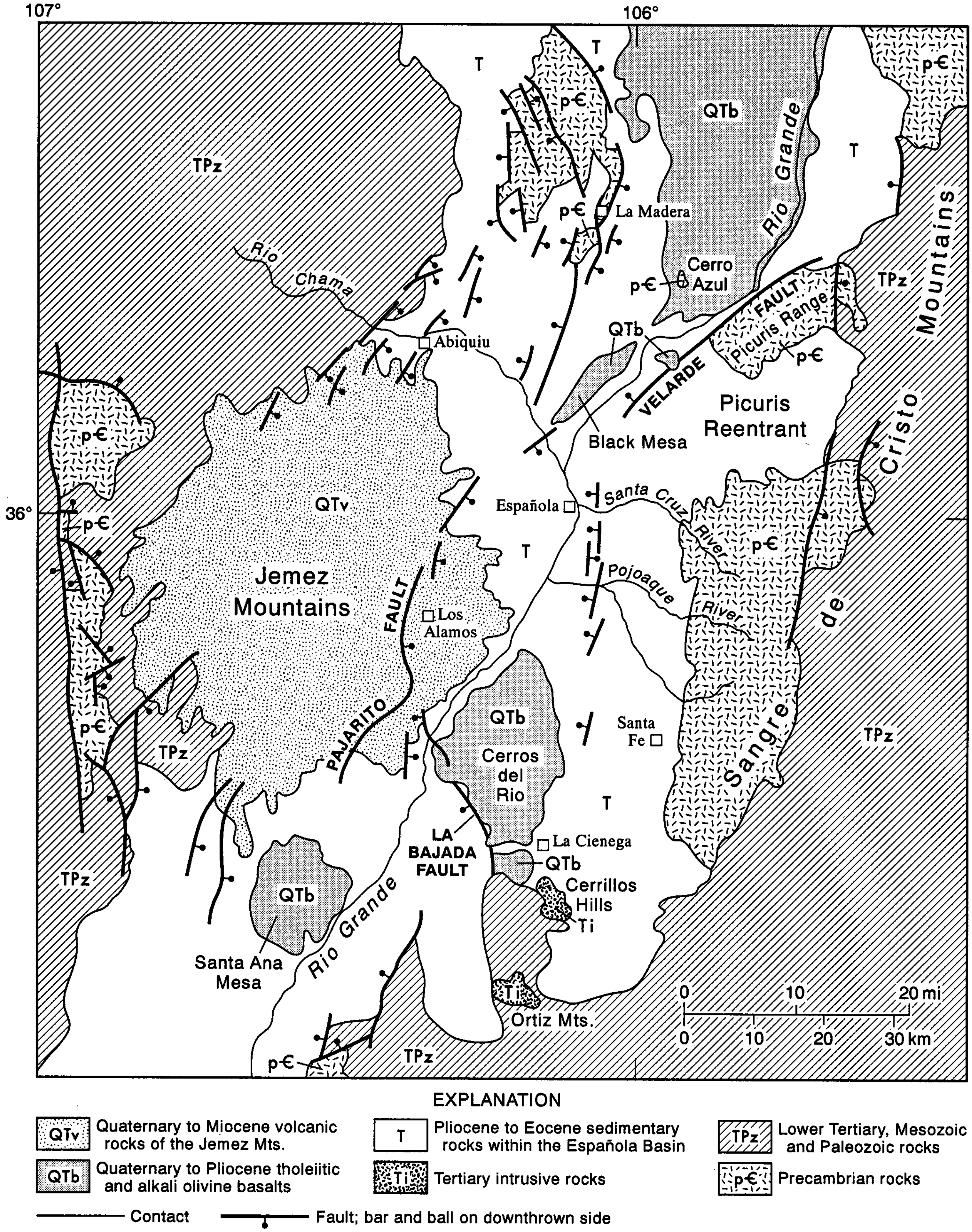

FIGURE W3—Geologic map of the Española Basin, from Manley (1979). White Rock Canyon is located within the northwestern portion of the Cerros del Rio volcanic field, north of the La Bajada fault (see Fig. W6).

At approximately 36° N, near the cities of Los Alamos and Española, New Mexico, the rift is crossed by a major crustal structure, the Jemez lineament. This lineament is defined by a SW–NE-trending line of volcanic fields, from the Springerville field in east-central Arizona to the Raton field of NE New Mexico and western Oklahoma (see Baldridge et al., 1989). The Jemez volcanic field, which has formed over the last 14 m.y., is located at the intersection of the Jemez lineament with the Rio Grande rift (Smith et al., 1970; Gardner et al., 1986).

Volcanism over the zone of extension and mantle diapirism represented by the Rio Grande rift shows considerable variation, from olivine tholeiites of the Albuquerque volcanoes (visible when looking west from the Albuquerque airport) to basanites and alkali olivine basalts of the Cerros del Rio volcanic field (Baldridge, 1979; Aubele, 1978; Duncker, 1988). The largest volume of basalt within the rift is in the Taos Plateau volcanic field, which consists of low-alkali tholeiitic rocks and interbedded intermediate and silicic lavas and tuffs (Dungan et al., 1984). Most of the basaltic volcanism visible within the rift has occurred over the last 5 m.y., although volcanism occurring along N–S faults on the rift margin within the Jemez volcanic field dates back to 14 Ma (Table W1).

White Rock Canyon

A diverse sequence of Miocene through Holocene rocks, in which are recorded volcanism, sedimentation, and erosion along the Rio Grande rift in the Española Basin, is exposed in White Rock Canyon. Sedimentary rocks of the lower and middle Santa Fe Group, overlying volcanogenic alluvial fan deposits (the Puye Formation), and Pliocene alluvium (upper Santa Fe Group), record alternating aggradation and canyon cutting centered near the present-day White Rock Canyon during late Tertiary time. This broad coalescent fan deposit (Santa Fe Group) extended beneath the eastern edge of what is now the Jemez Mountains. The thick basin fill accumulated to a local elevation of over 1860 m (6100 ft); this slowly aggrading base level existed from at least mid-Miocene to Pliocene time.

By mid-Pliocene time (~3 Ma), southwest-trending arroyos (from the east) and east- to southeast-trending canyons (from the Jemez Mountains) were deeply eroded to the level of the paleo-Rio Grande. Many of these paleocanyons are preserved under lavas and pyroclastic rocks of the Cerros del Rio volcanoes and by the two members of the Bandelier Tuff, erupted at 1.4 Ma and 1.0 Ma. Stratigraphic relationships may be seen by hiking up side canyons from White Rock Canyon.

The present location of the Rio Grande did not change much during activity of the Cerros del Rio volcanic field (~3 Ma). Although many of these lava flows are buried by the Bandelier Tuff, there are outcrops in canyons and on high ridges, around which the Bandelier Tuff flowed; basaltic lavas are intersected by water wells up to 10 km west of the Rio Grande (Dransfield and Gardner, 1985). In all of these cases, it appears that lavas erupted on the west side of the canyon flowed east and those on the east flowed west. The Cerros del Rio volcanic field covers over 800 km^2.

White Rock Canyon was repeatedly dammed by lava flows of the Cerros del Rio volcanic field. Lacustrine sedimentary rocks have been mapped in and north of White Rock Canyon; along the highway to Los Alamos there is a lava delta, where lavas from a vent on the west side of the valley flowed into a lake with a water surface of at least 1860 m (6100 ft). Similar lava deltas are visible in many canyons between Los Alamos Canyon and the village of White Rock.

Along the northwestern margin of White Rock Canyon is a 200 km^2 volcanogenic alluvial fan (the Plio-Pleistocene Puye Formation), which consists of volcanogenic sedimentary rocks and interbedded tuffs derived from the dacitic and rhyolitic domes of the Tschicoma Formation of the Jemez volcanic field (Turbeville et al., 1989, Fig. W4). The intertonguing Totavi Formation (Totavi Lentil of Griggs, 1964) includes the earliest dated deposits of the ancestral Rio Grande in northern New Mexico (4–4.5 Ma; Waresback, 1986). The river does not appear to have changed course considerably since before mid-Pleistocene time, when the growing Puye alluvial fan may have pushed the river several km east of its present position (Waresback, 1986). Deposition of this alluvial fan was con-

TABLE W1–Geochronology of volcanic events along the Río Grande, Cerros del Río. F.T. = fission track date.

Event	Age (Ma)	Method	Reference
Santa Fe Group sedimentary rocks	~21–1.0	K/Ar; F.T.	Hawley, 1978; Manley, 1979
Basaltic and andesitic volcanism, Cerros del Rio volcanic field	3.0–0.96	K/Ar; inferred from stratigraphy	Dethier (in press); Manley, 1976
Puye Formation alluvial fan, with a source area in the Jemez Mountains (contains interbedded tuffs from the Tshicoma volcanoes)	~5–2	K/Ar; inferred from stratigraphy	Turbeville, et al., 1989
Bandelier Tuff Upper (Tschirege) Member	1.12	K/Ar	Doell et al., 1968
Bandelier Tuff Lower (Otowi) Member	1.45	K/Ar	Doell et al., 1968
Damming of White Rock Canyon by slump blocks	>0.15	inferred from stratigraphy	Dethier (in press)
El Cajete Pumice (from Valles intracaldera rhyolite domes and craters)	0.13	K/Ar	Self et al., 1988

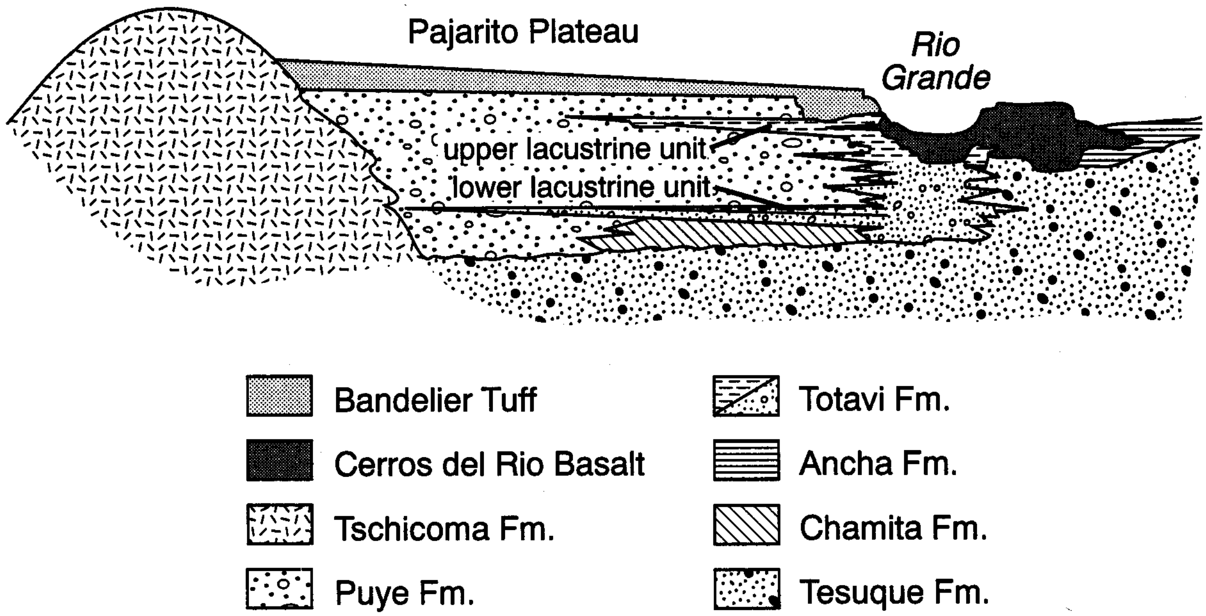

FIGURE W4—Schematic cross section of the Pajarito Plateau of the eastern Jemez Mountains volcanic field, the canyon of the Rio Grande, and Cerros del Rio, showing the relationships between the Puye alluvial fan (derived from the intermediate volcanic rocks of the Tschicoma Formation), sedimentary rocks of the Santa Fe Group (including ancestral river deposits—Totavi Formation, as well as Tesuque, Chamita, and Ancha Formations), and the basaltic rocks of the Cerros del Rio (from Turbeville et al., 1989).

temporaneous with basaltic volcanism, as basaltic ash layers were interbedded with the volcanogenic clastic rocks.

The upper portions of White Rock Canyon are flanked by slump blocks (toreva blocks), which appear to have been stable since at least mid-Pleistocene time (Bandelier Tuff is part of some of the slide blocks, but the blocks are overlain by 150 ka El Cajete pumice that has not been dislocated). Most of these slump blocks consist of basaltic lava flows that overlie rift sedimentary rocks.

Before about 2 Ma, the Rio Grande appears to have flowed at an elevation of less than 1834 m (6020 ft). Field relations demonstrate that the upper part of White Rock Canyon was at least 76 m (250 ft) deep at that time. Before eruption of the lower member of the Bandelier Tuff (1.45 Ma), side canyons west of the Rio Grande (south of Water Canyon) had been incised into basalt flows to a depth of at least 150 m (500 ft). White Rock Canyon was probably cut into landslide debris to an elevation of 1768 m (5800 ft).

Since eruption of the upper member of the Bandelier Tuff, White Rock Canyon has been deepened by about 122 m (400 ft). Elsewhere in the Española Basin, rapid incision of the river occurred during mid-Pleistocene time, probably in response to climatic change and to complex regional geomorphic factors, such as (1) integration with a through-flowing "Lower Rio Grande" system, and (2) upstream enlargement of the headwater areas to include the San Luis Valley region of Colorado. Soil development in the canyon supports the climatic hypothesis, but stream terrace remnants are small. The highest fluvial terrace preserved in this area is about 46 m above the Rio Grande and is older than 150 ka.

Hydrology of White Rock Canyon and Effects on Eruption Processes

The Santa Fe Group clastic sedimentary rocks underlying the Cerros del Rio volcanic field are generally considered to be a single hydrologic unit and part of the Tesuque aquifer system (Hearne, 1980; Coons and Kelly, 1984). Yields of 2725 m^3/day (500 gpm) have been reported from this aquifer. General flow is from recharge areas at the base of the Sangre de Cristo Mountains toward the west and the Rio Grande. Thickness of the Santa Fe Group and the aquifer is not known.

Under the Cerros del Rio volcanic field, the water table slopes from an elevation of 1798 to 1768 m (5900 to 5800 ft) along the eroded margins of the field down to 1676 to 1615 m (5500 to 5300 ft), where the aquifer reaches the surface at the Rio Grande (Borton, 1963). During the last 2 Ma, the level of this aquifer was possibly 150–175 m (492–575 ft) higher, depending upon the depth of erosion since then. Aquifer levels were between 90 and 275 m (300 and 900 ft) below the base of lava flows and cones that make up most of the field.

Vents of the Cerros del Rio located closest to the Rio Grande are maar volcanoes (Fig. W5). During the period from ~2.5 to ~1.5 Ma these were erupted at a time when the level of the Rio Grande (and the aquifer) was at an elevation of 1768 m (5800 feet). Bases of the northernmost maar volcanoes, located along northern White Rock Canyon, are at an elevation of about 1768 m (5800 ft); maars farther down White Rock Canyon are 30–60 m lower, but so was the river. Other vents, located at elevations over 100 m higher than those of the tuff rings (also farther from the ancestral river), are all composed of cinder cones, spatter cones, and associated lava flows.

All of the eruptions within the Cerros del Rio volcanic field, occurring where the aquifer was shallow (nearly at the surface), were phreatomagmatic (hydrovolcanic). Vents reaching surfaces 100 m or more above the water table formed cinder or spatter cones with little or no evidence of magma/water interaction. An excellent example of this effect can be seen by walking up Frijoles Canyon (a side canyon trending NNW from the Rio Grande). Basaltic volcanoes are exposed along this and adjacent canyons. Maar volcanoes are visible from elevations of ~1675 m (5500 ft) to 1706 m (5600 ft). From elevations of ~1800 m (5900 ft) and up, there are cinder cones. The elevation effect is also visible in the larger maars, where the craters are filled with scoria and lavas.

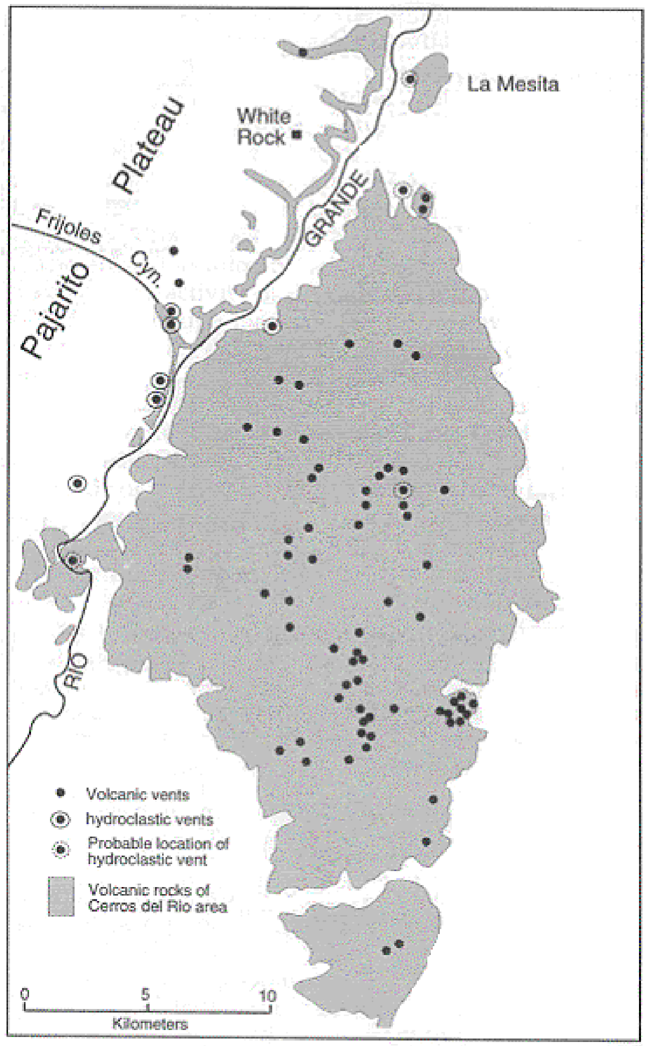

FIGURE W5—Sketch geologic map of the Cerros del Rio volcanic field, showing the location of known vents, including hydroclastic vents (tuff rings) (modified from Kelley, 1978).

Guide to White Rock Canyon

Reaching the head of White Rock Canyon (Fig. W6)

All of White Rock Canyon is accessible by foot, although much of it requires rigorous hiking and a little scrambling. The head of the Canyon, at the intersection of NM–502 and the Rio Grande (Otowi Bridge), is on land belonging to San Ildefonso Pueblo; you most obtain permission from the Pueblo before entering the canyon at this point. The most effective way of seeing the canyon is by boat, entering either from Otowi Bridge (permit required) or from Buckman Crossing (no permit required), which can be reached from Santa Fe. If you float, plan on at least one very long day, or perhaps two days, camping in the canyon.

Entering from the North

The easiest way of reaching this canyon is to drive north of Santa Fe along US–84/285 until you reach NM–502, turning west toward Los Alamos (15 mi from the intersection of Alamo Drive and St. Francis Drive at the northern end of Santa Fe). Most of the sedimentary units along this drive are arkosic sandstone, siltstone, and conglomerate of the Tesuque Formation of the Santa Fe Group. Type sections of this formation are exposed in arroyos below the Santa Fe Opera.

Travel west along NM–502 to Otowi Bridge on the Rio Grande (7 mi). If you enter the river and canyon here, you must acquire a permit from the offices of San Ildefonso Pueblo (fee required). Even if you do not enter the canyon here, there is a great view from the highway of La Mesita maar, discussed below.

Entering from the East

Travel north of Santa Fe (1.3 mi past the traffic light at St. Francis Drive and Alamo Drive) to the La Tierra Road (left off of US–84/285). Take this paved road due west for about 5 miles until it becomes a gravel road (be careful during the summer, for flash floods can inundate this road). This road becomes Buckman Road, which ends at Buckman Crossing on the Rio Grande.

The road continues WNW, following the old Chili Line railroad grade, through Santa Fe Group sedimentary rocks until it reaches the base of an escarpment formed by lavas of the Cerros del Rio volcanic field, which overlie the sedimentary rocks. At this point, the road turns north and continues along Cañada Ancho, at the base of the escarpment, for 6 mi until it reaches the Rio Grande at Buckman Crossing. This is an excellent place to begin your trip down White Rock Canyon. It is also the loca-

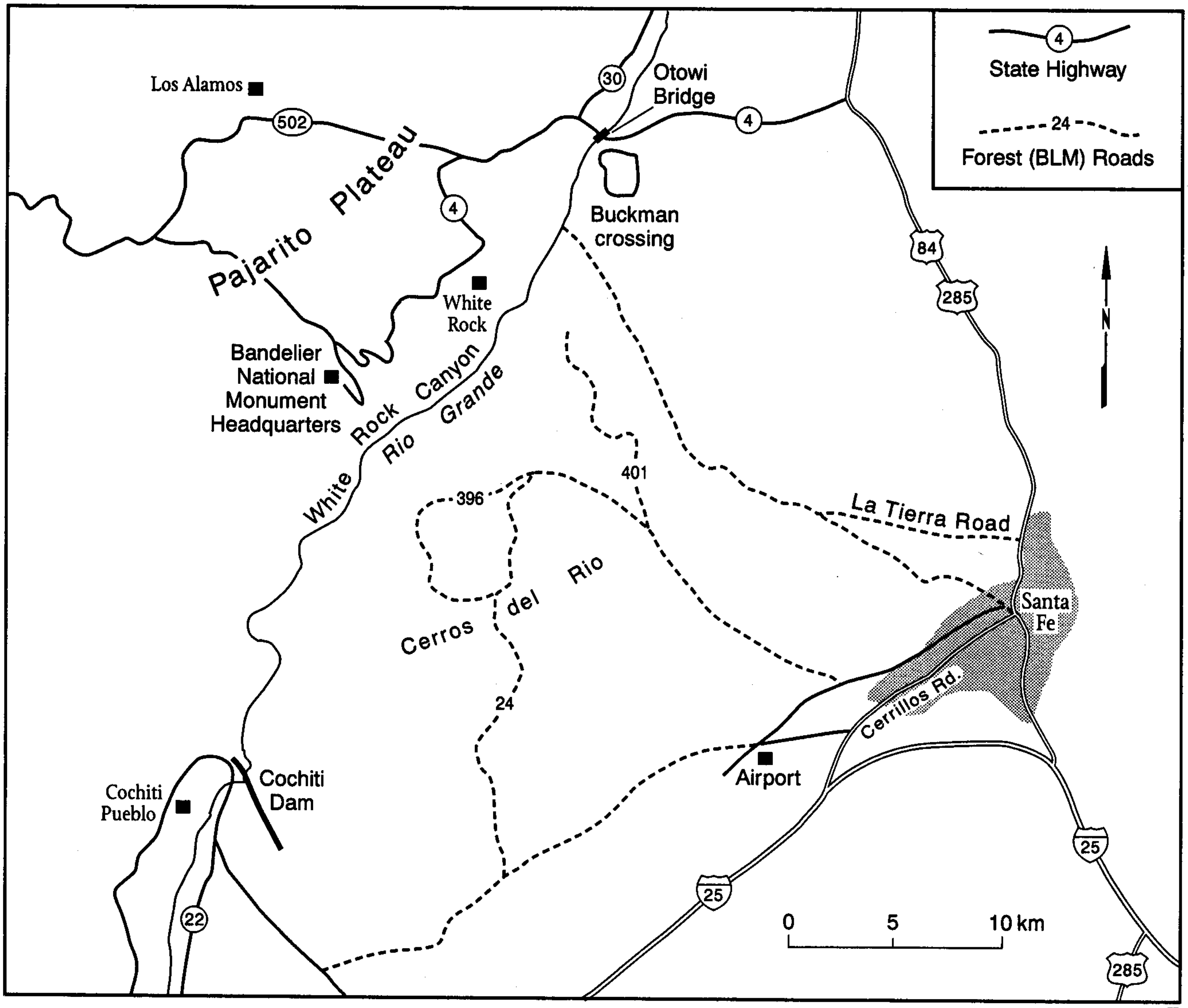

FIGURE W6—Access to White Rock Canyon: (1) via NM–502 to Otowi Bridge, (2) gravel and dirt roads from Santa Fe across the Cerros del Rio and to Buckman Crossing, (3) from Cochiti Dam, and (4) trails from White Rock village and Bandelier National Monument.

tion of two of the best exposed maar volcanoes along the canyon; La Mesita to the north of Buckman Crossing and the maar of Caja del Rio, south of the Crossing.

Entering from the West

White Rock Canyon can be reached by foot from the western side of the Rio Grande in several places from NM–4:

White Rock: Trails lead from the village of White Rock. Ask for their location at businesses within the village or at the new White Rock Information Center along NM–4.

Between White Rock and Bandelier National Monument: Trails are east of NM–4, between the village of White Rock and Bandelier National Monument. The most accessible of these trails is the one to Ancho Rapids. Trailheads are visible on the left (east) side of the highway (going toward Bandelier) as dirt roads with gates.

Bandelier National Monument: There are many trails in the Bandelier National Monument Wilderness Area. The easiest trail to the Rio Grande is that along Frijoles Canyon, below the Monument Headquarters (the Falls Trail), which will take you through several maar volcanoes in cross section. The geology along the Falls Trail is described later in this guide, in reverse, starting at the river. Other trails, including the one to Capulin Canyon, will require a very long day or camping overnight.

La Mesita maar

Introduction

La Mesita maar is the northernmost of the 70 volcanoes of the Cerros del Rio volcanic field. It is one of the best exposed maars in the volcanic field and is visible from NM–502 at Otowi Bridge. The southern end of this volcano can be reached from Buckman Crossing on the Rio Grande; to visit the northern half requires permission from San Ildefonso Pueblo. The purpose here is to describe the role of water in the development of La Mesita maar, as determined from detailed examination of the bedded deposits.

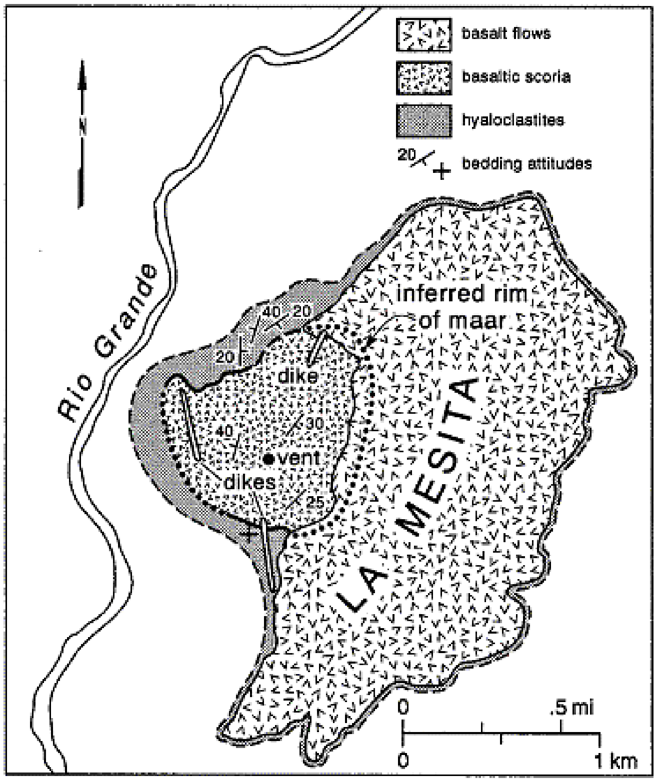

FIGURE W7—Geologic sketch map of La Mesita maar. The northern end of this maar is easily seen from the Otowi Bridge on NM–502.

Stratigraphy

La Mesita maar (Fig. W7) originally had a wide crater, but it is now filled with basaltic ash, scoria, and agglutinate from Strombolian eruptions. The maar rim beds that lie beneath the Strombolian fallout tephra make up the rim of the original maar volcano. They are about 100 m thick and are composed of fragments derived from hydroclastic eruption processes, as will be shown in a later section. The stratigraphic sequence, from maar rim beds to crater-filling scoria, is now exposed in cliffs for about 3 km along the course of the Rio Grande.

Paired coarse-fine bed sets. The rim beds of La Mesita maar are composed of a repetitive sequence of bedding sets consisting of coarse layers overlain by fine layers, with contacts being sharp to broadly transitional on a centimeter scale. The thickness ratio of coarse- to fine-grained layers varies greatly but averages about 1:1. The coarse-grained layers contain fragments that range from coarse ash to medium lapilli (~3–10 mm). The coarse-grained layers grade from those lacking fine particles (< ash size) with an open-work texture to those with a fine-grained matrix.

The coarse-grained layers with clast-supported texture are considered to be depleted in fine particles, which occurred by processes of elutriation within ascending eruption columns or pyroclastic surges. The more poorly sorted coarse-grained layers are believed to have also formed by one or the other of the two processes but with sorting processes not reaching completion. The fine-grained layers within single paired sets may have been derived by separation from the coarser debris. However, many of the fine layers contain coarser clasts and appear not to have undergone any sorting process.

Some of the rim beds contain bedding-plane sags, wavy units, and crossbedded units. Most of the layers, however, are plane beds commonly with inverse grading.

Within many of the coarse- or fine-grained beds there are laminations of coarser- or finer-grained ash but without sharp bedding planes (Figs. W8, W9). For example, in some of the coarse-grained beds there is a faint layering that is manifested only by light dusting of fine ash along a horizontal plane. Lines in some of the finer-grained beds are of slightly coarser ash grains, of one to several grain widths, that are imbedded in the silt-sized material. The ash in these laminations may increase or decrease up or down in the layer to produce small-scale graded bedding. The laminations within the section at La Mesita commonly show very subtle low-angle cross-bedding and lenticularity.

Each coarse-fine bedding set in the La Mesita maar rim beds is considered to be the result of an eruptive pulse, with tephra being deposited by fallout or by flowage processes. The initial assessment of whether beds are emplaced by flowage or by fallout processes is based upon certain bedding criteria. If internal laminations show subtle cross-bedding, the bedding set is interpreted to have been emplaced by flowage. Coarse-grained layers that lack bedding-plane sags at their bases and show inverse grading are also interpreted to have been emplaced by flowage processes. Another criterion of flowage emplacement is the sharply defined basal unconformity where the upper beds have eroded underlying beds, resulting in crosscut-

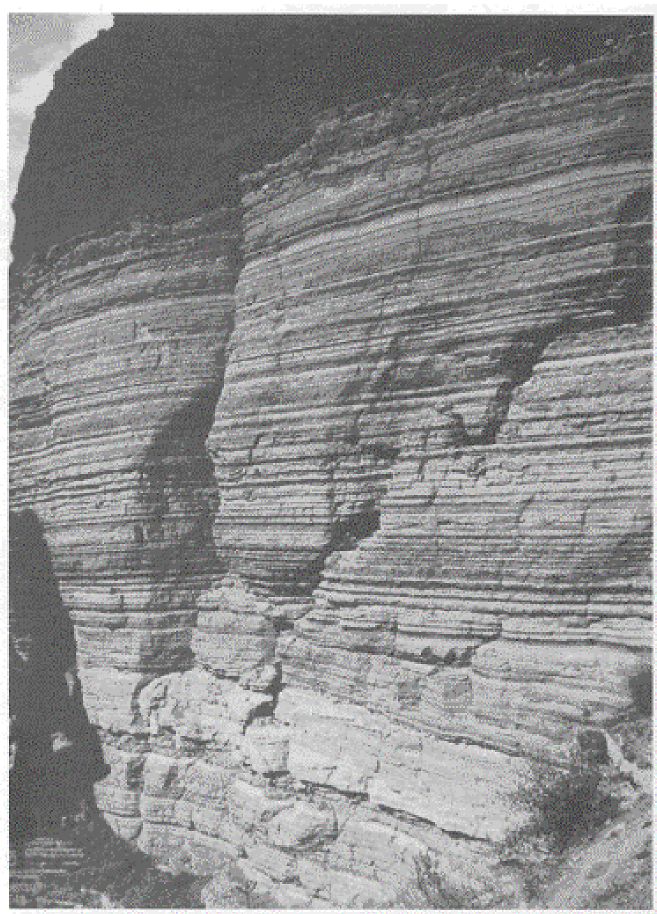

FIGURE W8—Bedding sets within the La Mesita maar deposits.

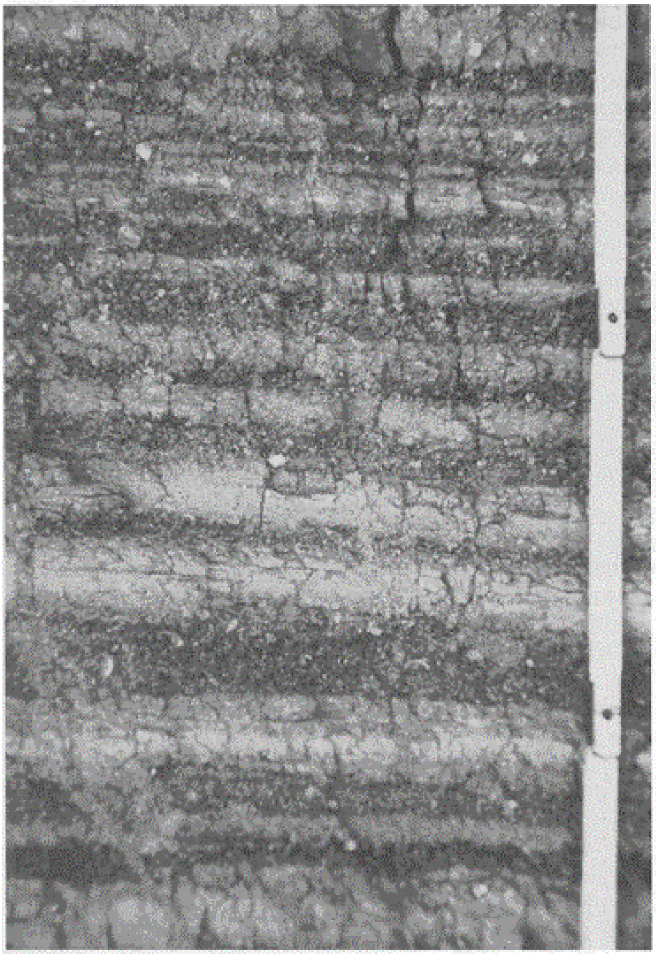

FIGURE W9—Close-up of bedding sets within La Mesita maar deposits. Bedding sets are a few cm thick. The scale is 2 cm wide.

ting relationships. Erosion by surges is also indicated by the mixing of two layers at their contact or the presence of material from the underlying layer in the overlying layer.

Evidence for deposition by fallout includes: (1) fragments at the base of a layer that distort the underlying layers (impact); (2) layers and laminations that drape smoothly over minor roughness elements of the underlying layer; (3) internal laminations that are parallel and are not associated with lenticular layers or other angular elements; and (4) relatively good sorting.

The explanation for why bedding pairs begin with a coarse-grained layer and end with a fine-grained layer depends upon whether the bed is believed to be of flow or fallout origin. Fallout beds with inverse grading can be interpreted in terms of eruptive energy: eruptions with progressively larger kinetic energy, expelling larger clasts as the eruption progresses (Wohletz and Sheridan, 1983). Normal grading can form during energetic eruptions, with the finest material elutriated from the eruption column, falling after the eruption phase.

Beds emplaced by flowage also show a progression from coarse to fine-grained tephra, but for a different reason. Coarse-grained basal units may form from the head of a passing surge, which is followed by the finer-grained and more poorly sorted body of the surge (Fisher et al., 1984, 1986). Because most of the beds have been assessed as emplaced by flowage at La Mesita maar, the following discussion assumes flow emplacement unless otherwise stated.

Description of the Eruption Sequence

The stratigraphic sequence at La Mesita maar is about 100 m thick, but complete sections could not be measured in any single locality. The upper portions of the sequence are well-preserved on the uphill side of a dike, whereas the lowermost part of the sequence is exposed in a large, dissected slide block, 400 m east of an elevation of 5487 ft along the Rio Grande (4 km downstream from the Otowi Bridge).

There are subtle color differences within the tuff beds, which reflect changes in clast compositions. Pink hues characterize beds containing abundant sand- and silt-sized material from the underlying Tesuque Formation. Yellowish hues are derived from palagonitized vesicular basaltic pumice fragments. Bluish-gray (medium gray) hues are typical of deposits containing abundant angular or subangular accessory to juvenile pyroclasts of hydroclastic (phreatomagmatic) origin. Dark gray and black deposits are mostly quartz xenocryst-bearing tachylitic scoria of Strombolian origin.

Beginning at the base, the first half meter of section is dominantly pink to medium gray, changing to yellowish gray over the next 3–4 m, and pink and gray for the next 20–30 m of interbedded fine-grained and coarse-grained layers. This sequence contains abundant lithic clasts (sand grains) derived from the underlying sandstones; these are progressively diluted upward with abundant juvenile hydroclastic particles. The uppermost sequence is mostly pink and contains a much higher percentage of accidental lithic clasts.

Petrology and granulometry of the sequence. Eleven samples of La Mesita tephra were sieved and characterized by scanning electron microscopy (SEM). In general samples are finer-grained in the middle of the eruption sequence and coarser-grained near the base and top; the strongest hydroclastic fragmentation occurred midway through the eruption. Figure W10 distinguishes scoria from hydroclastic tephra on a sorting-median diameter plot, which also shows relative fields of fallout and pyroclastic surge bed forms. Histograms of size frequency show a distinct polymodality for most samples, a feature typical of many tephra samples in general.

Clast shapes, as revealed by SEM, show several general trends: (1) lithic clasts of the Tesuque Formation are subangular to subrounded quartz and feldspar grains and ag-

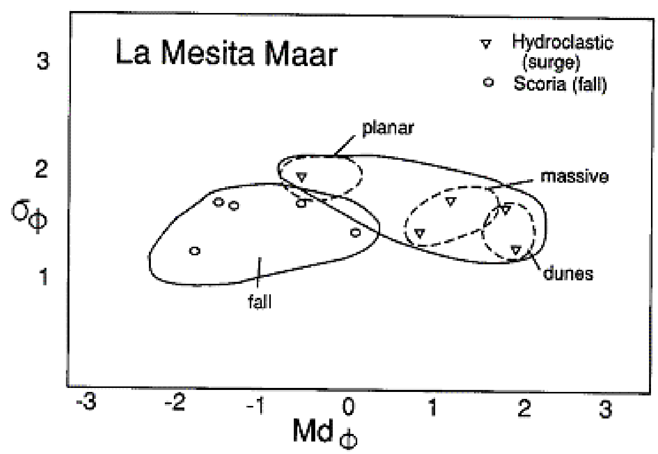

FIGURE W10—Sorting (σφ) vs. median diameter (Mdφ) plot for hydroclastic (phreatomagmatic) surge deposits and scoria fall deposits of the La Mesita maar.

gregates of mudstone; (2) juvenile clasts larger than 0.5 mm are dominantly scoriaceous while smaller grains consist of angular blocks and plates; (3) variable degree of palagonitization has produced hydration cracks and rinds that separate from glassy clasts; (4) some samples contain numerous spherical clasts with median sizes of about 2.5 ϕ (0.175 mm)—some of these appear to be alteration products, while most appear to be juvenile; (5) aggregation of fine ash is common; and (6) the scoria show fused surfaces, whereas hydroclastic shards have very irregular surfaces with abundant coatings of alteration products. The presence of spherical ash suggests some relatively fluid magma fragmented by hydrodynamic instabilities. However, the hydroclastic ash is dominantly angular, having been derived from brittle breakage of pre-vesiculated magma.

Conclusions regarding the origin of La Mesita maar

The hydroclastic tephra of the rim beds has abundant fine-grained clasts typical of hydroclastic eruptions. Experiments with water and thermite melt to simulate water-magma interactions indicate that explosive efficiency is strongly controlled by water-melt mass ratios and by confining pressure. Explosive efficiency is a percentage of mechanical to thermal energy, which is at its maximum when water-melt ratios are between 0.3 and 1.0 (Wohletz and Sheridan, 1983). The efficiency of fragmentation increases with explosive energy and the degree of water superheating.

One important measure of explosive energy and water-melt ratios is the amount of fine-grained ash in pyroclastic deposits ("fine-grained" is < 62.5 µm). The increased melt surface area caused by efficient fragmentation promotes highly efficient heat exchange between water and melt. More glass is produced because the particles are rapidly chilled.

Figure W11 (from Wohletz and McQueen, 1984) shows one of the models that can be used to interpret our data from La Mesita maar. As is shown for the model, rising magma encounters near-surface ground water; water is vaporized and magma and basement rocks are fractured and fragmented. Fragments from this level of magma–water interaction are composed of chilled juvenile pyroclasts and lithic clasts.

Figure W12 shows the variables that can be used to relate the deposits to the model shown in Figure W11. The volume of fine-grained glassy pyroclasts in the deposits is a measure of the amount of water in the system. The amount of lithic clasts (from rocks underlying the volcano) reflect the size, permeability, and volume of water within the aquifer that comes into contact with magma in the vent. Volcanic lithic clasts (crystalline) are a measure of the magma/water ratio in the system. Using these three parameters, we can define four fields, based upon clast components in hydroclastic deposits (Fig. W13).

Preliminary estimates suggest that in the lower part of the La Mesita maar, the fields move from C to D (high magma input and low to high volumes of water. Higher in the deposits, the fields alternate between A and B, which suggests enlargement of the vent area (increasing basement lithic clasts) and less magma input relative to the volume of available water.

In the lower half of the La Mesita deposits the layers occur as bedding sets, each consisting of a coarse-grained base and fine-grained top. We consider each bedding set to represent a single explosion of the eruption (see Fisher and Schmincke, 1984, p. 348); explosions progress from

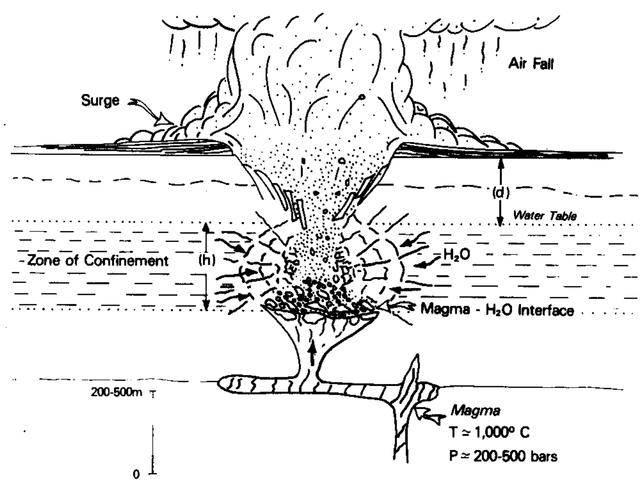

FIGURE W11—Model of hydroclastic (phreatomagmatic) eruption (from Wohletz and McQueen, 1984); explosive magma-water interactions are occurring within a shallow aquifer.

Rate and/or Volume

Rate and/or Volume: Water	Magma: Low		Magma: High	
Low	Ⓐ		Ⓒ	
	VL	low	VL	abund.
	BL	higher	BL	low
	G	low	G	low
High	Ⓑ		Ⓓ	
	VL	lower	VL	abund.
	BL	higher	BL	low
	G	abund.	G	abund.

VL = Volcanic lithics
BL = Basement lithics
G = Glass

FIGURE W12—Variables that can be used to relate maar deposits to the model shown in Fig. W11.

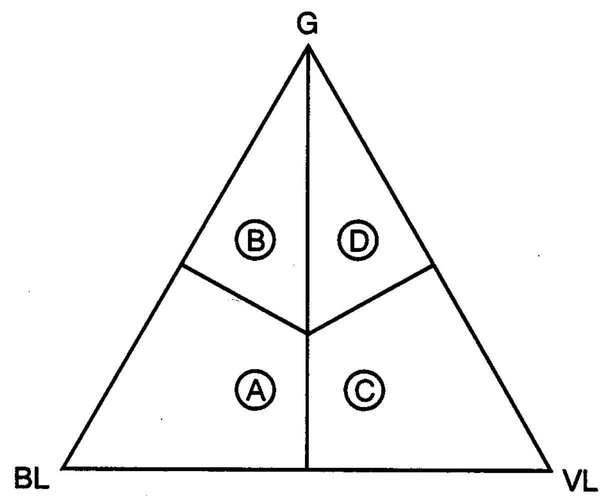

FIGURE W13—Four fields for hydroclasticdeposits, based upon component clasts; see Fig. W12 for explanation.

low to high efficiency where some water entered the vent, triggering the eruption and fracturing rocks adjacent to the vent. Subsequently, more water moves into the area of rising magma. We visualize a rapid, progressive build-up of released energy, higher eruption columns with associated surges, and greater production of fine-grained ash.

Near the top of the maar deposits lithic clasts from basement rock exceed those derived from the vent, but there is still alternation between coarse- and fine-grained (and glassy) hydroclastic tephra. Thus, these bedding sets are still characteristic of the same transport and depositional style, but enlargement of the vent results in a greater proportion of basement lithic clasts. With continued eruptive activity and construction of a volcanic tuff ring or cone, the hydrovolcanic activity ceases, followed by magmatic, Strombolian eruptions. One hypothesis explaining these changes is that fractures are closed by chilled magma along dike margins as the vent is enlarged, closing off inflow of ground water.

Another hypothesis is that vesicle growth gives higher surface area within rising magma, which promotes efficient interaction with external water. Hydrovolcanism is absent where pyroclastic sections are greater than 100 m thick. Vesicle growth in many basaltic magmas typically occur at depths of > 100 m (Heiken, 1971). It is possible that the conduit pressure increased as the tuff ring grew to the point at which vesiculation was inadequate to promote interaction with ground water at depths < 100 m, and the eruption ended with Strombolian activity and filling of the crater with a lava lake.

Buckman maar

An oval, 1 x 1.2 km, 150-m-high mesa, located along the Caja del Rio, 2.2 km SE of Buckman Crossing, Buckman maar (unofficial name) is an eroded maar topped with deposits of scoria and bombs and a basaltic lava flow. The summit is most easily approached from the east, along a trail from Buckman Road.

The general stratigraphic sequence consists of massive, block-bearing hyalotuff, overlain by well-bedded orangish-brown hyalotuff (mostly plane beds). Interbedded with the well-bedded tuffs are blocks of pink, fine-grained sandstone, up to 30 m long. Dikes cut the tuff sequence and were feeders for the lava lake and flows overlying the maar. A lava flow capping the mesa south of this maar has an age of 2.3 ± 0.1 Ma.

Beginning of float trip

(distances are measured from Buckman Crossing)

From Buckman Crossing to the mouth of Water Canyon (west), 0–8 km, the widest section of White Rock Canyon (2 km to 2.4 km wide) is characterized by large slump blocks on both sides of the river. Capping the slumps next to the river (E side at km 0.1) are slack-water deposits, 35 m thick, that record damming of the river sometime before 60 ka.

Mortandad Canyon

(location A, west side, 1 km south of the Crossing)

Mortandad Canyon is little modified by slumping and contains a well-exposed section of rocks typical of White Rock Canyon (Fig. W14). This section includes 76 m of sandstones and conglomerates, which are intertongued Santa Fe Group sedimentary rocks (metamorphic and plutonic provenance) and Puye Formation (volcanic provenance), and ancestral Rio Grande deposits (Totavi Formation of Waresback, 1986).

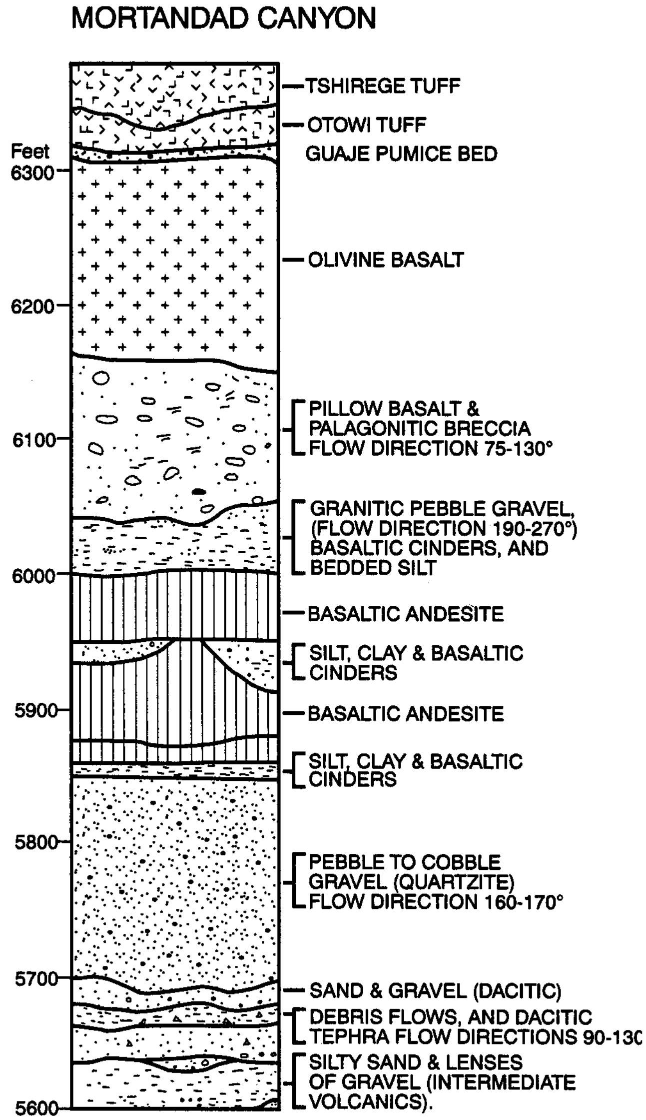

FIGURE W14—Stratigraphic section, Mortandad Canyon (A, west side of the river; elevations are in feet).

140 m of lavas of the Cerros del Rio overlie this with some interbedded river gravels. The lower third consists of basaltic andesite flows and some scoria, the middle third is a lava delta, consisting of interbedded palagonitized hyaloclastites and pillow lavas (the foreset beds of the lava delta indicate flow from west to east); these appear to have flowed into one of the lakes formed by damming of the Rio Grande by lava flows or slumps. The upper third of the basalt section is a massive olivine basalt flow with well-developed columnar jointing near the flow top, which has a K–Ar age of 2.3 ± 0.1 Ma.

The section is overlain by members of the Bandelier Tuff, including the Guaje pumice fall and distal Otowi (lower) Member (1.4 Ma) and Tshirege (upper) Member (1.0 Ma). Paleovalleys and paleocanyons cut into the Cerros del Rio basalts, west of the Rio Grande, are partly filled with Bandelier Tuff; most of the canyons seen today are resurrected paleocanyons.

North end of Sagebrush Flats

(location B, immediatly south of the Crossing, east side of the river)

The ridge immediately south of Buckman Crossing is modified by slumping, but a section of about 140 m can be pieced together. It consists mostly of intertongued sandstones and conglomerates of both the Santa Fe Group and Puye Formation. This is the only location in the canyon where the Puye Formation crops out east of the Rio Grande. Olivine basalt flows are exposed in the middle and top of the section (Fig. W15).

Southwest Sagebrush Flats

(location C, 6 km south of the Crossing, east side of the river)

Thirty meters of interbedded sandstones, cinders, and phreatomagmatic tuffs are exposed below the base of a paleocanyon that is partly filled with 100 m of the Tshirege (upper) Member ignimbrite of the Bandelier Tuff (Fig. W16). This is one of only three or four places in the canyon where Bandelier Tuff ignimbrites are preserved on the canyon's east side.

Chino Mesa and Montoso maar

(location D, 7.9 km south of the Crossing, east side of the river)

Most of the 210 m of section exposed here consists of interbedded phreatomagmatic tuffs, thin basalt flows, and an andesite flow (Fig. W17). The andesite of Montoso Peak is exposed along the Canyon for nearly 10 km, and may have been a major unit responsible for one of the episodes of damming the Rio Grande.

The Montoso maar is located in a side canyon (east side of river), 2.3 km due north of Montoso Peak. Aubele (1978, 1979) interprets this as the dissected throat of a collapsed maar. The amphitheater-shaped, inward dipping tuff beds of these phreatomagmatic deposits overlie well-exposed talus deposits and show no evidence of collapse. It is possible that this maar was erupted in a paleocanyon, coating the walls of the canyon with sticky surge and fall deposits, which extended beyond the edge of this paleocanyon and out onto the lava plateau. On the plateau, these deposits are overlain by later lava flows. The canyon-filling tuff deposits were later cut by dikes feeding younger lava flows.

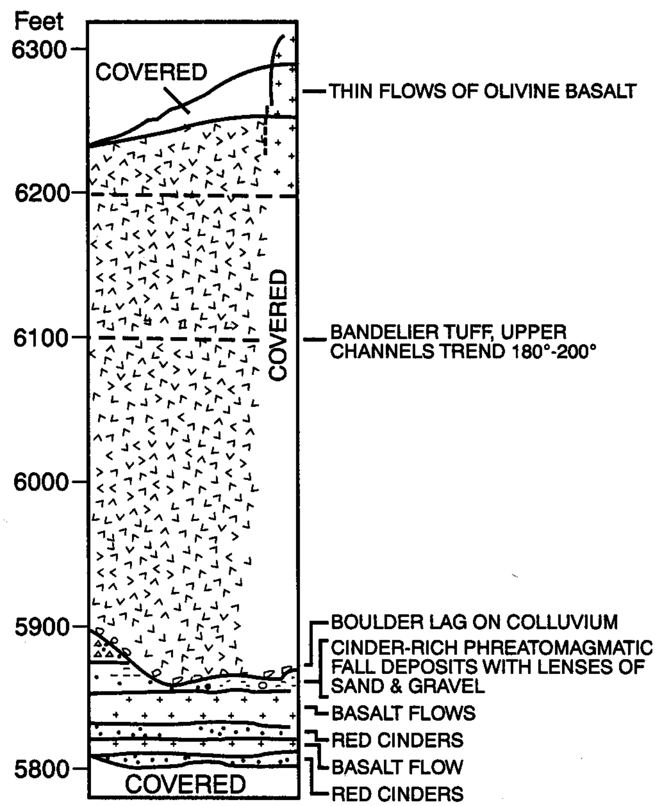

FIGURE W16—Stratigraphic section, southwest end of Sagebrush Flats (C, east side of the river; elevations are in feet).

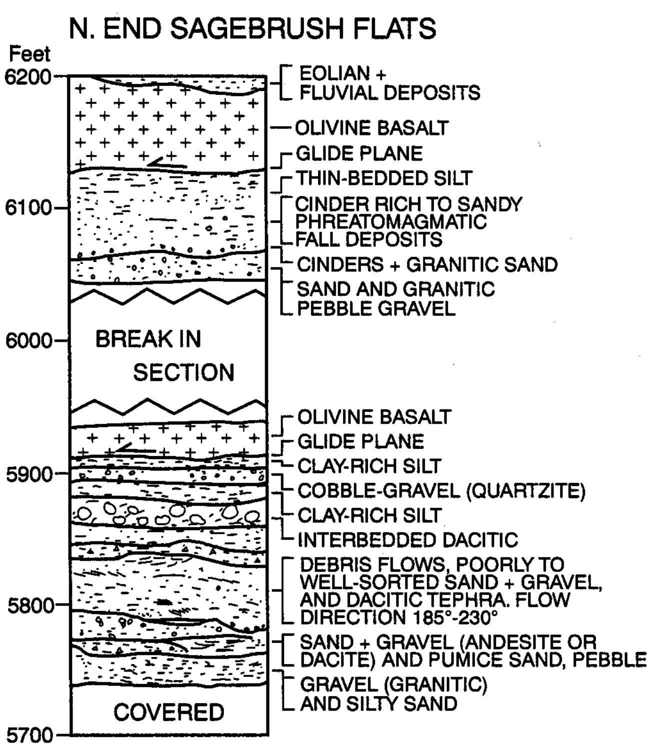

FIGURE W15—Stratigraphic section, north end of Sagebrush Flats (B, east side of the river; elevations are in feet).

Chino Mesa, Montoso Peak, and the Montoso maar may also be reached via U.S. Forest Service and powerline roads across the central Cerros del Rio. These roads should be tackled only with vehicles having high clearance and passengers with tough kidneys (see IAVECI field trip guide excursion 8A, Day 6, Stop 2, Baldridge, 1989).

Chaquehui and Frijoles Canyons

(location E, 12.7 km south of Buckman Crossing, west side of the river)

While floating down the river between these two canyons, notice the interplay between the ancestral Rio Grande and volcanic activity, exemplified by deposits that are well-exposed in the western canyon walls. There are several maar volcanoes, each 2–3 km in diameter (reconstruction); their distal surge deposits are visible, interbedded with thin tholeiitic basalt flows, many of which are overflow from crater lava lakes within the maar craters (Fig. W18). These maars and lavas were cut by drainages from the Jemez volcanic field, which in turn were filled by the thick andesitic lavas from the Cerros del Rio field, cut again by streams in resurrected canyons, partly buried

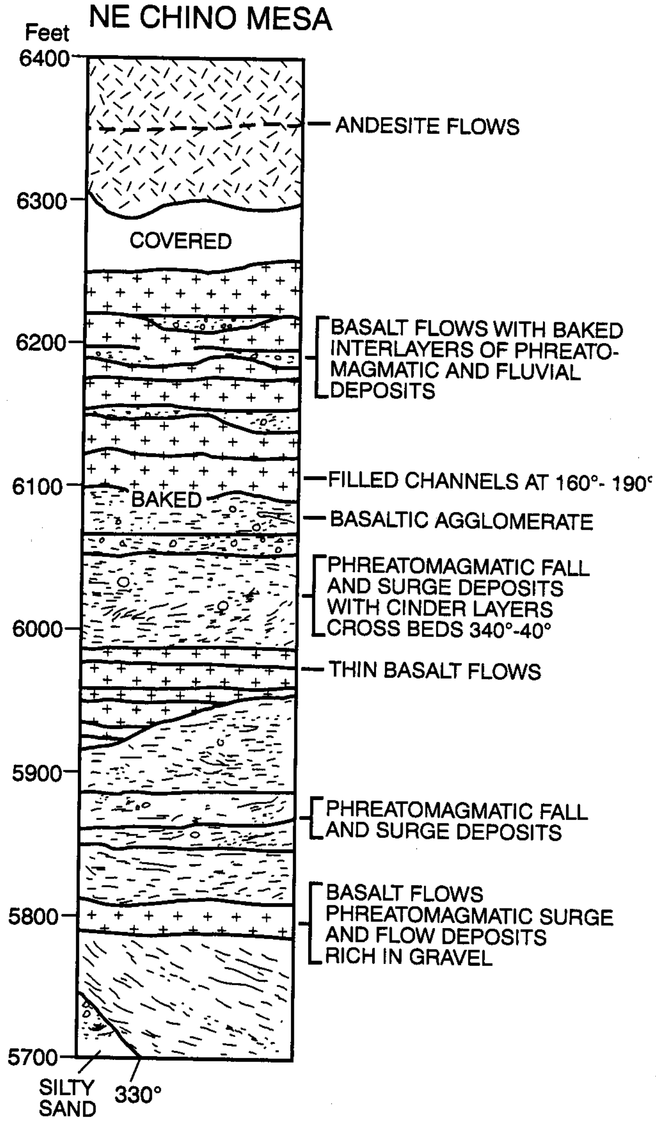

FIGURE W17—Stratigraphic section, northeast Chino Mesa (D, east side of the river; elevations are in feet). This section is representative of the stratigraphy of the Cerros del Rio.

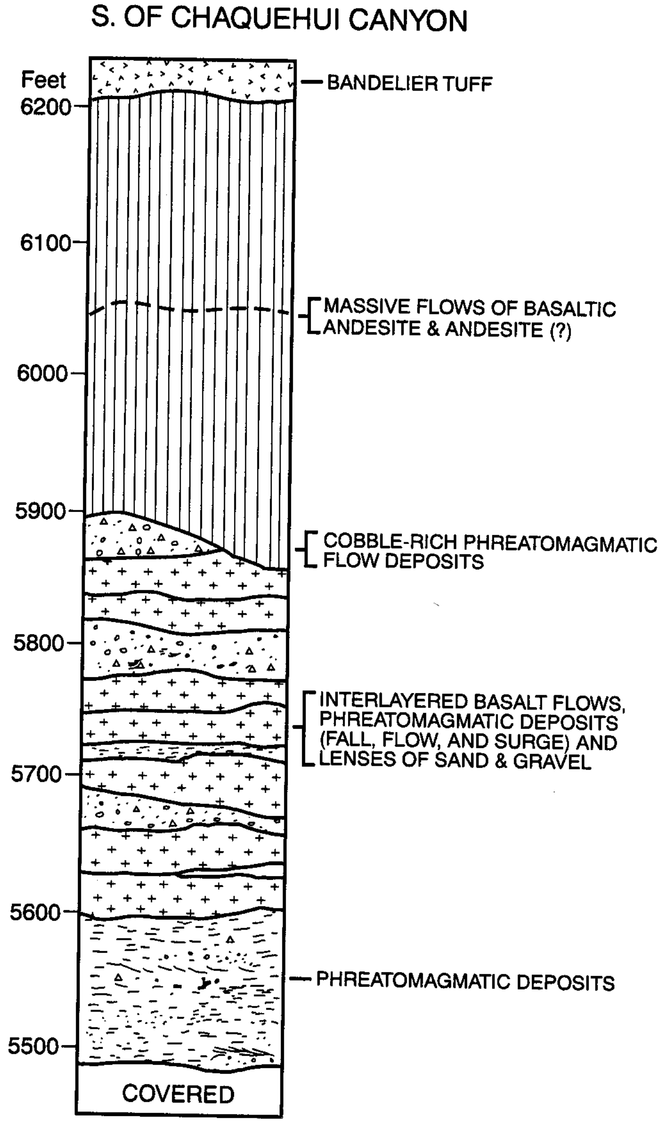

FIGURE W18—Stratigraphic section, immediately south of Chaquehui Canyon (E, west side of the river; elevations are in feet).

by canyon-filling ignimbrites of the Bandelier Tuff, then cut again by streams to form the canyons visible today.

Sidetrip up Frijoles Canyon to Bandelier National Monument Headquarters.

This well-used trail is within a national monument; **do not collect samples!** A walk of 1.5 km up (north) Frijoles Canyon to Upper Frijoles Falls will take you through a well-exposed section of these maar volcanoes. The uppermost of these maars can be walked from its distal margins at the Rio Grande to its throat at the upper falls, where the crater is filled with scoria deposits and crater lake lavas (Figs. W19, W20). This maar is about 3 km in diameter and 30 m thick at the crater rim. It is overlain by thick andesite flows that are probably from a vent located east of the river.

North of Upper Frijoles Falls (and upsection from the river), there is a cinder cone exposed to the east of Frijoles Canyon, above the flats above upper falls. This line of vents includes another large cinder cone, located along NM-4, 2.3 km north of Upper Frijoles Falls. As was mentioned earlier, those vents near river level (and the water table) formed phreatomagmatic tuff rings, whereas those about 100 m above the river were erupted as cinder cones. Another interpretation is that the cinder cones were erupted later, after deeper incision of the river canyon and lowering of the water table.

Above Upper Frijoles Falls, the canyon broadens, and you can see a well-exposed paleocanyon along the northeastern canyon wall, which is filled with upper member Bandelier Tuff ignimbrite. Frijoles Canyon was also filled with Bandelier Tuff, but much of the paleo-Frijoles Canyon has been resurrected, leaving only patches of tuff clinging to paleocanyon walls. As you continue up the canyon, the basaltic lavas of this paleocanyon are exposed along the trail.

As the trail begins to climb again, toward the Bandelier Headquarters, it passes through nonwelded ignimbrite of the upper member of the Bandelier Tuff. There are no Plinian pumice fall deposits at this location because the upper member pumice deposit is located northwest of the Valles caldera.

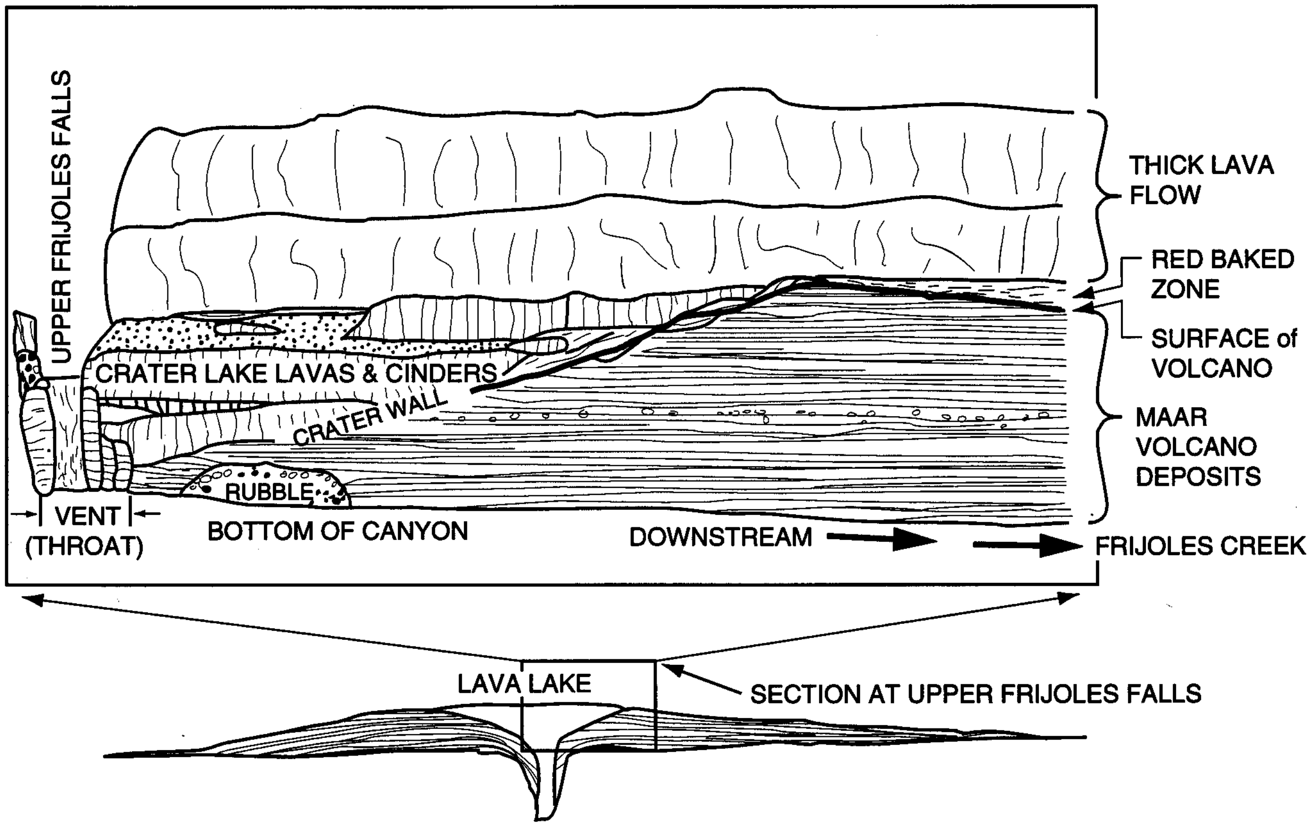

FIGURE W19—Exposures and reconstruction of the maar volcano at Upper Frijoles Falls, Frijoles Canyon (west side of the river).

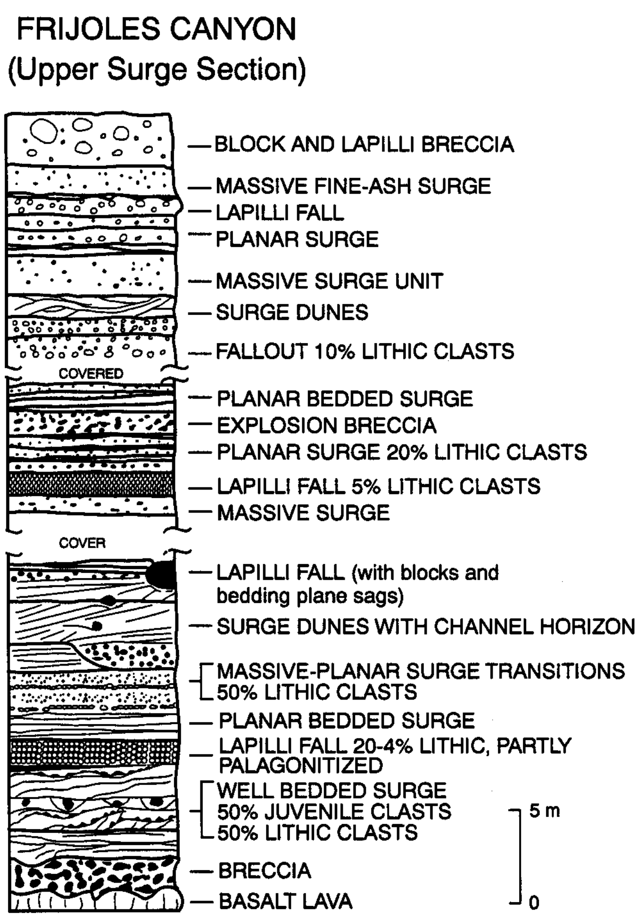

FIGURE W20—Stratigraphic section through the maar volcano at Upper Frijoles Falls, Frijoles Canyon.

When you have reached the Headquarters Visitor Center, enjoy the museum and well-preserved ruins. Ask at the Visitor Center about other geological trail guides.

Frijoles Canyon to Lummis and Alamo Canyons

(17–17.9 km south of Buckman Crossing, west side of the river)

This stretch of canyon consists of mostly basaltic and andesitic lava flows, with some interbedded proximal phreatomagmatic tuffs. These rocks were cut by canyons that were filled by ignimbrites of the Bandelier Tuff; some, but not all, of these canyons have been resurrected by erosion.

Several of the basalt flows of the Cerros del Rio are underlain by the lower member of the Bandelier Tuff (1.4 Ma) and overlain by the upper member of the Bandelier (1.0 Ma).

Arroyo Montoso enters from the east 1 km upstream from Alamo Canyon. There are thick (> 100 m), nearly continuous exposures of coarse phreatomagmatic deposits, dipping dikes, and interlayered basalt flows for 1.5 km up the arroyo from the Rio Grande. The basaltic sequence is capped by approximately 30 m of ignimbrite (lower Bandelier Tuff?), which lie beneath a glassy Cerros del Rio andesite flow.

Quaternary gravel deposits, containing clasts < 4 m in diameter, crop out 30–50 m above the Rio Grande downstream from Alamo Canyon. These gravels are older than 60 ka and may have been deposited after breaching of landslide dams located upstream.

Alamo Canyon to Sanchez Canyon

Along the canyon for 2 km downstream from Alamo Canyon, the river cuts through another maar. Diatomites are interbedded with the phreatomagmatic deposits east of the canyon. Flows of basaltic andesite and andesite from the Cerros del Rio cap these deposits. West of the river, thick exposures of proximal phreatomagmatic deposits crop out in cliff exposures and lie beneath ignimbrite (Bandelier?) and basalt flows. Interlayered mid-Pliocene cobble gravels and basalt flows lie at water level over much of this reach, and pink arkosic sandstone of the Santa Fe Group is exposed beneath the gravel at several locations.

Between Sanchez Canyon and Rio Chiquito
(22.5 km south of Buckman Crossing)

Faults extending north from the La Bajada fault (and escarpment) cross the canyon through this section. If you drive from Albuquerque to Santa Fe, you climb this escarpment on Interstate Highway 25. These faults and escarpment mark the mouth of White Rock Canyon, where the Rio Grande flows out into the Santo Domingo Basin and opens into the main part of Cochiti Reservoir. There is no evidence for movement along these faults during the Quaternary. Several basaltic vents are located along these faults, but none have been mapped in detail. Scattered patches of distal lower member (Otowi) Bandelier Tuff ignimbrite are interbeddedwith basaltic lavas and tuffs of the Cerros del Rio.

As the Rio Grande turns to the ESE (mouth of the Rio Chiquito), the cliffs on the north bank consist of a 35-m-thick section of hyaloclastite. The deposit consists of orangish-yellow tuffs and highly vesicular bombs. Large bombs increase in number relative to matrix until there is no matrix and the deposit consists of agglutinated bombs that have flowed as rootless lava flows. The largest bombs are 1.3 m x 0.3 m; all are very vesicular, with vesicles up to 1 cm in diameter. Smaller clasts are bomb fragments. The yellow-orange matrix consists of coarse ash, lapilli, and small bombs; bedding plane sags are common. The tuffs are cut by a basaltic dike with an attitude of N10E; this grades upward into a scoria deposit.

Turning toward the south again, the river cuts through another maar deposit consisting of well-bedded pale brown coarse to medium ash and lapilli. Beds are 2–10 cm thick, with ripples, dunes, and bedding plane sags. The well-bedded deposits consist of mostly lithic clasts, including fluvial cobbles and pebbles. Overlying the well-bedded tuffs is 15 m of massive to poorly bedded volcanic breccia, interbedded with yellow tuff.

Boat Ramp, Cochiti Reservoir
(27.5 km south of Buckman Crossing)

From here, it is possible to exit the canyon and river via Cochiti Pueblo, La Bajada escarpment and Interstate Highway 25.

References

Aubele, J., 1978, Geology of the Cerros del Rio volcanic field, *in* J. Hawley (ed.), Guidebook to the Rio Grande rift in New Mexico and Colorado: New Mexico Bureau of Mines and Mineral Resources, Circular 163, pp. 198–201.

Baldridge, W. S., 1979, Petrology and petrogenesis of Plio-Pleistocene rocks from the central Rio Grande rift, New Mexico, and their relation to rift structure; *in* R. Riecker (ed.), Rio Grande rift—tectonics and magmatism: American Geophysics Union, Washington DC, pp. 323–353.

Baldridge, W. S., 1984, Rio Grande rift—problems and perspectives: New Mexico Geological Society, Guidebook 35, pp. 1–12.

Baldridge, W. S., Perry, F. V., Vaniman, D. T, Nealey, L. D., Leavy, B. D., Laughlin, A. W., Kyle, P., Bartov, Y., Steintz, G., Gladney, E. S., 1989, IAVCEI Field Guide, Excursion 8A—Miocene to Quaternary volcanism and extensional tectonics, central Río Grande rift and southeastern Colorado Plateau, New Mexico and Arizona; *in* Chapin, C. E. and Zidek, J. (eds.), Field excursions to volcanic terranes in the western United States, Volume 1—Southern Rocky Mountain region: New Mexico Bureau of Mines and Mineral Resources, Memior 46, pp. 187–228.

Borton, R. L., 1968, General geology and hydrology of north-central Santa Fe County, New Mexico: *abs in* Guidebook of the San Juan–San Miguel–La Plata region: New Mexico Geological Society, Guidebook 19, pp. 210–211.

Cavazza, W., 1986, Miocene sediment dispersal in the central Española Basin, Rio Grande rift, New Mexico, U.S.A.: Sedimentary Geology, v. 51, pp. 119–135.

Chapin, C. E., 1979, Evolution of the Rio Grande rift—a summary; *in* R. Riecker (ed.), Rio Grande rift—tectonics and magmatism: American Geophysics Union, Washington DC, pp. 1–5.

Coons, L. M., and Kelly, T. E., 1984, Regional hydrogeology and the effect of structural control on the flow of ground water in the Rio Grande trough, northern New Mexico: New Mexico Geological Society, Guidebook 35, pp. 241–244.

Dethier, D., (in press), Geology of the White Rock quadrangle, Los Alamos and Santa Fe counties, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Geologic Map 73, scale 1:24,000.

Doell, R. R., Dalrymple, G. B., Smith, R. L., and Bailey, R. A., 1968, Paleomagnetism, potassium-argon ages, and geology of rhyolites and associated rocks of the Valles caldera, New Mexico *in* Coats, R. R., Hay, R. L., and Anderson, C. A. (eds.), Studies in volcanology—a memoir in honor of Howell Williams: Geological Society of America, Memoir 116, pp. 211–248.

Dransfield, B. J., and Gardner, J. N., 1985, Subsurface geology of the Pajarito Plateau, Española Basin, New Mexico: Los Alamos National Laboratory Report, LA10455–MS, 15 pp.

Duncker, K. E., 1988, Trace-element geochemistry and stable isotope constraints on the petrogenesis of Cerros del Rio lavas, Jemez Mountains, New Mexico: Unpublished MS thesis, University of Texas (Arlington), 158 pp.

Dungan, M. A., Muehlberger, W. R., Leininger, L., Peterson, C., McMillan, N. J., Gunn, G., Lindstrom, M., and Haskin, L., 1984, Volcanic and sedimentary stratigraphy of the Rio Grande gorge and the late Cenozoic geologic evolution of the southern San Luis Valley: New Mexico Geological Society, Guidebook 35, pp. 157–170.

Fisher, R. V., Heiken, G., Purtymun, W., and Wohletz, K., 1984, Cerros del Rio, New Mexico maars: American Geophysical Union, EOS, Transactions, v. 65, no. 45, p. 1145.

Fisher, R. V., Heiken, G., and Wohletz, K. H, 1986, Base Surges: International Association of Volcanology and Chemistry of the Earth's Interior, International Congress (Abst), American Geophysical Union, p. 101.

Fisher, R. V., and Schmincke, H. U., 1984, Pyroclastic Rocks: Springer–Verlag, New York, 472 pp.

Galusha, T, and Blick, J. C., 1971, Stratigraphy of the Santa Fe Group, New Mexico: American Museum of Natural History, Bulletin 144, 127 pp.

Gardner, J. N., Goff, F., Garcia, S., and Hagan, R. C., 1986, Stratigraphic relations and lithologic variations in the Jemez volcanic field, New Mexico: Journal of Geophysical Research, v. 91, no. B2, pp. 1763–1778.

Griggs, R. L., 1964, Geology and ground-water resources of the Los Alamos area, New Mexico: U.S. Geological Survey, Water-supply Paper 1753, 104 pp.

Hawley, J. W., 1978, Correlation chart 2—Middle to upper Cenozoic stratigraphic units in selected areas of the Rio Grande rift in New Mexico *in* J. Hawley (ed.), Guidebook to the Rio Grande rift in New Mexico and Colorado: New Mexico Bureau of Mines and Mineral Resources, Circular 163, pp. 239–241.

Hearne,G. A., 1980, Mathematical model of the Tesuque aquifer system underlying Pojoaque River basin and vicinity, New Mexico: U.S. Geological Survey, Open-file Report 80–1023, 181 pp.

Heiken, G. H., 1971, Tuff rings of the Fort Rock–Christmas Lake Valley Basin, south-central Oregon: Journal of Geophysical Research, v. 76, pp. 5615–5626.

Kelley, V.C., 1978, Geology of the Española Basin, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Geologic Map 48, scale 1:24,000.

Machette, M. N., 1978, Geologic map of the San Acacia Quadrangle, Socorro County, New Mexico: U.S. Geological Survey, Geologic Quadrangle Map GQ–1415, scale 1:24,000.

Manley, K., 1976, K–Ar age determinations of Pliocene basalts from the Española Basin, New Mexico: Isochron/West, no. 16, pp. 29–30.

Manley, K., 1979, Stratigraphy and structure of the Española Basin, Rio Grande rift, New Mexico; *in* R. Riecker (ed.), Rio Grande rift—tectonics and magmatism: American Geophysical Union, Washington DC, pp. 71–86.

Self, S., Kircher, D. E., and Wolff, J. A., 1988, The El Cajete Series, Valles caldera, New Mexico: Journal of Geophysical Research, v. 93, no. B6, pp. 6113–6118.

Smith, R. L., Bailey, R. A., and Ross, C. S., 1970, Geologic map of the Jemez Mountains,New Mexico: U.S. Geological Survey, Miscellaneous Geologic Investigation Map I–571, scale 1:125,000.

Turbeville, B. N., Waresback, D. B., and Self, S., 1989, Lava-dome growth and explosive volcanism in the Jemez Mountains, New Mexico—evidence from the Plio–Pleistocene Puye alluvial fan: Journal of Volcanology and Geothermal Research, v. 36, no. 4, pp. 267–291.

Waresback, D. B., 1986, The Puye Formation, New Mexico—analysis of a continental rift-filling volcanoclastic alluvial fan sequence: Unpublished MS thesis, University of Texas (Arlington), 225 pp.

Wohletz, K. H. and Sheridan, M. F., 1983, Hydrovolcanic explosions—II, evolution of basaltic tuff rings and tuff cones: American Journal of Science, v. 283, pp. 385–413.

Wohletz, K. H. and McQueen, R., 1984, Experimental studies of hydromagmatic volcanism; *in* Explosive volcanism—inception, evolution, and hazards: National Academic Press, Washington, D. C., pp. 158–169.