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The Terraces Of Cochiti Canyon: Soil Development And Relation To Tectonism In The Pajarito Fault Zone

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THE TERRACES OF COCHITI CANYON: SOIL DEVELOPMENT AND RELATION TO TECTONISM IN THE PAJARITO FAULT ZONE

by

SCOTT BENJAMIN ABY

Bachelor of Science, Geology, Humboldt State University, 1993

THESIS

Submitted in Partial Fulfillment of the
Requirements for the Degree of

Master of Science in Earth and Planetary Sciences

The University of New Mexico
Albuquerque, New Mexico

May, 1997

Dedication

This work is dedicated first and foremost to my good dog and field assistant, Rio. Since he does not read well, someday I will tie a copy of this page to a slow rabbit.
Additional thanks go out to my parents Max and Sue for a lifetime of support and lots of great money, and to Lluvia Lawyer for being so very lovable and for loving me back so well.

Acknowledgments

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My sincere personal thanks and love go out to
Les McFadden for sharing his unique outlook on soils and geomorphology and for his open-minded guidance,
to Frank Pazzaglia for many thought-provoking comments and discussions,
to Gary Smith for critical help and an indispensible on-site orientation to the geology of the field area,
to Cindy Jaramillo and Alice Quatrochi for helping me to find all the hoops I needed to jump through and prodding me when necessary,
and to everyone else who helped out along the way.

THE TERRACES OF COCHITI CANYON: SOIL DEVELOPMENT AND RELATION TO TECTONISM IN THE PAJARITO FAULT ZONE

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ABSTRACT OF THESIS

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ABSTRACT

The history of motion on splays of the Pajarito Fault Zone in the area of Cochiti Canyon in the Jemez Mountains is examined using morphometric techniques and investigations of river terrace soils and deformation. Hypsometry, sinuosity, and long-profile analysis all suggest that the main splay of the Pajarito Fault has been active in the late Pleistocene and possibly Holocene, but no calibration is available that numerically constrains this activity. Terraces were mapped and correlated based on soil development and landscape position. Correlation revealed the presence of three strath terraces between 87 and 18.5 meters above present grade and one fill terrace 14 meters above grade. Terraces are informally designated as the Ridge (~500-600 ka), Canada (~350-240 ka), Rio (~125-175 ka), and Ash (~60 ka) terraces. Age estimates were derived from comparison to dated soil

chronosequenses, a calibrated varnish-cation ratio age, and the presence of ash from the ~60 ka El Cajete eruption in a terrace deposit. Long profiles of terraces and the modern stream valley were constructed by surveying the modern floodplain, the upper surface of each terrace deposit (the terrace tread), and the of terrace deposits with underlying bedrock (the strath or the base of fill). Offset of terraces ranged from 5 to 110 meters. Rates of offset are between 0.03 and 0.13 mm/yr. Offset across two splays of the fault zone has been nearly constant through time, and faulting will presumably continue on these splays. Incision rates were calculated by measuring the height of terrace remnants above grade on apparently undeformed segments of terrace profiles. Incision rates have been nearly constant from 1.22 to ~60 ka, twice as high from ~60 ka to the present, and significantly higher during the period 1.61 to 1.22 Ma.

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INTRODUCTION

"...whoever in publishing the results of a scientific inquiry sets forth at the same time the process by which it was attained, contributes doubly to the cause of science."

- Grove Karl Gilbert

An Introduction to this Report

Philosophical Considerations

It is perhaps uncommon to begin a scientific report with a philosophical discussion these days. Space constraints on journal publishers and editors no longer allow for the beautiful language of Powell or the poetry that can be seen in Gilbert's Henry Mountains report. The subtlety of the passive voice or the imagery engendered by an allusion to historic events that may have occurred in one's field area have been lost under the tide of data that has resulted from both the proliferation of geologists and the practice of measuring the substance of a career by the weight of the resultant vitae. To some extent this change in style has advanced our science by making the research of many geologists easily available. I have benefited from this practice myself by being able to publish undergraduate research. However, while attempting to fit a 100+ page senior thesis into a less-than four page *Geology* article I realized that to do so was to entirely abandon the guidance of Gilbert as stated in the quote I have chosen as my standard. I additionally noticed that the first things I removed from my thesis were the things I did not want there in the first place--the assumptions I made, the potentially unfounded generalities, the personal opinions--all of the awkward or subjective material that weakened an argument for which I had developed a parental affection. As I found myself doing this I began to wonder how prevalent (and how valid) the practice was. We are human beings exploring an exquisitely complex Earth. Opinions are necessary, generalizations are necessary, reliance on the theories of those who are respected is necessary if we are to begin to put the geologic history of a region into an understandable context. Since these practices are an integral part of the experience of geologists, and since the philosophical framework of a geologist affects the conclusions that they draw, it is my contention that they should be an integral part of a geologic report. A Master's thesis does not have the same space limitations as a journal article and I therefore have the liberty to including this type of information.

The way I have chosen to organize this report combines, I hope, a semi-chronological treatment of the methods I used to investigate my chosen problem (in deference to G.K.'s suggestion), with a respect for the value of my readers' time. Also, with the idea of emphasizing that this research had the dual purpose of answering a specific scientific question as well as teaching me how to go about answering such a question, I occasionally use the first person. I realize that this is unconventional in a scientific context, but it is becoming more common and is even recommended by some geologists I have worked with. I believe that it will help to put the various sections in perspective if it is remembered that each of them was written *as* I was researching and/or conducting a particular phase of the research.

Organization of this Report

It was clear from my first, chance visit to Cochiti Canyon (figure 1) that the area had a complex Quaternary history. Once I had decided to make this area the object of my Master's research, the first step I took was to explore the geologic history of New Mexico as a whole and of the Jemez Mountains in particular. This investigation of the regional geology of the study area created the context within which the rest of my research was carried out. In order to create a similar context, a summary of my findings makes up the first sections of this report. Those who are familiar with the geology of the Jemez

Mountains can skip this section or use it as a reference for the later discussions.

While investigating the geologic literature generated by studies of the Jemez Mountains, I determined that the goal of my research was specifically to constrain the timing of faulting on the Pajarito Fault System in the area of Cochiti Canyon. The second section of this report discusses previous research on this fault zone. While reading these previous studies, I began to investigate the methods by which other scientists had approached similar problems in other parts of the world. Some of these methods were not applicable to my study area because preservation of necessary deposits had not occurred (e.g. trenching of fault-scarp colluvium), and others could not be used because I did not have access to certain equipment (e.g. seismic profiling). The methods that did appear to be applicable included analysis of basin morphometry, correlation of terrace remnants through the use of soil development and landscape position, and construction of paleo-long profiles. The discussions of each of these methods follow a similar pattern. I begin with an introduction to the concepts behind a particular method and any assumptions which may be necessary in its application. Since I was trying to determine which methods I would employ and which ones were inapplicable or (in my mind) fundamentally flawed, these parts of the various sections tend to be extremely critical. Nevertheless, the inclusion of each method here indicates that I believed them to have substantial redeeming value as applied to the questions I tried to address. The second portion of each method section shows how I applied the particular method to my area, and any unique problems I encountered. The results of each investigation are then given, and any conclusions I felt could be reasonably drawn are enumerated. In the morphometry section, where multiple methods are discussed and conclusions are drawn from each, I have combined the application and results sections. This method of presentation is intended to show the process by which I familiarized myself with a technique, and to give the reader some indication of both the mechanics of a particular method and my personal approach. Following this protracted discussion of the methods I used to investigate deformation in the study area, I use the evidence generated to create a cohesive Quaternary geologic history for Cochiti Canyon, compare the results of my investigation to those of other terrace studies in the Jemez Mountains, and then conclude with a summary of what has been learned.

REGIONAL GEOLOGY

OF THE

STUDY AREA

Geology of Central New Mexico

A general geologic history of this region can be constructed from previous investigations of deposits in the Jemez Mountains and the northern Rio Grande rift. The following summary emphasize events that have a direct bearing on the study area. I begin by presenting an overview of the Rio Grande Rift and north central New Mexico, and then give a more detailed account of the volcanic history of the Jemez Volcanic Field.

Opening of the Rio Grande rift in northern New Mexico began at approximately 24-30 Ma with the formation of a series of shallow, broad, internally drained basins (Baldridge and Olson, 1989; Ingersoll, *et al.*, 1990). These basins were partially filled by clastic materials derived from uplifted Precambrian and Paleozoic sedimentary, igneous, and metamorphic terranes, as well as volcanic centers in northern New Mexico and Colorado. All basin fill younger than 28 Ma is known collectively as the Santa Fe Group (Ingersoll, *et al.*, 1990). Basalts (up to 16.5 Ma) interbedded with Santa Fe Group clastic deposits heralded the beginning of Keres Group (13-7 Ma) volcanic activity in the Jemez Mountains (figure 1).

A second stage of rifting, marked by distinctly steeper dipping fault zones, began by 13 Ma (Gardner *et al*, 1986). This new style of rifting was accompanied by voluminous eruption of basalts, rhyolites, andesites, and dacites. Approximately one half the total volume of the Jemez volcanic field was erupted during this period along the Cañada de Cochiti fault zone to the west of the study area (figure 1). The rate of volcanism and rifting had decreased by 7 Ma and the Cañada de Cochiti fault zone was inactive by approximately 4.5 Ma (?). Some investigators believe that rift-bounding structures stepped inward during this period and the Pajarito fault zone became the active western margin of the Rio Grande rift in this area (Gardner *et al.*, 1986). This assertion is refuted by others who see evidence of faulting on the Cañada de Cochiti fault zone after 4.5 Ma (G.A. Smith, written communication, 1996). Between 7-2 Ma, dacites, basalts, and rhyolites of the Polvadera Group were extruded, and the volcaniclastic apron of the Puye

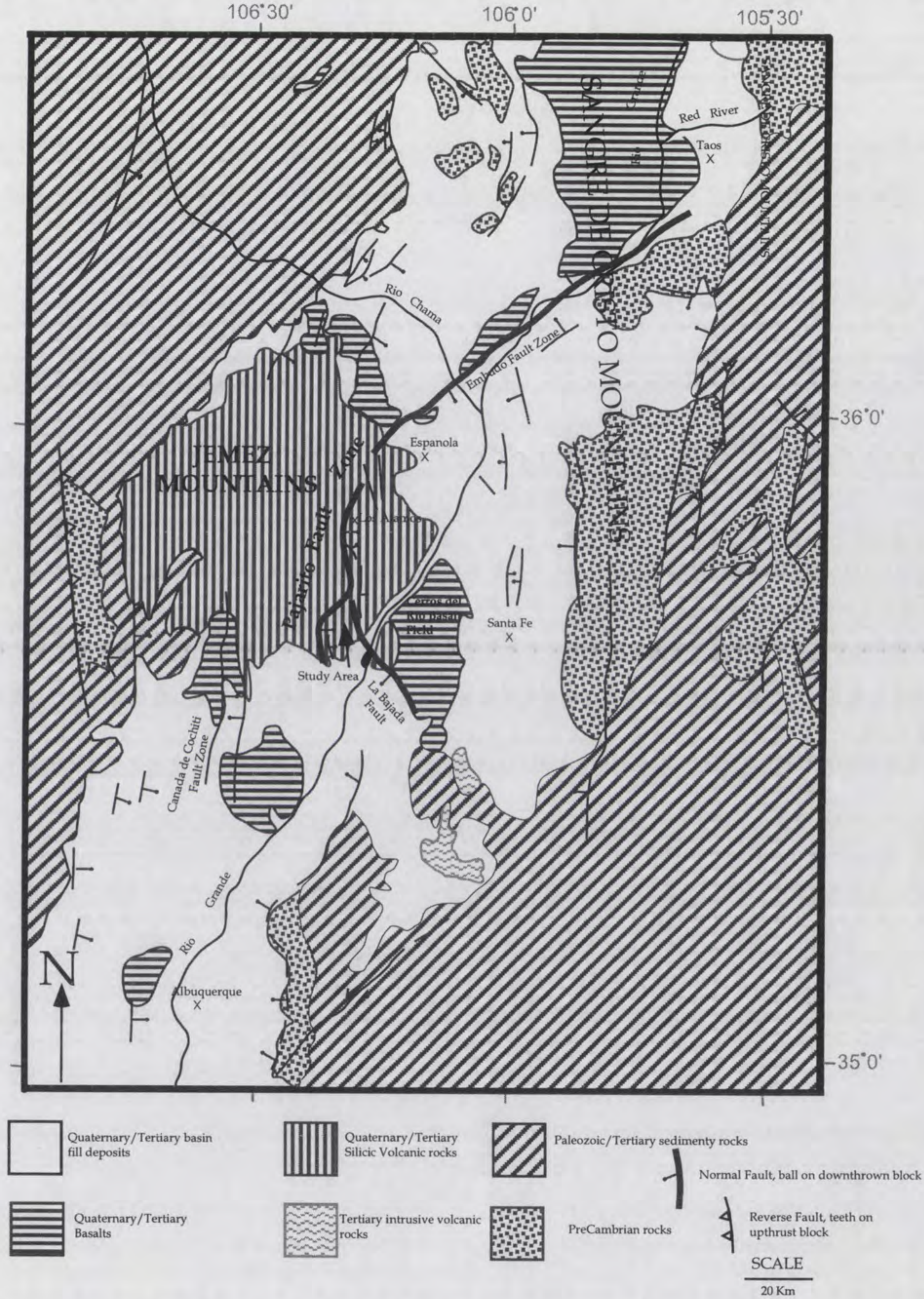


Figure 1. Generalized geology of central New Mexico and location of field area within New Mexico. Modfied from Gonzalez (1993).

Formation was shed eastward into the Rio Grande rift from the Jemez Mountains (Waresback and Turbeville, 1990). The Polvadera Group is not found in the Cochiti Canyon area.

The most recent phase of volcanic activity in the Jemez Mountains has been marked by the eruption of the rhyolitic ignimbrites of the lower (1.61 Ma) and upper (1.22 Ma) units of the Bandelier Tuff (Spell, *et al.* 1990). The combined volume of these deposits equals that of all other volcanic material in the Jemez Mountains (Gardner *et al.*, 1986). The Bandelier Tuff, various rhyolites in the Valles caldera, and the El Cajete Pumice together comprise the Tewa Group.

Upper Bandelier Tuff is offset by up to 110 meters across the Pajarito Fault in the study area. Cochiti Canyon lies within the axis of El Cajete Pumice deposition, and this unit provides a late Quaternary marker within terrace deposits. The El Cajete Pumice has previously been dated between 130-170 ka (Self *et al.*, 1988) but recent carbon-14 and thermoluminescence dates suggest eruption at approximately 60 ka (Reneau *et al.*, 1994).

An important event during the evolution of the Rio Grande system (the regional base-level control) was the addition of a large drainage area due to headward erosion through the Taos Plateau approximately 600 ka (Wells *et al.*, 1987). This increase in drainage area resulted in an increase in discharge (and therefore stream power), and would have favored accelerated stream incision since ~600 ka. Baselevel fall and incision by the Rio Grande would result in an increase in gradient at the mouth of its tributaries (of which Cochiti Canyon is one) which would then propagate upstream, causing incision in the tributaries. Gonzalez (1993) has shown that, north of White Rock Canyon, the Rio Grande has been incising for most of the late Quaternary, but periods of smaller scale aggradation have interrupted this general trend. Regional uplift is favored by Gonzalez (1993) and Dethier and Harrington (1987) as the primary driving force behind regional incision, but climate is indicated as the primary influence on the *rate* of incision through time, and in this way controls the formation of river terraces. In Cochiti Canyon, aggradation does not seem to have been as important in landscape evolution, as no major valley-fill deposits are preserved. However, periods of local baselevel stability are indicated by the presence of several terraces. Climate may have been the main variable controlling formation of these terraces as well. One other potentially important factor influencing incision in the Cochiti Canyon area is the local baselevel control exerted by basalt flows (~1.5-4 Ma) of the Cerros del Rio field found at the mouth of the canyon.

History of the Jemez Mountains

Introduction

The Jemez Mountains are a volcanic edifice created by the eruption of basaltic-to-rhyolitic composition magmas onto Precambrian to Tertiary sedimentary and igneous rocks. Although the Jemez Mountains lie along the Jemez Volcano-Tectonic Lineament (a line of volcanic centers and localized faulting stretching from eastern Arizona to the Great Plains), volcanism in this area seems to have been initiated by Rio Grande rift tectonism (Aldrich, 1986). Initial volcanism is indicated by the presence of 16.5 Ma basalts interbedded with basin-filling sediments of the Santa Fe Group near Saint Peter's Dome. The most recent volcanic activity involved a single plinian-style eruption from a vent on the southern margin of the Valle Grande approximately 60 ka (Reneau, *et al.*, 1994). Volcanic activity during this time span can be divided into three phases. Formations produced during these three episodes are grouped into the Keres, Polvadera, and Tewa Groups (in order of decreasing age). Although it is now known that the ages of rocks within each phase overlap to some extent, the division is still useful because of petrologic distinctions between rocks of the three different groups (Gardner *et al.*, 1986). Before discussing the nature and timing of each volcanic episode, a note about the formations which have been partially buried by Jemez volcanism is in order.

Pre-Jemez Volcanism Rocks

Pre-volcanic rocks in this part of New Mexico include Precambrian granites; Paleozoic marine shale, limestone, and sandstone; Mesozoic marine shales, marine and terrestrial sandstones and evaporites; and Tertiary basin-filling eolian, lacustine, and alluvial deposits. All of these units are exposed at some location in or near the Jemez Mountains. The Paleozoic and Mesozoic sedimentary rocks are predominantly fine-grained clastics and fossiliferous marine limestones. The limestones contain stringers of chert and silica-replaced fossils. Tertiary rocks are considerably more heterogeneous both within and between formations than are older rocks. In general, these rocks are coarser-grained terrigenous alluvium deposited within basins along the axis of the rift. Clasts within these deposits include a wide range of sedimentary, igneous and metamorphic rock types. Santa Fe Group deposits are by far the most common Tertiary units exposed in the Jemez Mountains, but these deposits make no substantial contribution to terrace deposits in the study area.

Keres Group

Keres Group volcanism was dominated by eruptions from vents along the Cañada de Cochiti fault zone in the central and south-central part of the present Jemez Mountains between approximately 13 and 6 Ma. Volcanic deposits of this group include the Canovas Canyon Rhyolite, Bearhead Rhyolite (including Peralta Tuff Member) , and Paliza Canyon Formation. The Paliza Canyon Formation is composed of a wide variety of rock types, but andesite dominates the formation and makes up nearly half the volume of the entire volcanic field (Self, *et al.*, 1986). Emplacement of Keres Group rocks occurred as hypabyssal intrusion, localized vent eruptions, and during construction of (presumably) overlapping composite volcanoes. Sedimetary deposits shed from Keres-group highlands are defined as the Cochiti Formation (Bailey, *et al.*, 1969). The Cochiti Formation therefore contains rocks from all of the three Keres Group units and generally reflects the volumetric dominance of Paliza Canyon Formation andesites. Some confusion has arisen due to inconsistent and contradictory definitions of the Cochiti Formation. Smith and Lavine (1994) have attempted to resolve this issue, but the nomenclature of these deposits has little bearing on the Quaternary geology of Cochiti Canyon, and will not be discussed here.

Polvadera Group

Polvadera Group rocks dominate the northern and particularly the northeastern part of the Jemez Mountains. The oldest formation in this group is the Lobato Basalt. This unit is contemporaneous with basalts of the Paliza Canyon Formation and forms the caprock of Lobato Mesa in the northeastern Jemez Mountains. The somewhat younger Tschicoma Formation is composed of 3-7 Ma dacites and minor andesites emplaced as flows and domes . This formation is second only to the Bandelier Tuff (discussed below) in present aerial extent (see map of Smith, *et al.*, 1970). The Tschicoma formation is interbedded with the volcanogenic alluvium of the Puye Formation. This 7 to

1.7 Ma unit was deposited as an alluvial-fan complex by streams flowing to the east away from mountains created by Tschicoma Formation volcanic activity (Waresback and Turbeville, 1990). The third volcanic deposit of the Polvadera Group is the El Rechuelos Rhyolite, which forms several small domes west and north of Polvadera Peak. Although not formally included in the Polvadera Group, basalts of El Alto, Santa Ana Mesa, and Cerros del Rio were erupted around the southern, northern , and eastern margins of the Jemez Mountains during Polvadera time.

Tewa Group

Volumetrically, the Tewa Group is dominated by 600 cubic kilometers of high-silica rhyolite (Heiken *et al.*, 1990) of Bandelier Tuff (combined upper or Tshirege and lower or Otowi Members). Pre-Bandelier Tewa Group deposits include two domes within the Toledo Embayment known as the Cerro Rubio Quartz Latite, and the San Diego Canyon ignimbrites (Heiken *et al.*, 1990) exposed in the north rim of Valles Calderaand in San Diego Canyon. Pumice derived from San Diego Canyon ignibrite eruptions was reworked and deposited near the mouth of Cochiti Canyon. The Bandelier Tuff was deposited following two cataclysmic eruptions from the Valles (1.22 Ma) and Toledo (1.61 Ma) calderas (Spell *et al*, 1990). These two calderas are essentially superimposed on each other. The Toledo embayment, once thought to be the center for the lower Bandelier eruption, is now believed to be an older, unrelated topographic feature (Heiken *et al.*, 1990). The Cerro Toledo Rhyolite consists of tuffs and domes erupted from vents within the Toledo Embayment and to the south as far as Rabbit Mountain (*Smith et* al, 1970). Eruption of these domes and tuffs seems to span the period from just prior to the lower Bandelier eruption (1.61 Ma) to just before the upper Bandelier eruption (1.22 Ma).

Valles Rhyolite is composed of domes, flows, and tuffs of high-silica rhyolite erupted within the Valles Caldera after the upper Bandelier eruption. These eruptions occurred between 1.04 and 0.365 Ma and were associated with both resurgent dome formation and moat volcanism (Gardner *et al.*, 1986). Resurgence since eruption of the Bandelier Tuff has raised the central part of the caldera to an elevation of 3430 m at Redondo Peak. The mechanisms of resurgence are not well understood, and the radius of the zone of uplift associated with resurgence has not been well documented (Francis, 1993, p. 298). The youngest eruptions in the Jemez Mountains were centered at a vent on the southern rim of Valles caldera and lead to deposition of the El Cajete Pumice, Battleship Rock Ignimbrite, and Bonco Bonito Obsidian (Self, *et al.* 1988). The Battleship Rock Ignimbrite and Bonco Bonito Obsidian are confined to the upper reaches of San Diego Canyon where this canyon forms the only drainage out of the Valles caldera. The axis of dispersal of El Cajete Pumice stretches from the vent site to the southeast toward Cochiti Pueblo. These eruptions have recently been dated at approximately 60 ka, based on radiocarbon ages derived from trees killed by blasts from the El Cajete crater (Reneau, *et al.,* 1994).

THE PAJARITO FAULT ZONE

Introduction and Outline of the Problem

The Pajarito Fault Zone (or Fault System) extends for more than 50 kilometers along the eastern edge of the Jemez Mountains (figure 1). Faults within the zone are normal or oblique-normal and dominantly eastward dipping. Previous studies of the fault zone have focused on either the potential risk to Los Alamos National Laboratory (e.g. Wasch, *et al*, 1988, Gardner and House, 1987) or on the role the fault plays in the kinematics of the region (e.g. Golembek, 1983; Baldridge and Olson, 1989). Studies of potential seismic hazard have shown that the fault is "capable" as defined by government regulations (has had recurrent movement within 0.5 Ma or movement within 0.035 Ma) but have been hampered by a lack of age control on Quaternary fluvial and hillslope deposits. Kinematic studies have shown that the fault zone is related to opening of the Rio Grande Rift (Baldridge and Olson, 1989). Long-term offset along various splays of the fault zone can be calculated by measuring offset in the upper Bandelier Tuff (1.22 Ma, Spell, *et al.*, 1990). Offset on major splays of the fault system ranges between 60 and 200 m (Golembek, 1983; Gardner and House, 1987), yielding offset rates of 0.05 to 0.17 mm/yr. The question remains: *When* during the last 1.22 My have faults within the zone been active? Cochiti Canyon offers some hope of resolving this question because it is one of the only places along the fault zone where several spatialy distinct Quaternary fluvial deposits are preserved. These deposits are located on the hanging-wall block of the main splay of the fault zone and are cut by at least two other splays (figure 2, 3, appendix 1). If some age constraints can be placed on terraces/ terrace deposits and these terraces have been deformed by fault movements, then it will be possible to create a history of movement on individual faults *during* the last 1.22 million years.

Cochiti Canyon is the focus of this study because it appears to have the most extensive suite of terrace deposits close to the main splay of the Pajarito Fault Zone. The drainage basin of Cochiti Canyon covers 78 km^2 and the trunk stream flows for ~27 km in a southeasterly direction (figure 4). The upper 16 km of the Rio Chiquito (the stream that flows within Cochiti

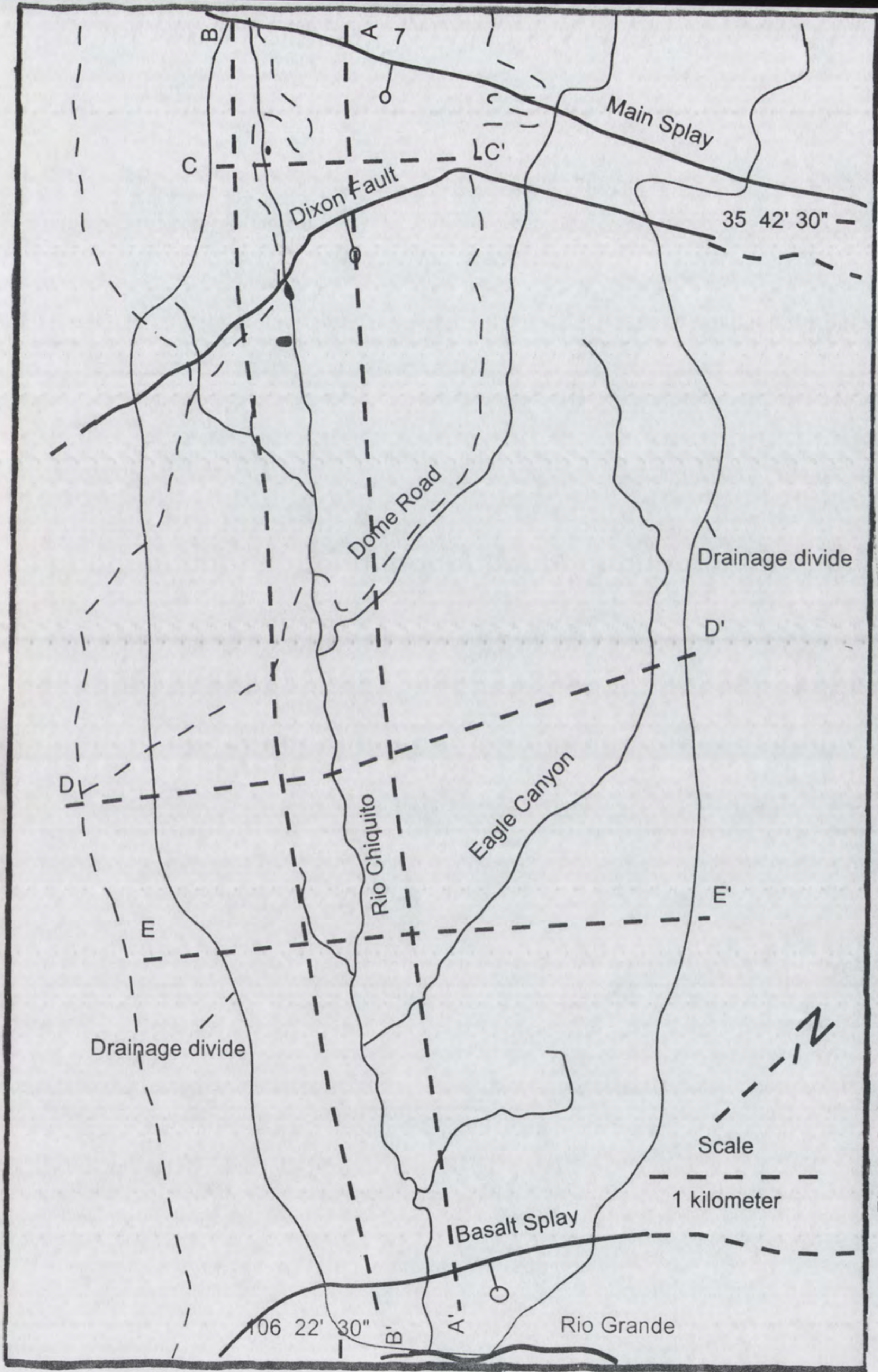


Figure 2. Map of study area showing locations of cross sections (heavy dash), faults (thick lines), roads (thin dash), streams, and outline of drainage basin. See appendix 1 for topographic referenc.

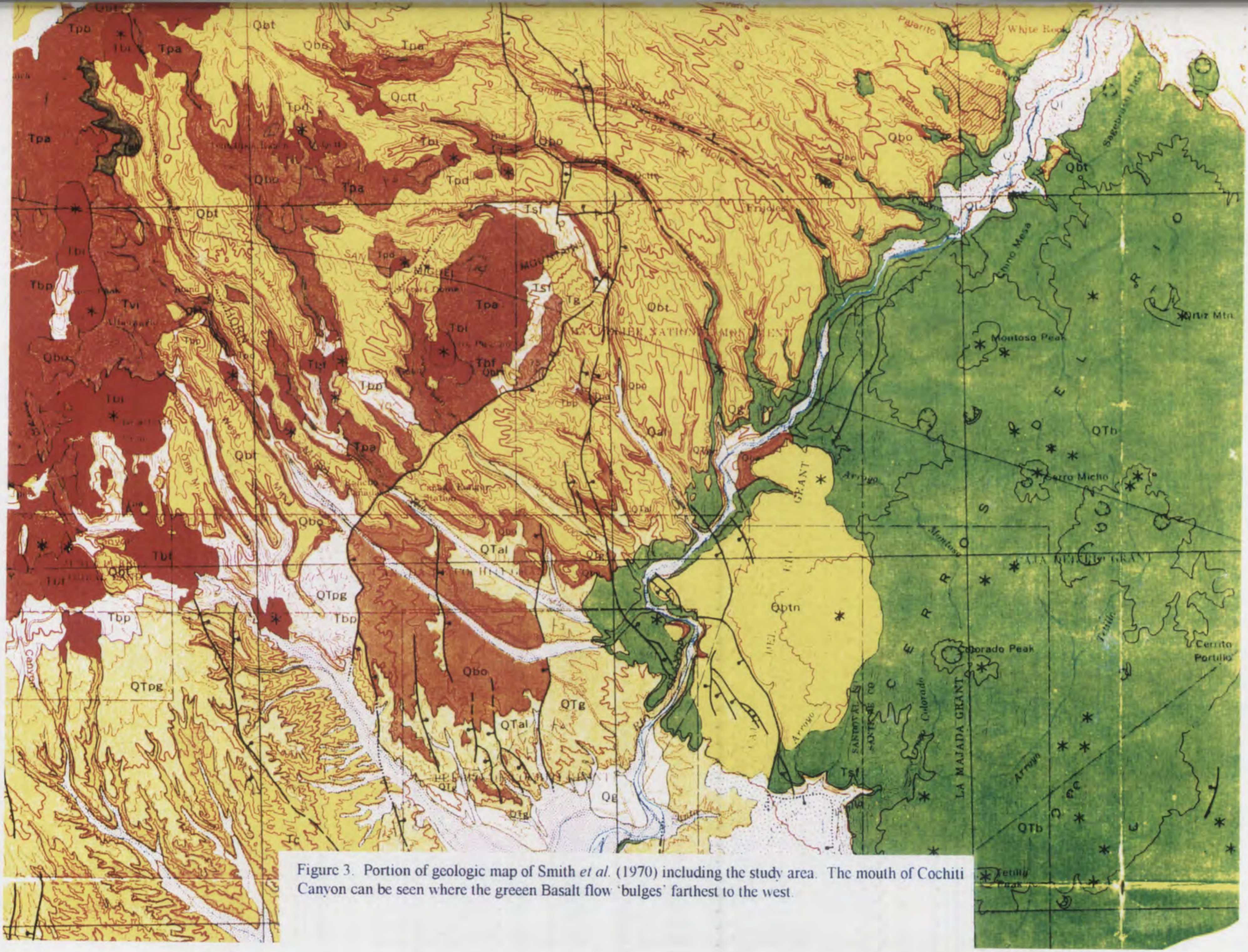


Figure 3. Portion of geologic map of Smith *et al.* (1970) including the study area. The mouth of Cochiti Canyon can be seen where the greeen Basalt flow 'bulges' farthest to the west.

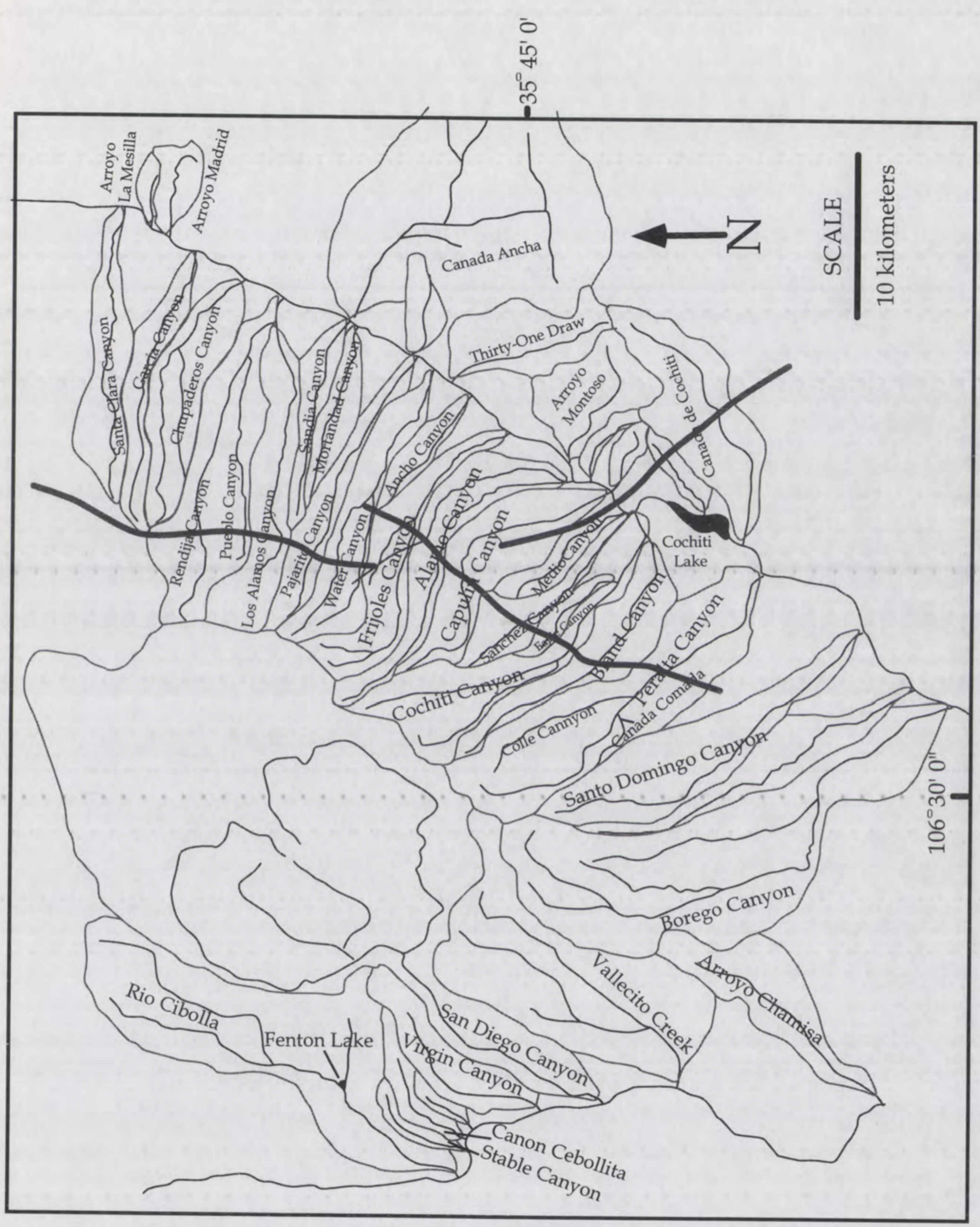


Figure 4. Map of drainge basins in the Jemez Mountains analyzed during this study. Drainage divides, streams, and major faults were taken from U.S.G.S. 1:100,000 scale topographic maps.

Canyon) are confined within a 200-300 meter deep, 0.2-1.2 km wide canyon incised into cliff-forming Bandelier Tuff and more or less slope-forming units of the Keres Group. This canyon is separated from those to the northeast and southwest by mesas formed in Bandelier Tuff and capped by a thin, discontinuous veneer of reworked ~60 ka El Cajete Pumice Bandelier-derived sand, and/or colluvial/alluvial deposits derived from volcanic deposits that rise above the level of the Bandelier Tuff . The lower 11 km of the stream flow on the downthrown block of the main splay of the fault zone. As the Rio Chiquito flows onto the hanging wall of this fault, the canyon broadens to 1.5 km and the first fluvial terraces occur on the northeast side of the canyon, although the drainage divide is still delineated by mesas formed in Bandelier Tuff for another kilometer downstream. Below this point the Rio Chiquito flows through an alluvial valley flanked on either side by one to four distinct fluvial terraces as well as outlying buttes of Bandelier Tuff. The last 2.3 km of the Rio Chiquito's course plunge through a narrow gorge cut into a ~1.7 Ma basalt flow of the Cerros del Rio before entering the Rio Grande near the upper reaches of the man-made reservoir called Cochiti Lake (figure 4).

Previous Fault Studies

As mentioned previously, earlier studies of the Pajarito Fault Zone have focused on either the regional kinematic significance of the system, or on the seismic risk posed to Los Alamos National Laboratory. The seismic hazard study of Gardner and House (1987) incorporated previous work and serves as a comprehensive review of information up to the date of the publication. In this report, Gardner and House (1987) divide the fault zone into three geographical segments, the southern, central, and northern. These segments are discussed below and all information is from Gardner and House (1987) unless otherwise noted.

The Southern Segment

This segment includes my study area and extends northward to the southern boundary of Los Alamos County. Maximum displacement of the upper Bandelier Tuff across the main splay of the fault zone is approximately 200 meters (a rate of .02 mm/yr. for the last 1.22 My) on the mesa south of Frijoles Canyon (figure 4), and total displacement across all splays in this same area is 290 m. According to this study, displacement of pre-Bandelier rocks "...easily exceeds 300 m" (p. 11) based on juxtaposition of different-age units on either side of the fault. All displacement indicators measured show normal displacement (see their figure 3). At the mouth of Bland Canyon, the main splay of the fault zone is exposed in a cut-bank exposure and juxtaposes post-Bandelier alluvium against 6.81 Ma Peralta Tuff (Gardner, *et al.,* 1986). The fault plane dips ~70 degrees to the southeast and displaces alluvium by 6 m. Three parallel seismic lines were run near this location and appear to confirm the displacement seen in outcrop, and also showed an additional 3 m of vertical offset. However, my own observations indicate that the alluvium on the upthrown side of the fault is at most 1.5 meters thick while the seismic lines indicate up to 3 m. This observation indicates that the seismic velocities used to calculate thicknesses on these lines may be inaccurate. Assuming that all of the offset observed in outcrop (6 m) occurred during a single faulting event yields an estimate of one, magnitude 7.8 earthquake (method of Slemmons, 1977, annotated in Gardner and House, 1987).

The Central Segment

This segment includes all of the fault zone within Los Alamos County. Displacement of the Bandelier Tuff is up to 120 meters in this segment, but the fault zone is considerably more complicated than in the southern segment. For example, numerous down-to-the-west faults are observed in the northern part of the central segment, and stratigraphic offsets and slickenside data (see p. 26, and figure 3 of Gardner and House) argue for some substantial strike-slip movement in this segment. Post-displacement alluvial fans have been abandoned and incised along this segment, and a line of foliage is seen crossing a small landslide on the fault scarp. Varying displacement of pre-Bandelier units and Bandelier tuff indicate recurrent movements. Discontinuities in stream profiles in Water, Del Valle and Pajarito canyon show a steepening of gradients upstream of the fault zone. Gradient changes here are on the order of 1-3 degrees. Using minimum incision rates calculated based on total incision of the Bandelier Tuff since its emplacement indicate that these gradient changes are the result of 60-110 m of displacement in the last 0.5-0.3 m.y. (rates of .02-.04 mm/yr). Seismic refraction profiles across down-to-the-west faults indicate damming of 2.5 to ~3 m of alluvium on the upstream (downthrown) side of these faults, but no age control is provided.

The Northern Segment

The northern segment extends from the northern boundary of Los Alamos County to the northern tip of the Embudo Fault Zone (figure 1). Within this segment, evidence for offset includes: displacement of a ~350 ka geomorphic surface by 50 m, ~.2 m displacement of an apparently young cut-and-fill sequence in a ~350-240 ka terrace, and apparent displacement of a ~42 ka paleochannel by 0.5 m.. Much of the movement on this segment is right lateral and vertical displacements do not therefore give a good indication of fault activity.

In addition to the thorough discussion of evidence for the recency of faulting, Gardner *et al.* (1987) also provide a list of historic earthquakes that may have been felt in the Los Alamos area, and a map showing microseimicity in the region from 1973 to 1984. The list of historic earthquakes does not bear directly on this study but demonstrates the lack of significant, historical macroseismicity. The microseismic data show the highest concentration of small earthquakes along the Nacimiento range-front (their figure 10, 11), with other concentrations along the Embudo Fault Zone and to the west of Albuquerque. The lack of a large number of earthquakes along the Pajarito Fault Zone on this plot either indicates that the fault zone is inactive or that strain is currently accumulating here.

Objective of this study

As stated in the introduction to this section, the purpose of this study has been to determine when the Pajarito Fault zone has been active since emplacement of the Bandelier Tuff. The answer to this question is relevant to seismic hazard assessment, but touches on more purely academic issues as well. For instance: Is the Rio Grande rift an active structure or is it a failed rift? (Baldrige, *et al*, 1989), How do previously existing fault zones respond to cataclysmic volcanic events and subsequent resurgence?, Do such cataclysmic events initiate tectonism?, How do streams respond to the interplay between tectonism, volcanism, and climate change? Although I have no illusions of resolving all these questions with this one study, determining the timing of tectonism on the Pajarito Fault Zone during the Quaternary provides one essential piece for each of these puzzles.

METHODS

Basin Morphometry

Introduction

Morphometry is the study of the size and/or shape of landscape elements and the relationship between various quantitative parameters of drainage basins (Ritter, 1978). The quantities measured can be broken into three categories: linear, areal, and relief parameters. Linear parameters give some measure of the "plan view" characteristics of a basin. This category includes stream length within each order (stream orders were defined using the Strahler method, see Ritter (1978) for discussion, bifurcation of one stream order to another, total number of streams in each order, and total length of streams in each order or within the basin (see appendix 2 for ,mathmatical descriptions of all morphometric parameters used in this report). Areal parameters measure some two-dimensional characteristic of a basin such as basin area, basin shape, length of steams per unit area (drainage density), or number of streams per unit area (stream frequency). Relief parameters define the three-dimensional relationship of basin characteristics. One commonly defined relief parameter is basin hypsometry. This parameter is a measure of the proportion of a basin which lies within some elevation range, and is usually shown by drawing a hypsometric curve (see discussion below).

The hypothesis behind the study of morphometry is that basins with different primary controls (tectonism, climate, basin size, lithology, ...) will have correspondingly different morphometric characteristics. If this hypothesis is correct, then one need only control all factors except one (e.g. tectonic activity) to asses the impact of that particular variable on the landscape. Control of all other factors is accomplished by careful selection of basins, since one can obviously not run a 1:1 scale experiment for millions of years. Any conclusions drawn from morphometric analysis are dependent on an accurate inference being made between the morphometric parameter in question and some landscape-shaping process. This is a critically important assumption that must be kept in mind when evaluating the results of morphometric analysis.

In practice, the analysis of morphometry is complicated by the inability to truly control all landscape-shaping factors except one. This difficulty increases in proportion to the resolution with which one examines the variability between basins (i.e. the closer you look the more variability you will find within basins which at first seem to have identical controlling factors). One way to overcome this problem may be to increase the accuracy of measurement, but a point of diminishing returns is rapidly approached as morphometric analysis is extremely time-consuming. Further error/uncertainty is introduced by the way in which data is collected. First, one must question the accuracy of the maps used. U.S.G.S. topographic maps are the most commonly used database, and these have a sizable allowable error. Adjoining maps may not have been made using the same technique, at the same time, or by the same individual(s). The person creating the maps will be subject to all of the inaccuracy inherent in being a human performing a mechanical task. Some other human (with all the same sources of error/inconsistency) will then be responsible for measuring hundreds of individual stream lengths, basin areas, basin perimeters, and other basin parameters. Despite these sources of inaccuracy, some useful generalizations can often be made, and basins that have different morphological controlling factors can be compared.

Application and Results

A number of different morphometric parameters where examined (see appendix 1 for definitions of parameters used in this report). Each of these parameters was hypothesized to have some relation to fault activity, but it was not possible to know which one would be most effective. The discussion below is divided into sections that each describe how a specific technique was applied to this particular area. The techniques are arranged in the order in which they were applied, as information gained from one analysis helped me to decide what technique to try next. I have included both successes and failures in the hope that my experience might allow future workers to more quickly focus their efforts.

Morphometric Analysis Using 1:100,000 Scale Maps

In an attempt to understand/quantify the effect of the Pajarito Fault Zone on landscape evolution (and thus aid in determining the timing of movement on the fault), basin area, stream lengths within each order, and drainage density were measured from "blue-line" drainage nets constructed using 1:100,000 scale maps of large portions of the Jemez Mountains (table 1). From these parameters, morphometric properties of basins were calculated and compared (figure 4). Many different combinations of data from this set were plotted, however, most graphs showed no systematic covariation. A few of the plots that did indicate some consistent relation between parameters are shown in figure 5. Several general trends can be seen, such as increasing number of streams per order with increasing basin size and decreasing drainage density with increasing basin size. These trends have been shown for streams in general many times (e.g. Ritter, 1978). This analysis gave little

Name of Basin	Total length of Streams by Order: First order	Second Order	Third Order	Total length of all Streams	Average Basin Area	Basin Area Above Fault*
Santa Clara	17.42	17.45	8.76	34.87	59.71	
Garcia	12.69	16.36	0.00	29.36	33.23	3.20
Chupaderos	34.44	15.07	0.31	65.53	43.22	N.A.
Rendija	52.64	16.00	16.02	68.64	83.08	34.12
Los Alamos	35.10	20.93	0.00	56.03	64.51	28.04
Sandia	14.61	2.69	0.00	17.30	14.06	N.A.
Mortandad	25.69	4.63	0.00	30.32	26.54	N.A.
Pajarito	18.38	11.85	0.00	30.23	33.16	16.65
Ancho	16.53	3.71	0.00	21.57	17.54	N.A.
Water	88.96	17.89	1.33	106.85	50.75	22.28
Frijoles	16.64	18.68	0.00	35.32	50.61	34.43
Alamo/Lucero	32.71	16.60	0.00	49.31	49.65	24.52
Capulin	29.20	16.94	0.00	46.14	50.76	26.76
Medio	13.47	1.50	0.00	14.97	16.83	5.30
Sanchez	12.35	11.75	0.00	35.22	19.33	14.10
Cochiti	43.45	20.95	11.12	64.40	77.81	55.69
Bland	23.91	19.79	0.00	54.61	44.78	24.55
Peralta	65.25	23.41	10.91	97.05	120.31	87.79
Santo Domingo	52.43	20.80	8.39	73.23	79.55	N.A.
One south of SD	23.68	13.16	0.00	71.97	41.40	N.A.
Borego	108.70	35.67	35.13	150.34	415.81	N.A.
Arroyo Chamisa	21.04	17.52	5.97	62.46	49.85	N.A.
Vallecitos	108.80	26.19	23.90	193.37	186.97	N.A.
San Diego	221.11	170.78	58.38	391.89	1032.83	N.A.
Virgin	30.61	20.80	0.00	51.41	46.20	N.A.
One north of V	6.22	3.79	0.00	25.19	7.78	N.A.
Cebolita	23.58	3.23	15.18	26.81	23.72	N.A.
Stable	7.07	3.72	0.00	10.79	10.00	N.A.
Arroyo la Mesilla	4.41	0.00	0.00	4.41	4.41	3.27
Arroyo Madrid	11.70	2.39	0.00	14.09	22.06	14.71
One north of 31	6.70	1.89	0.00	8.59	8.59	7.73
Thirtyone Draw	16.02	8.93	0.00	24.95	26.50	31.86
Arroyo Montoso	20.65	9.42	1.55	31.62	30.07	39.59
One south of AM	6.47	0.00	0.00	6.47	6.47	8.74
Two north of CdC	6.77	0.00	0.00	6.77	6.77	8.89
One north of Cdc	6.86	1.51	0.00	8.37	12.36	8.72
Canada de Cochiti	34.26	14.96	3.99	53.21	84.35	58.84

Table 1. Data generated from 1:100,000 scale topographic maps. See appendix 1 for definitions of parameters. N.A. indicates a parameter that is not applicable to a particular basin.

*Area of basin measured from the point where the stream crosses the main splay of the Pajarito Fault Zone

Name of Basin	**Number of Streams by (Order** First Order	Second Order	Third Order	**Drainage Density**
Santa Clara	4.00	2.00	1.00	0.58
Garcia	3.00	1.00	0.00	0.88
Chupaderos	6.00	2.00	1.00	1.52
Rendija	14.00	4.00	1.00	0.83
Los Alamos	5.00	1.00	0.00	0.87
Sandia	2.00	1.00	0.00	1.23
Mortandad	3.00	1.00	0.00	1.14
Pajarito	3.00	1.00	0.00	0.91
Ancho	3.00	1.00	0.00	1.23
Water	5.00	2.00	1.00	2.11
Frijoles	4.00	1.00	0.00	0.70
Alamo/Lucero	5.00	2.00	0.00	0.99
Capulin	7.00	2.00	0.00	0.91
Medio	2.00	1.00	0.00	0.89
Sanchez	4.00	1.00	0.00	1.82
Cochiti	10.00	2.00	1.00	0.83
Bland	8.00	1.00	0.00	1.22
Peralta	17.00	2.00	0.00	0.81
Santo Domingo	14.00	2.00	1.00	0.92
One south of SD	4.00	1.00	0.00	1.74
Borego	32.00	6.00	1.00	0.36
Arroyo Chamisa	7.00	2.00	1.00	1.25
Vallecitos	31.00	7.00	2.00	1.03
San Diego	84.00	17.00	3.00	0.38
Virgin	11.00	1.00	0.00	1.11
One north of V	3.00	1.00	0.00	3.24
Cebolita	4.00	1.00	0.00	1.13
Stable	2.00	1.00	0.00	1.08
Arroyo la Mesilla	1.00	N.A.	N.A.	1.35
Arroyo Madrid	2.00	1.00	N.A.	1.50
One north of 31	2.00	1.00	N.A.	1.11
Thirtyone Draw	6.00	1.00	N.A.	0.83
Arroyo Montoso	7.00	2.00	1.00	0.76
One south of AM	1.00	N.A.	N.A.	0.74
Two north of CdC	1.00	N.A.	N.A.	0.76
One north of Cdc	12.00	1.00	N.A.	1.42
Canada de Cochiti	9.00	2.00	1.00	1.43

Table 1 continued.

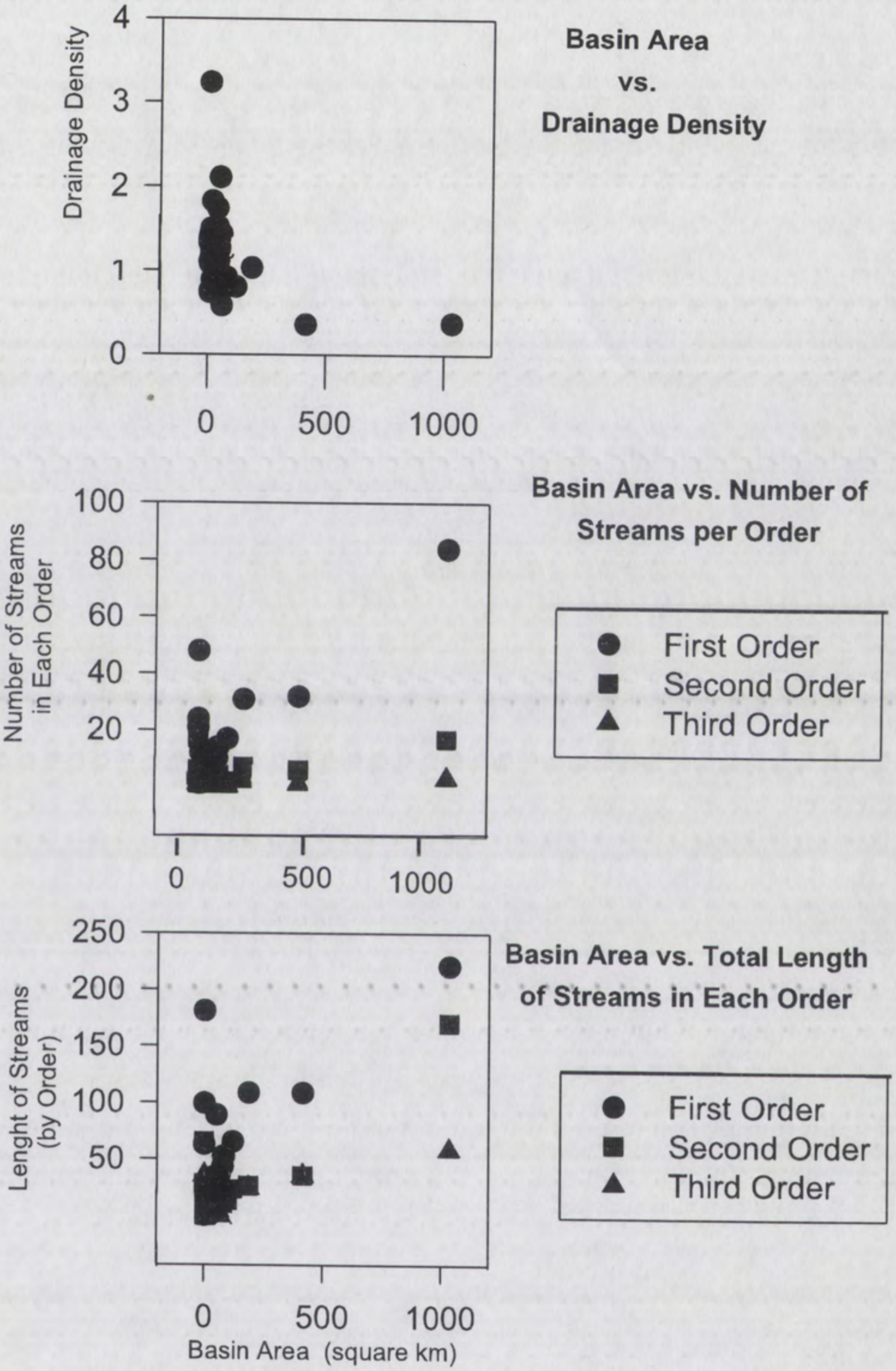


Figure 5. Plots of selected morphometric parameters calculated using 1:100,000 scale maps.

insight into the specific effect of tectonics on basin parameters because a number of controlling variables were not held constant and because of the coarse scale used.

One additional observation can be made from simply observing the shape of drainage basins in figure 4. Basins throughout the Jemez Mountains are elongated in a radial pattern about the Valles Caldera. This observation may indicate that the basins are relatively "youthful" (Ritter, 1978). During basin formation, headward erosion in the direction of regional slope originally dominates, with integration of smaller basins taking place later. It is not often possible to give a quantitative definiton of 'young' in relation to landscapes, but all of the basins of the eastern Jemez Mountains have formed since eruption of the upper Bandelier Tuff 1.22 million years ago.

Morphometric Analysis Using 1:24,000 Scale Maps

To increase the resolution of analysis, drainage nets were created for *selected basins* from 1:24,000 scale topographic maps. Basins were selected that are generally comparable to those in or near the study area with respect to rock-type and basin size (table 2). These two characteristics were hypothesized to exert the greatest control on basin morphometry. Basin size has been shown to be proportional to discharge (Ritter, 1978), and rock-type has obvious effects on landscape evolution through its control on erodability (e.g. Bull, 1991). In an area with the climatic and lithologic parameters characteristic of the Jemez Mountains, tens of square kilometers of basin area can mean the difference between a perennial stream and one which flows for only a few months each year. Within the study area this is illustrated by Bland and Cochiti Canyons. Observations and conversations with individuals familiar with the area indicate that Bland Canyon, with a basin area of 44.8 km^3, usually

Parameter*	**Basin Name** Cochiti	Bland	Virgin	Capulin	Water	Alamo
Basin Area	78.00	45.00	46.00	51.00	51.00	16.00
Area Above Fault	56.00	25.00	0.00	34.00	22.00	
Number of streams by order:						
first	805.00	594.00	427.00	632.00	570.00	359.00
second	139.00	116.00	71.00	116.00	103.00	58.00
third	24.00	27.00	9.00	25.00	19.00	8.00
fourth	5.00	6.00	1.00	6.00	5.00	2.00
fifth	2.00	1.00	N.A.	2.00	1.00	1.00
sixth	1.00	N.A.	N.A.	1.00	N.A.	N.A.
Average length of streams by order:						
first	0.26	0.07	0.35	0.23	0.24	0.35
second	0.41	0.45	0.54	0.40	0.34	0.56
third	1.70	0.90	2.50	1.26	1.40	1.70
fourth	3.60	1.40	23.00	2.40	2.00	9.00
fifth	8.70	20.80	N.A.	9.70	15.00	3.50
sixth	N.A.	N.A.	N.A.	1.20	N.A.	N.A.
Drainage density	4.60	5.77	5.02	5.10	4.50	3.92
Bifurcation of streams by order:						
1 to 2	5.80	5.10	6.00	5.40	5.50	6.20
2 to 3	5.80	4.30	7.90	4.60	5.40	7.30
3 to 4	4.80	4.50	9.00	4.20	3.80	4.00
4 to 5	2.50	6.00	N.A.	3.00	5.00	2.00
5 to 6	1.00	N.A.	N.A.	2.00	N.A.	N.A.
Basin Shape	3.40	2.25	2.30	2.43	2.13	0.64

Percent of listed Lithology (by area) for selected basins

Formation / **Canyon**	Bandelier	Peralta	Various Rhyolites	Paliza Canyon	Quaternary Alluvium	Miscellan-eous	Tchicoma
Cochiti	71.00	4.30	4.60	13.40	6.70	0.00	0.00
Bland	56.70	16.10	5.97	0.80	10.80	9.20	0.00
Alamo	77.00	0.00	8.97	5.30	0.82	8.10	0.00
Virgin	89.80	0.00	0.00	1.20	0.00	8.95	0.00
Water	72.20	0.00	0.00	0.00	0.00	3.00	26.80
Capulin	70.50	0.00	14.40	15.10	0.00	2.00	0.00

Table 2. Morphometric data from 1:24,000 scale maps for selected basins in the Jemez Mountains. See appendix 1 for definitions of parameters. N.A. indicates a parameter that is not applicable.

flows past the Pajarito fault only during the summer monsoons and spring snow-melt. Cochiti Canyon, with an area of 77.8 km^3, usually flows well past this same location during all but a few weeks during some years. Drainage nets constructed at 1:24,000 scale included not only the "blueline" streams (allegedly perennial, although Bland Canyon and many others in the Jemez would then have no streams highlighted in blue), but those that could be inferred from contour lines. By attempting to hold drainage area and rock-type nearly constant, it was hoped that a tectonic "signature" might reveal itself.

Drainage nets were made by overlaying topographic maps with tracing paper on a light table and tracing all obvious streams. 'Obvious', in this case, is relative to the person tracing the streams and further illustrates that morphometric analyses from different studies should be compared in only a general way. Stream lengths, basin areas, basin perimeters, basin lengths, and the area of various sections of each basin were then measured using a Lasico series 1280 digitizer with a computer interface. This procedure allowed data to be entered directly into a Microsoft Excel spreadsheet.

Cochiti Canyon is of a unique size among basins draining the eastern Jemez Mountains and flanked on both sides by upper Bandelier Tuff (table 1, figure 3). Most other basins in this region are at least 10 square kilometers smaller. Cochiti Canyon also contains an unusually wide range of rock types (table 2, figure 3). Since basin size and rock-type are the most important factors controlling the evolution of a basin, comparison of Cochiti Canyon to dissimilar basins would be misleading. For this reason, Bland Canyon was chosen as representative of the study area (figure 4). Bland Canyon is not only of comparable size to several other basins in the Jemez Mountains, but is more lithologically homogeneous than Cochiti Canyon. Results of some of the analyses done for Bland Canyon and similarly sized basins are presented in table 2 and figure 6. Drainage density does not seem to vary in a systematic way with basin area (figure 6 a,b), either when the total basin area or just the area of the basin above the main splay of the Pajarito Fault Zone is used. Basin shape (see appendix 2 for description of parameters) increases with increasing basin area (figure 6 c,d), but after consideration of the processes involved in basin elongation and widening, I believe this is merely a function of the greater number of tributaries in a larger basin. At this scale, the law of stream orders (Horton, 1947) holds (figure 6 c,d), but this too is a function of natural changes during the evolution of a basin.

I realized, after considering these and other combinations of my data, that I was looking in the wrong place for a tectonic signature in the landcape. I was attempting to find out if the Pajarito Fault System had been affecting the morphometry of these basins during specific time periods. Parameters such as drainage density and overall hypsometry are related to the entire geologic history of the basin, and are most heavily influenced by discharge and rock type. The Pajarito fault is not an extremely active fault (compared to, for example, faults in California with rates of movement on scales of millimeters per year) and therefore should not be expected to produce a strong effect in the entire basin. In this sense, it is not possible to determine the time periods during which a fault has been active with this tool. However, it might be possible to determine if a fault has affected the area immediately adjoining the fault zone during the last few (ten?) thousand years. With this in mind, I began an analysis of the sinuosity of streams, long profiles of streams, and

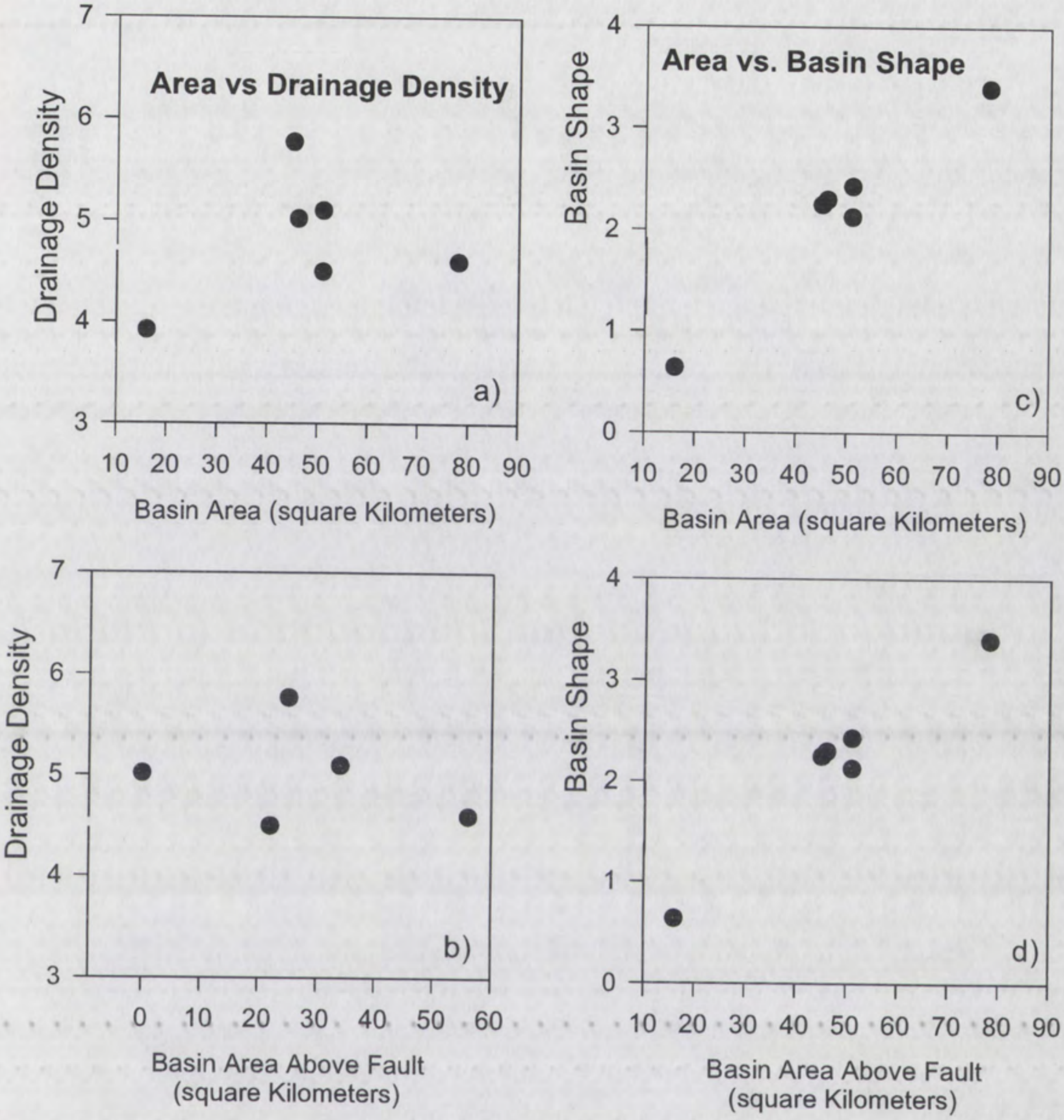


Figure 6. Plots of selected morphometric parameters calculated from 1:24,000 scale maps. Basin area above fault (b, d) is measured upstream of the main splay of the Pajarito Fault Zone (see figure 3).

hypsometry of portions of basins within one kilometer on either side of the main splay of the Pajarito Fault Zone.

Hypsometry of Basins Near the Main Splay of the Fault Zone

In order to asses the effect (if any) of the Pajarito Fault on valley hillslopes during the Holocene, hypsometric curves were constructed for portions of streams within one kilometer up and downstream from the main splay of the fault system. Hypsometric curves give a graphic display of the distribution of elevation within a basin. The hypsometric curve is constructed by:

1) Breaking the basin (or portion of a basin) of interest into sections based on contour lines. For convenience sake, I broke each two kilometer long section of stream into ten equal portions.

2) Measuring the area of each interval.

3) Plotting the area above some contour divided by the total area in the basin, against the elevation range of each slice divided by the total elevation range of the basin (in this case the local relief). This procedure is simply designed to normalize parameters (e.g. local relief and basin area) so that basins of different size and relief can be more easily compared. See Ritter, 1978, p170 for a full discussion.

The hypothesis was that, if splays of the Pajarito Fault Zone had been episodically active in the recent past and this activity had been preserved in hillslope morphometry, then offset would be seen as a distinct elevation "bench" in the hypsometric curve that would not be obvious on a topographic map. Results of this analysis can be seen in figure 7. Several canyons do seem to have an anomalous bench in their plots (Eagle, Sanchez, Capulin Tributary, Turkey Spring, Frijoles, and One south of Water), but this bench is not at a consistent elevation above the fault (as would be expected if they were all effected by the same offset events). In addition, variations in the erodability of bedrock (either within the Bandelier Tuff or because of formation contacts located near the fault zone) seem to control the position of these anomalies in the canyons I have had the opportunity to examine closely in the field (Eagle and Sanchez canyons).

These plots do not immediately reveal a tectonic signature, but some interesting observations can be made. First, let us further examine what the curves might be telling us about the history of the hillslopes near the fault zone. If much of a basin's area is at high elevation (the hypsometric curve is convex), it can be reasonably assumed that the hillslopes in that basin have not been able to keep up with stream incision. In other words, baselevel for that stream has dropped at a rate greater than the rate of hillslope adjustment to baselevel lowering. In the case of stream segments near fault zones, this baselevel fall can then reasonably be attributed to fault offsets. Conversely, a basin that has relatively little of its area at high elevation (a concave hypsometric curve) can be seen as more 'mature', or to have had a more stable baselevel for some substantial amount of time. The stream within such a basin can be assumed to be graded (see definition of graded in the discussion of terraces as indicators of tectonics below), and little recent faulting is implied. Finally, an equal distribution of elevation (a straight hypsometric curve) indicates that hillslope retreat has kept up with base level fall during the recent geologic past, and that either faulting has been moderate or the hillslopes have the ability to rapidly adjust to baselevel fall. For the basin segments examined here, it is reasonable to assume that the ability to respond to baselevel fall is constant between basins for similar rock types. It should be

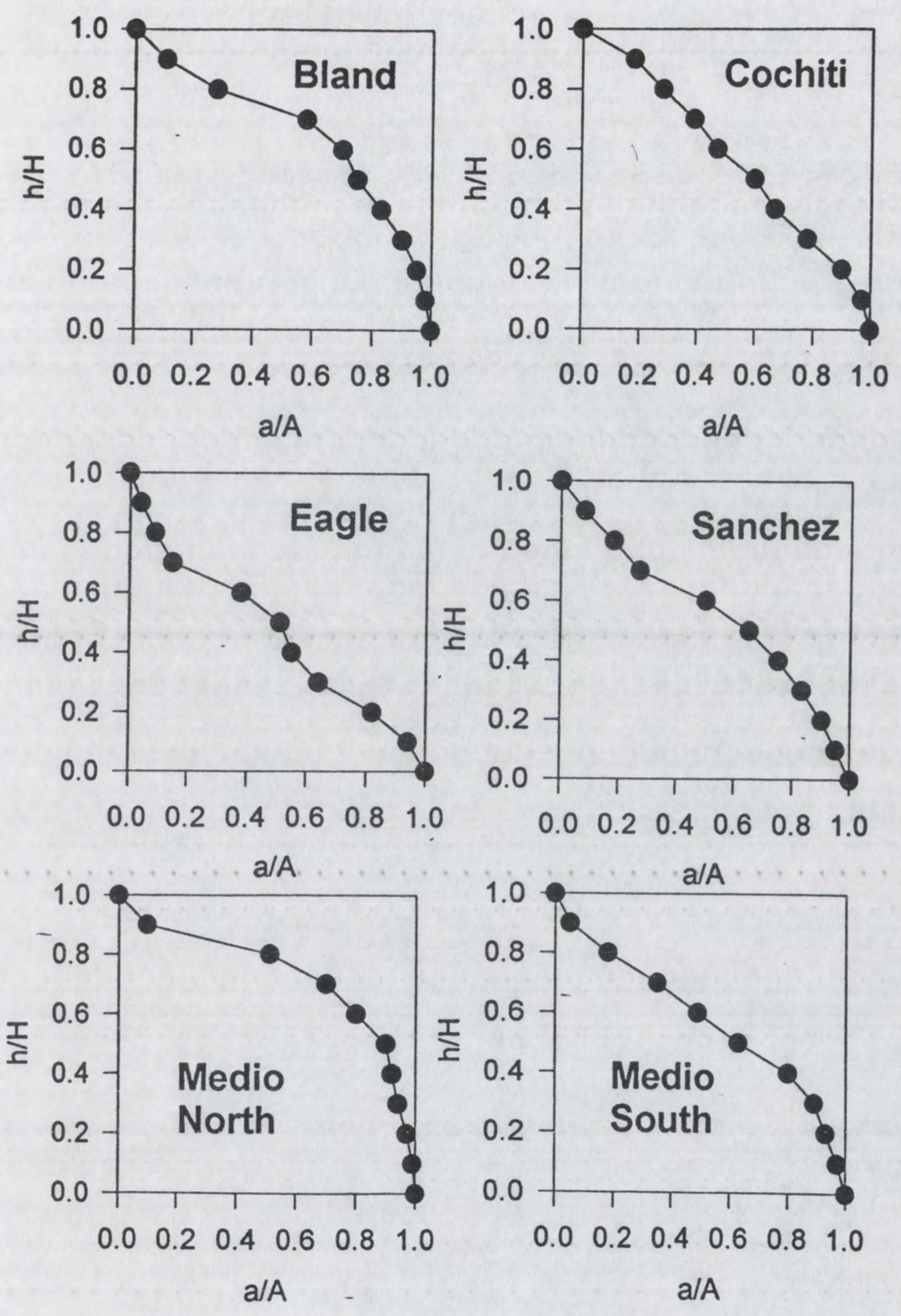


Figure 7. Hypsometric curves for portions of basins approximately one kilometer up-and-downstream from the main splay of the Pajarito Fault Zone. See text for an explaination of the plots and figure 4 for location of basins.

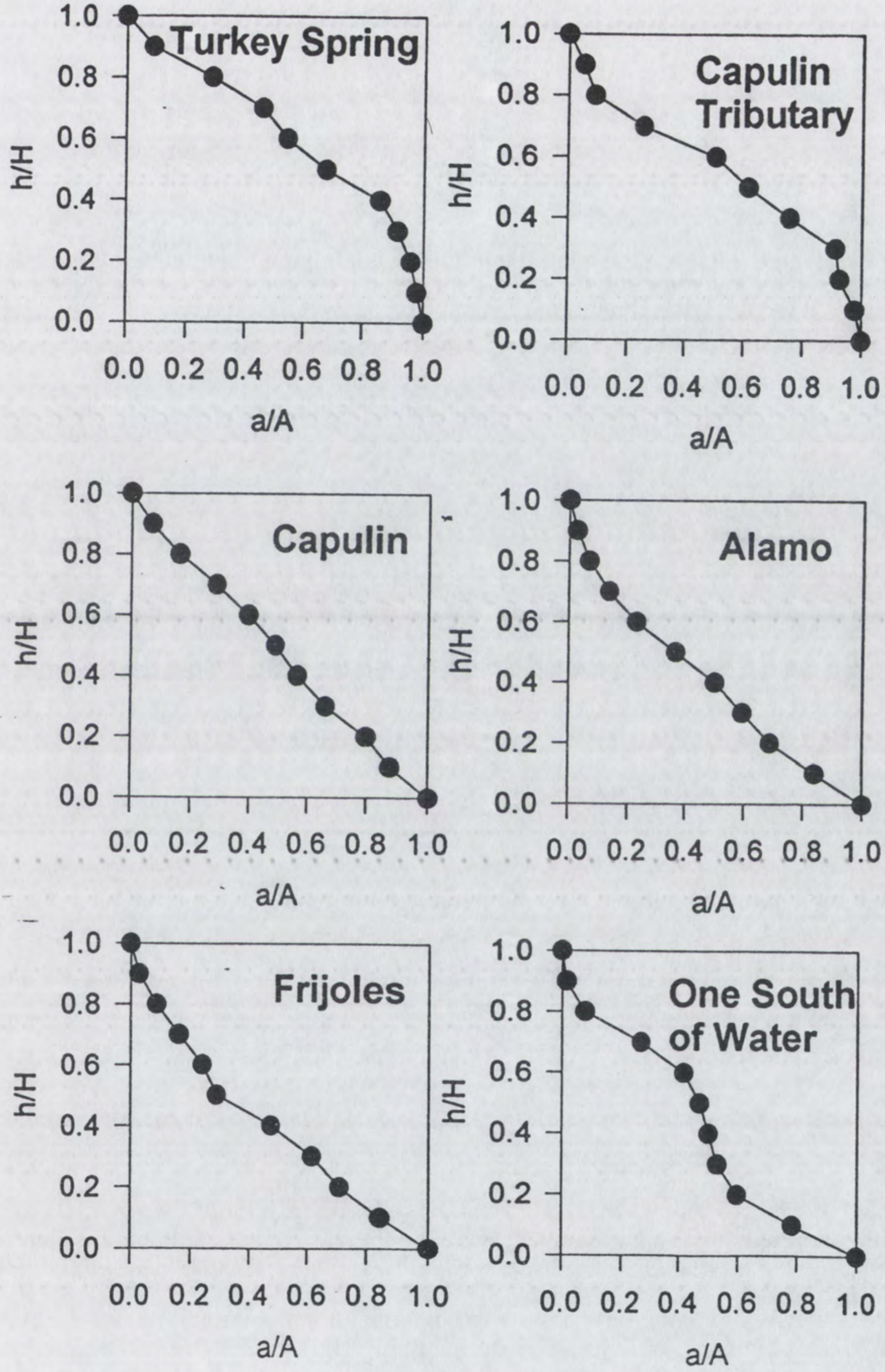


Figure 7 continued.

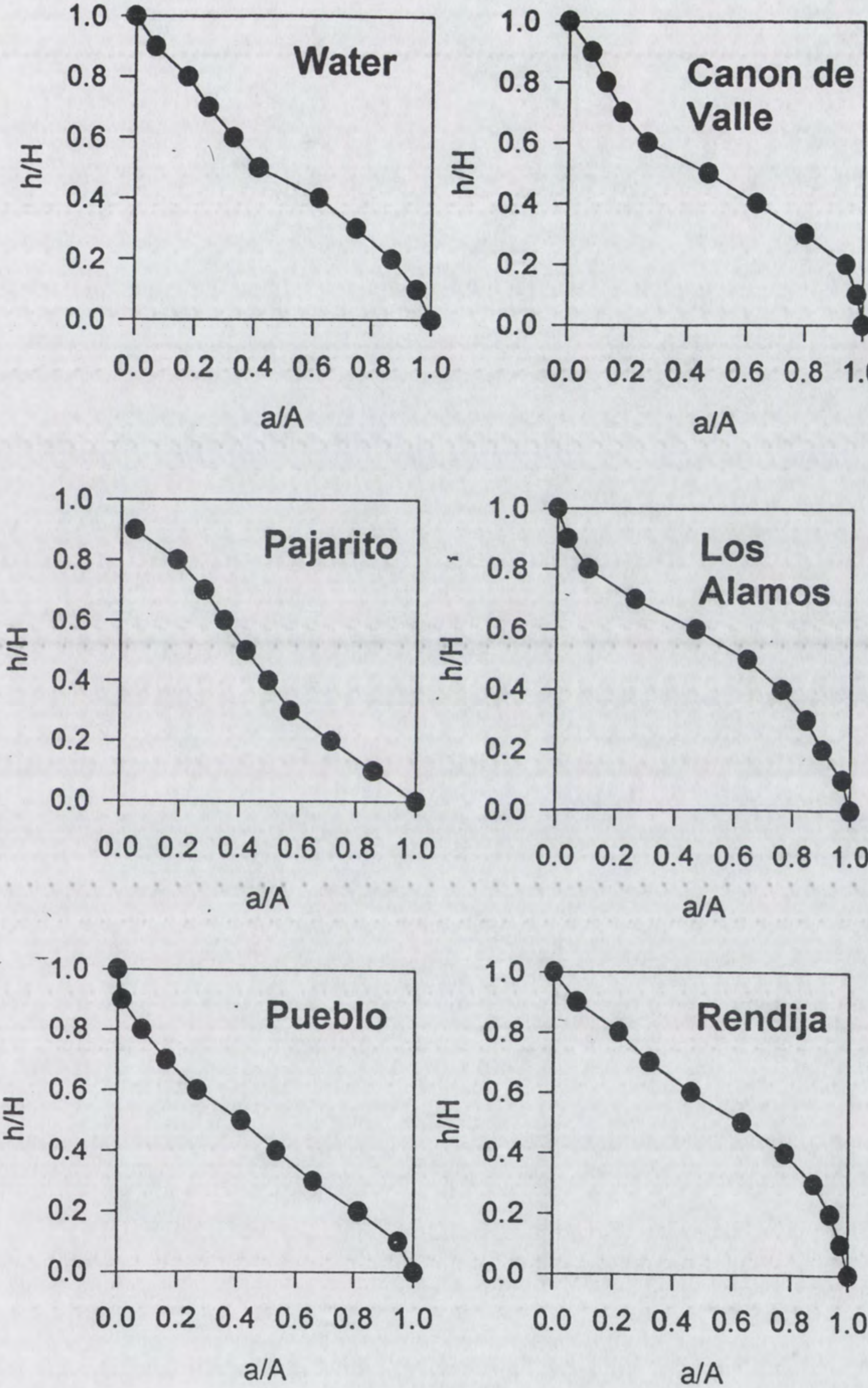


Figure 7 continued.

remembered that hypsometric curves tell you little about the actual shape of a canyon. For example, a perfectly 'V'-shaped canyon will have the same hypsometric curve whether the canyon walls lie at a 10 degree or an 80 degree angle (since elevation is *distributed* in a similar manner).

Examples of each of the three 'end-member' types of hypsometric curves discussed above can be seen in figure 7. Capulin Canyon has a fairly straight hypsometric curve. The hillslopes within this canyon are apparently adjusted to the amount of baselevel fall they have experienced in their recent history. If the inference between baselevel fall and hillslope response proposed above is correct, then this canyon appears to have been experiencing protracted base level fall and incision, or the hillslopes within this basin have only recently compensated for a previous period of (rapid?) baselevel fall.

Pajarito, Alamo, and Frijoles canyons all exhibit concave hypsometric curves, implying that baselevel has been stable for some extended period of time while hillslopes have slowly retreated. These canyons are all located in the central part of the Pajarito fault Zone (figure 1,4), and appear to be structurally separated from the study area by a complex zone near Capulin Canyon where several cross-faults have been mapped and where the La Bajada Fault intersects the Pajarito Fault Zone (figure 1). It is possible to suggest from these data that this segment of the fault zone has been only moderately active in the late Pleistocene (the exact time period is relative to the time it takes hillslopes to adjust to base level fall).

Bland, Cochiti, Sanchez, Turkey Spring, two portions of Medio Canyon, Los Alamos, and Rendija canyons all have predominantly convex hypsometric curves (figure 7). With the exception of Los Alamos and Rendija Canyons, these canyons are all located in the southern part of the Pajarito Fault Zone, near or within the study area (figure 4, 7). Again, if the inferences made about hillslope response to baselevel fall are correct, then these canyons have experienced rapid baselevel fall in the relatively recent past, and the southern segment of the Pajarito Fault Zone would be seen as the most recently active. Two observations complicate this simple picture. Basin size may also play an important role in determining the response of individual basins, with relatively large basins responding to baselevel fall more quickly than smaller basins. Second, rock-type is not constant for all basins examined.

The most important clue as to the timing of activity on the Pajarito Fault that can be gleaned from the analysis of these hypsometric curves is the fact that canyons with predominantly convex hypsometric curves are all within a few kilometers of Cochiti Canyon (figure 4). It can be further noted that, within this subset, those with smaller drainage basins (Medio North and Medio South) and those in less resistant bedrock (Medio North, Medio South, and Bland) have the most dramatically concave hypsometric curves (figure 1, 7). Smaller basins are not as able to maintain a graded state during baselevel fall, or to compensate for baselevel fall as rapidly as are larger basins. Therefore, these basins may preserve a longer record of baselevel change, or may preserve such changes more faithfully.

The relative concavity of all these curves can be interpreted as indicating that this segment of the fault zone has been more active than other segments in the relatively recent past. One additional observation adds to this discussion. Post-Bandelier offset is greatest at Cochiti and Bland Canyons and decreases to the north until the intersection with the La Bajada Fault (Golembek, 1993) while hypsometric curves become more drastically convex in this same direction. This suggests the possibility that relatively recent faulting began near Cochiti Canyon and has propagated to the north--with the most recent faulting located in the northern part of this fault segment.

Hypsometric curves offer one additional means by which fluvial systems may be examined, but conclusions about basin history and the effect of faults on hillslopes must ultimately rely on field observations of individual basins and knowledge of process rates. The rates of interest in this case are: rate of hillslope erosion (and lag time between stream incision and hillslope response), bedrock erosion rate of streams, and rate of offset along the faults of interest. These rates are not yet available for the Jemez Mountains, so further investigation of these relationships must be left to future workers. Lastly, I believe that my observation that smaller basins have the most dramatically convex hypsometric curves within the study area should be taken as an important clue to future workers attempting to resolve tectonic activity by means of hypsometry. Merrits and Vincent (1989) have previously suggested that first-order streams (on 'blue-line' drainage nets) are the best recorders of varying uplift rate in tectonically active, humid Humboldt County, California. My analysis indicates that smaller-order basins may also be preserving the record of less-active tectonics in the arid southwest.

Long Profiles of Streams

While hypsometric curves can give an indication of how hillslopes have reacted to changes in baselevel, long profiles (simply a plot of downstream distance versus stream elevation) can reveal how stream-channels themselves are responding to these same changes. It can be assumed that stream channel elevation will respond more quickly to variations in base level than will the adjoining hillslopes, and will therefore reveal tectonic activity on a shorter (but still unknown) time scale. The hypothesis behind the use of long profiles as indicators of the recency of tectonic activity hinges on the fact that a faulting event will cause a local increase in stream gradient. This increase in stream gradient will in turn cause a local increase of stream power that will act to smooth the locally steep portion of a stream's gradient. Hence, if a locally steep portion of a stream (a nickpoint) exists near a fault, it may be inferred that the fault has been active 'recently'. How long 'recent' is in years is a function of individual stream power (the power of the stream to remove a nickpoint) and the resistance of a particular bedrock to erosion, since these factors will control the rapidity of profile smoothing after a faulting event. Tectonic activity is not the only variable that can produce a nickpoint. Lithologic changes appear to be the most common cause of nickpoint formation in the study area, and increases in stream power due to increases of discharge at tributary confluences can also cause rapid changes in trunk-stream gradient. In order to isolate tectonically-induced nickpoints in streams that cross the Pajarito Fault Zone, these other factors must first be ruled out. I attempted to accomplish this by first constructing the long profiles and then determining if any visible nickpoints coincided with mapped geologic contacts or stream confluences (figure 8). In the following discussion I address individual features in five canyons within or near the study area. Features noted on the profiles of these canyons (e.g. nickpoints, fault traces, lithologic changes,...) are illustrative of those seen on the profiles of canyons further north. I have limited full descriptions to these canyons because I am most familiar with their geology from first-hand observations and mapping.

Peralta Canyon

Beginning downstream, two splays of the Pajarito Fault Zone cross this canyon at between 5500 and 5600 feet, but no dramatic nickpoint can be seen (figure 8a). A small nickpoint at 5800 feet coincides with a rhyolitic intrusion that forms a dramatic falls in Colle Canyon (a tributary of Peralta that trends north from this point; figure 4). Two more splays of the Pajarito system cross the canyon at approximately 6050 feet elevation, and a substantial increase in gradient is evident above this point. However, above this point the stream flows on 'low-permeability bedrock" while below this point the stream flows in "permeable fill" (Smith, written communication 1996) Several small nickpoints upstream of this point seem to coincide with numerous, mapped geologic contacts between rhyolitic flows and flow breccias (Goff *et al.*, 1990; Smithe *et al,* 1970).

Bland Canyon

This canyon is cut entirely within thin alluvium over Peralta Tuff above the Pajarito Fault and thin alluvium over lower Bandelier Tuff below the fault. Where they are exposed, these units appear to be similar in their resistance to stream erosion. The main splay of the fault zone is exposed in the southern wall of Bland Canyon (Gardner and House, 1987, fig 9) and appears to offset post-Bandelier alluvium) by up to 6 meters here. No age determination was made for the post-Bandelier alluvium by Gardner and House (1987), but I attempt to estimate an age for this deposit in following section of this report. Although the profile of Bland canyon is perhaps closer to an "ideal" graded profile than any other canyon examined, the most prominent discontinuities in the profile do cluster near this fault (figure 8b). For this reason, this section of the profile has been plotted on a smaller scale (figure 10). Here it can be seen that the main nickpoint is indeed located just upstream of the main splay of the fault system. Although this could conceivably be the result of the change in lithology that takes place here, the previously mentioned evidence for recent fault movement here lends credence to the hypothesis that this nickpoint is fault generated and that long profiles may be an effective means of locating active fault splays. No nickpoint is seen where Bland Creek crosses another splay of the fault zone (the Dixon fault) downstream of the main splay. This splay clearly offsets two terraces as it crosses Cochiti Canyon, but no nickpoint can be seen in the profile of Cochiti Canyon. One must conclude that, either the change in rock type noted for the main splay is more important than hypothesized, or that the most recent activity on the Dixon fault has been erased because it occurred at an earlier time. One other fault (with down-to-the-west displacement) is mapped crossing Bland Canyon before the confluence with Cochiti Canyon. This fault is shown juxtaposing 'River Gravels' (well-rounded cobbles of Precambrian quartzites) against 'Older Alluvium' (predominantly volcanic sand and gravel "...partly interbedded with..." Rio Grande-type gravels (Smith *et al.,*1970) and lower Bandelier Tuff. This fault does not displace the upper Bandelier Tuff to the north and does not affect the long profile of Bland Canyon.

Cochiti Canyon

The upper reaches of the Rio Chiquito (the stream in Cochiti Canyon) experience many changes in rock type (Goff *et al*, 1990) and these are reflected in a markedly irregular stream profile on the upthrown block of the Pajarito fault (figure 8c). In particular, nickpoints can be seen where Quaternary landslides have been mapped (Goff *et al*, 1990). Immediately upstream of the main splay of the fault, a small nickpoint (that forms a falls in

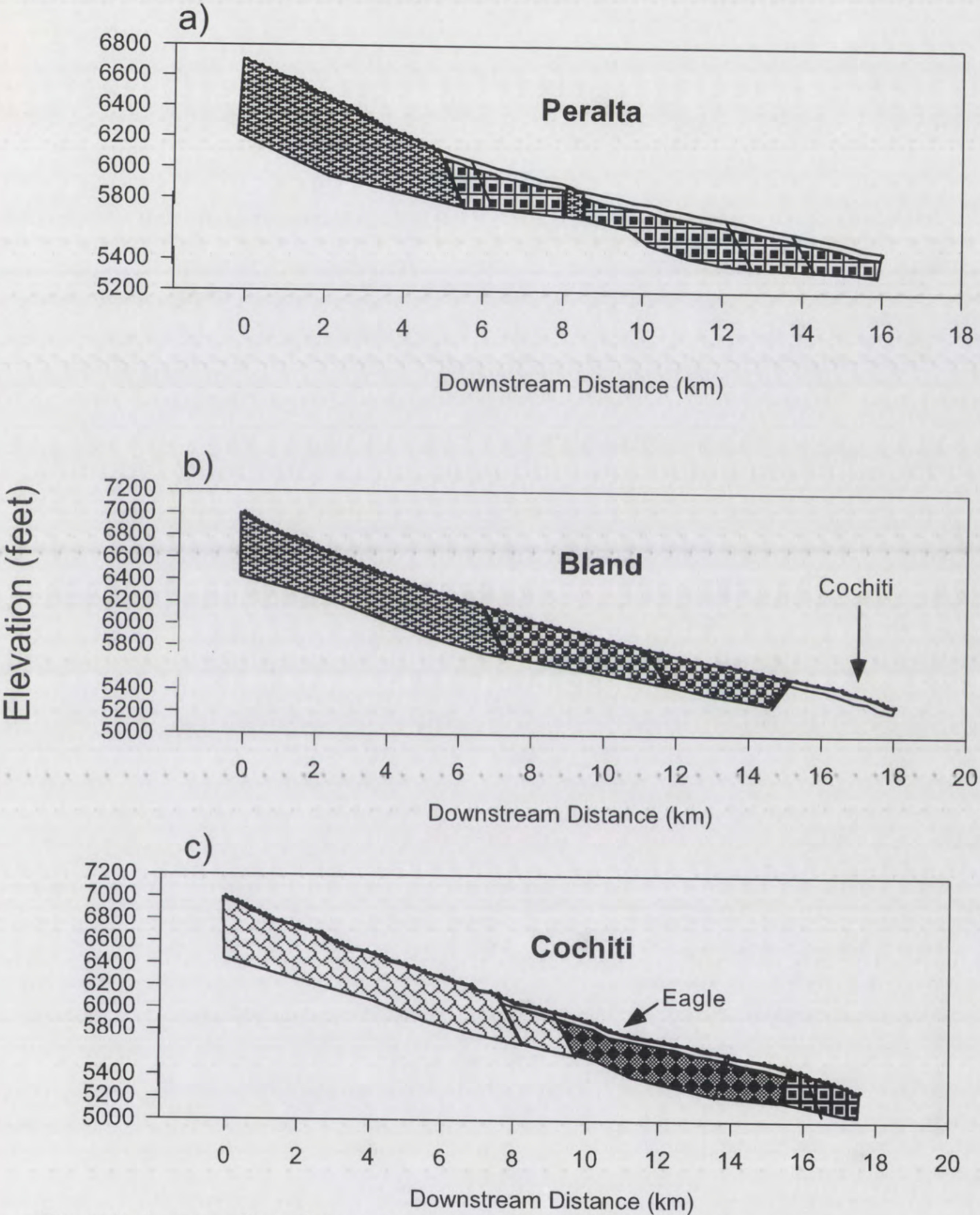
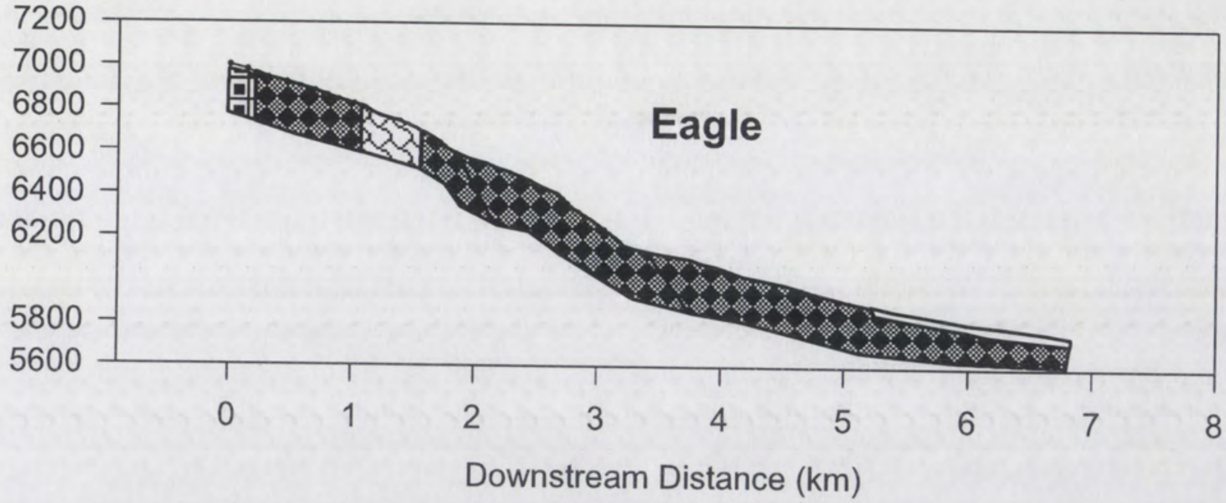


Figure 8. Long profiles of streams constructed from 1:24,000 scale maps. Lithology shown by Smith et al. (1970) is superimposed. Confluences are indicated by arrows. See figure 8b for Key. The profiles are arranged from north to south.

Figure 8 continued.

d)



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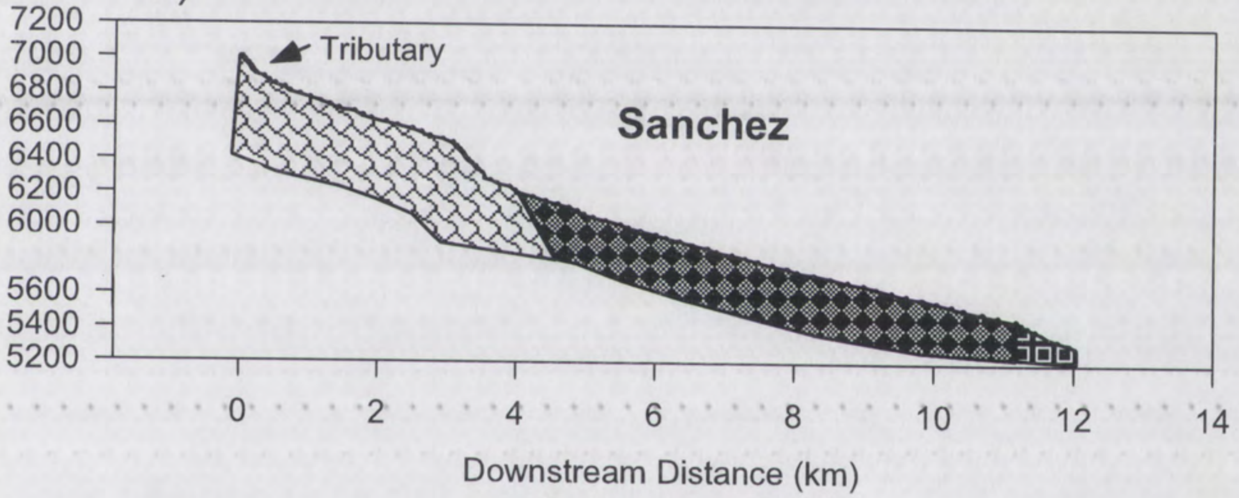

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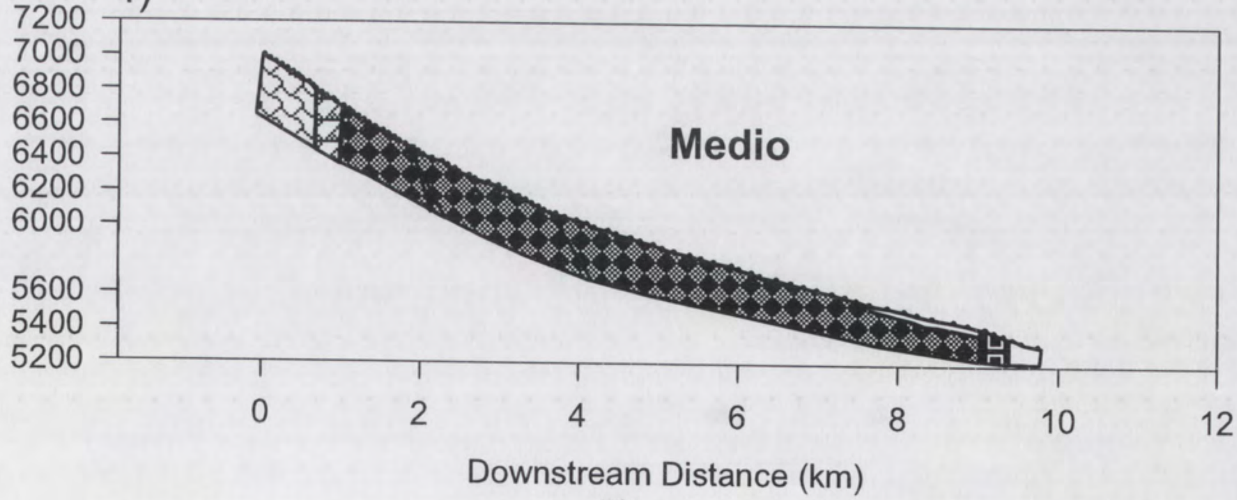
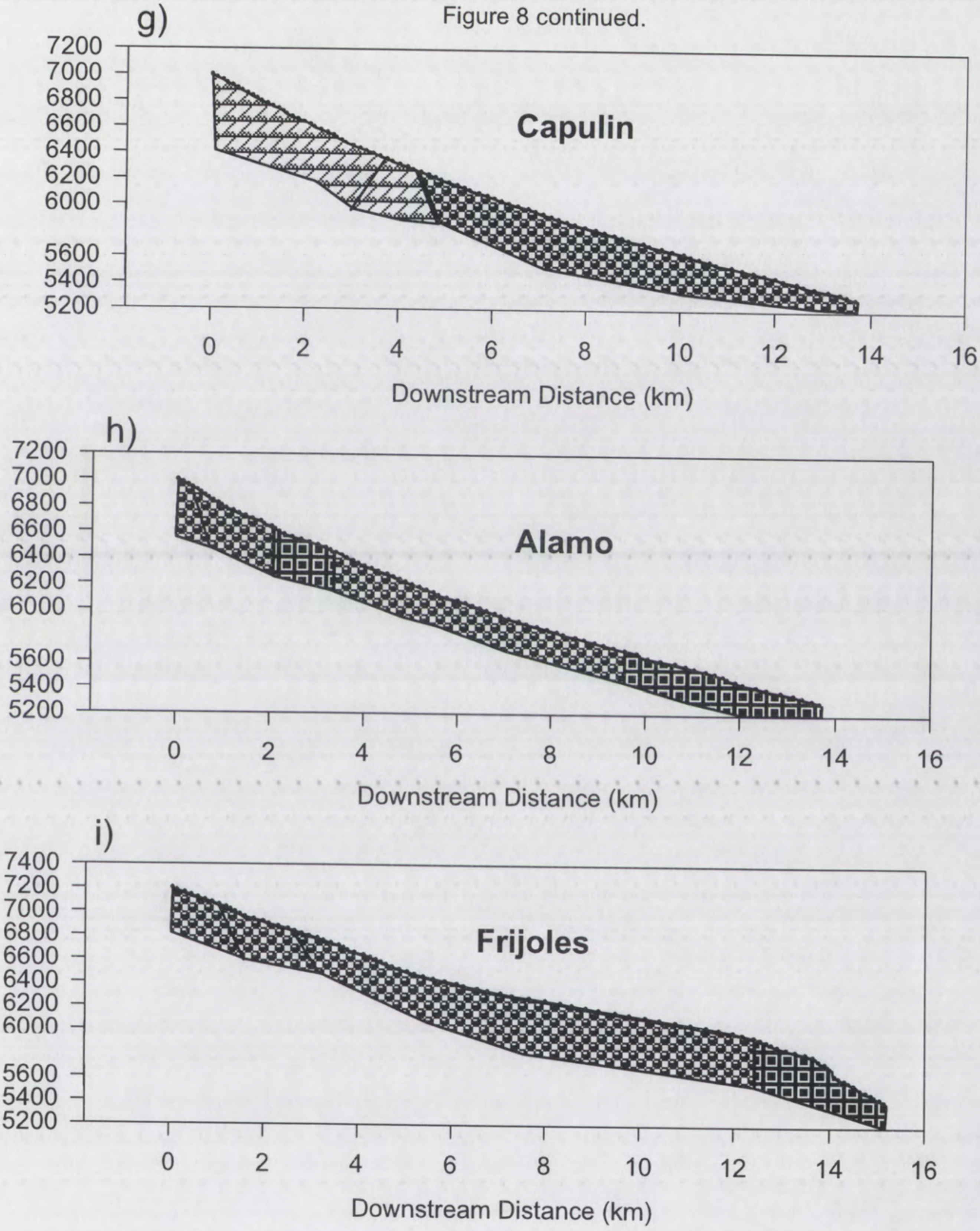


Figure 8 continued.



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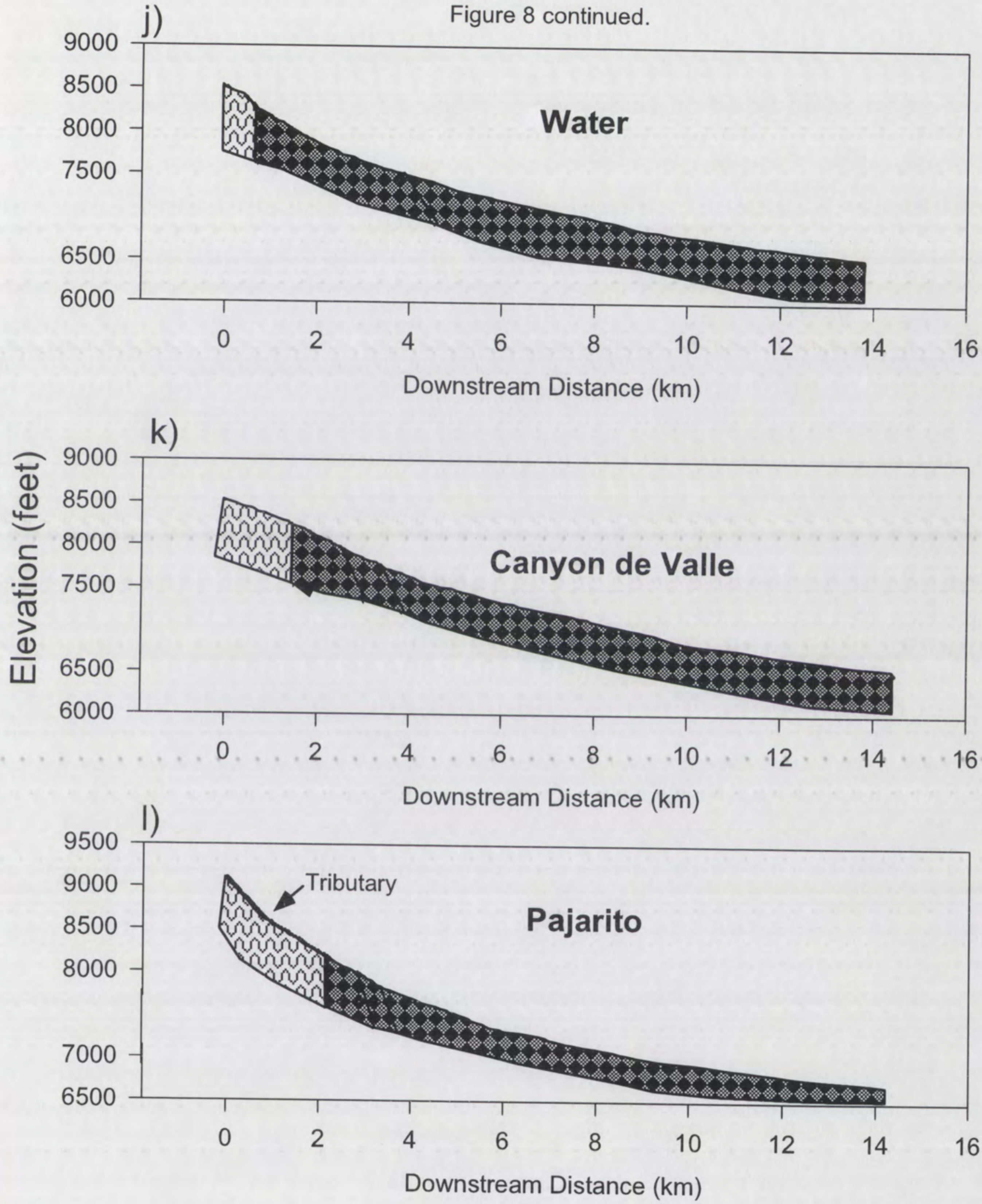
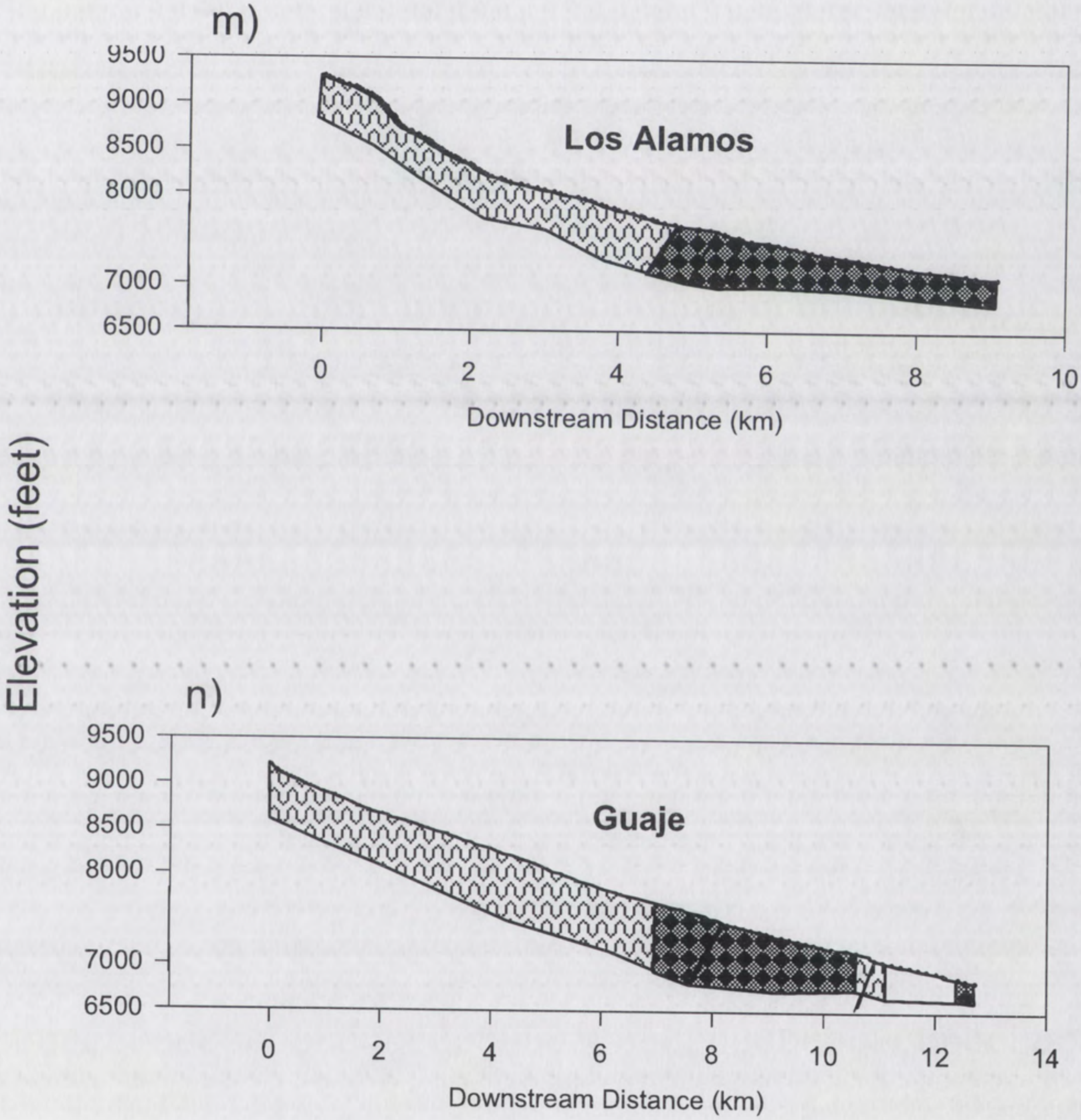


Figure 8 continued



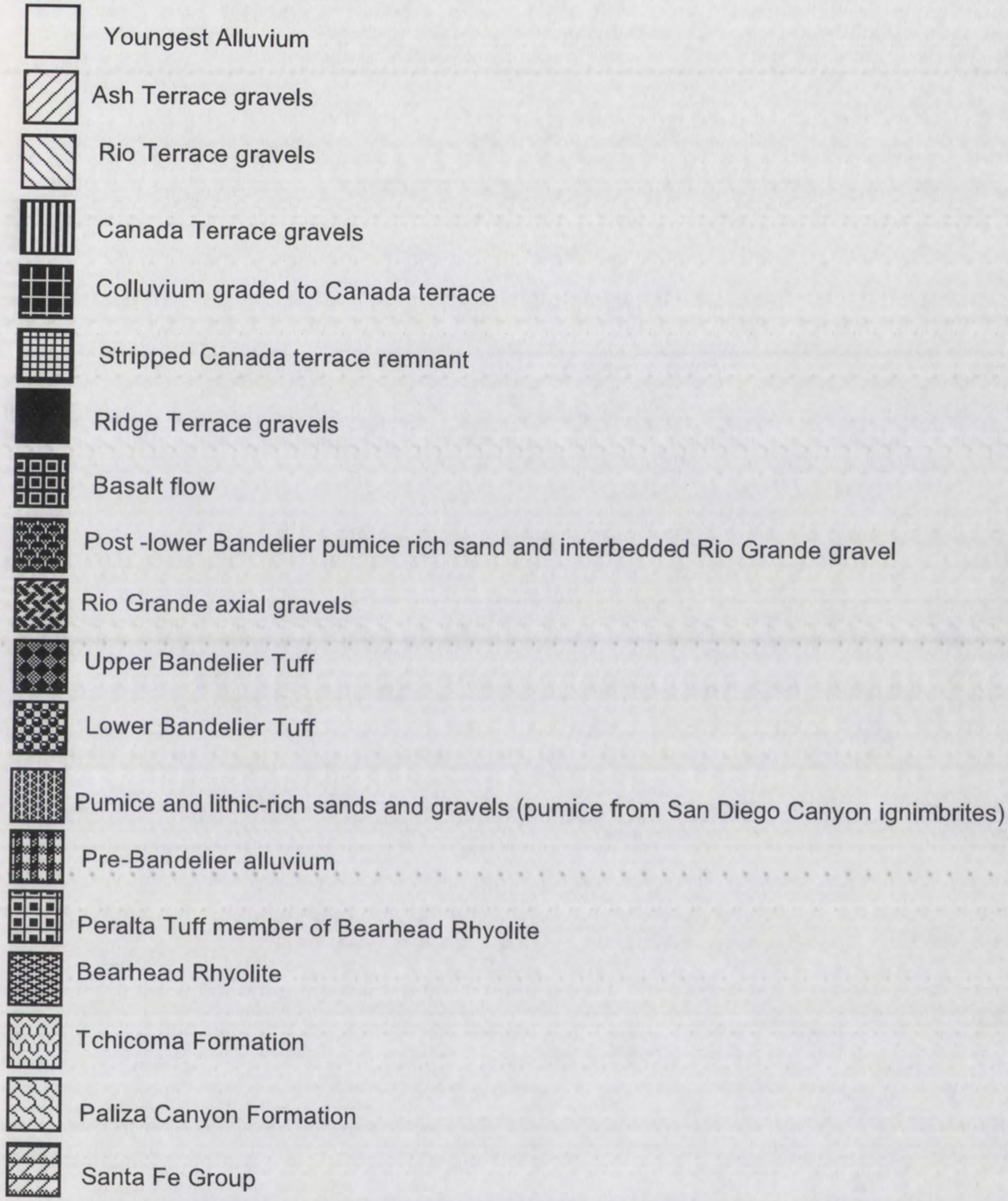


Figure 8b. Key to lithology for figures 8a, 10, 11, 12, and 13.

Figure 9. Exposure of the main splay of the Pajarito Fault Zone in a cut-bank of Bland Creek. The fault juxtaposes the Peralta Tuff member of the Bearhead Rhyolite (~6.8 Ma) on the right, against Quaternary stream gravels on the left. The degree of soil development in the stream gravels (whict extend across the fault onto a surface beveled in the Tuff on the footwall) suggests that they are approximately 85 +/-20 ka..

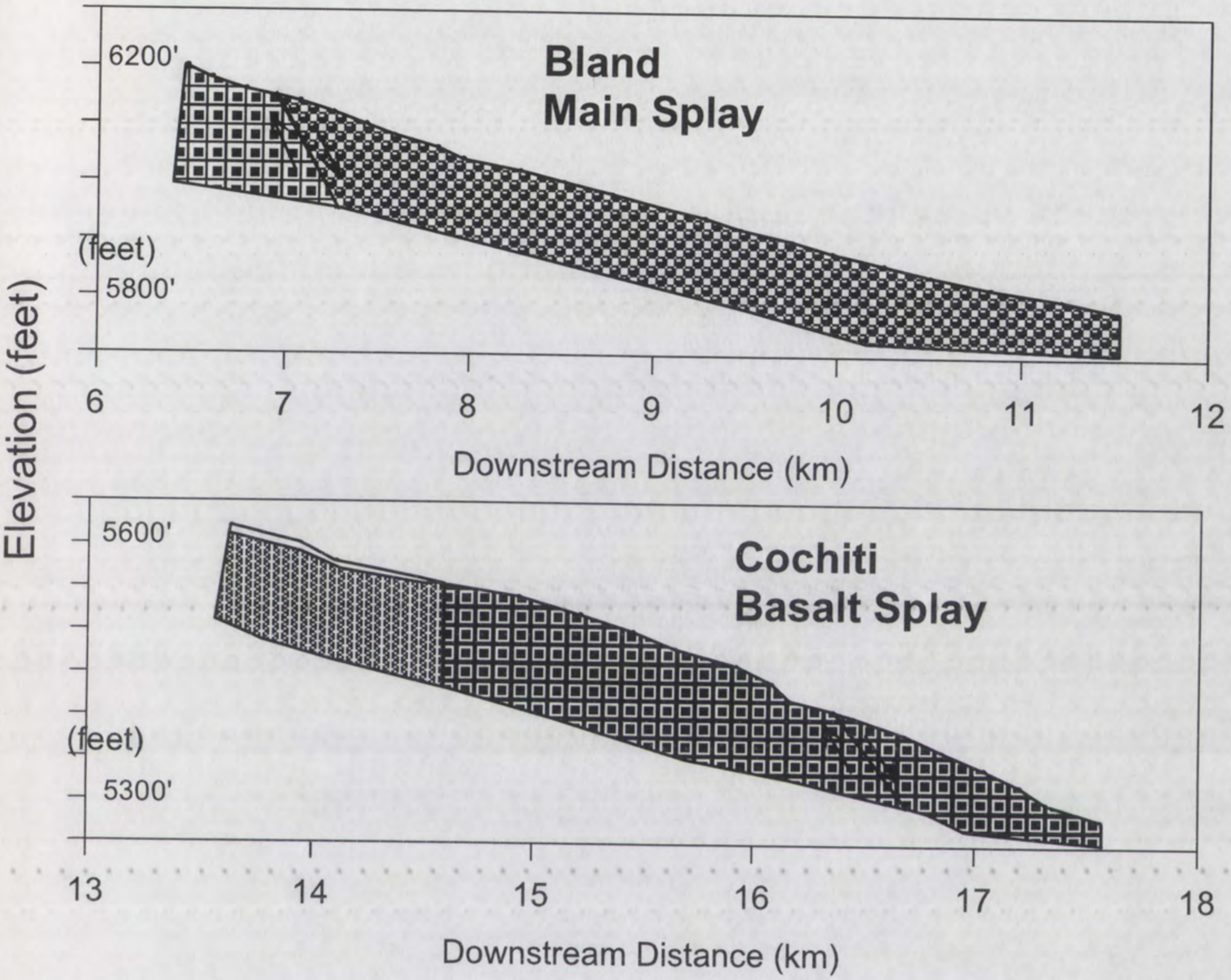


Figure 10. Enlargements of portions of long profiles of selected streams near fault splays. See figure 8b for key.

the canyon) can be seen cutting through andesite of the Paliza Canyon Formation. In this case it is not clear whether the nickpoint is due to the rock-type change across the fault (andesite to alluvium) or due to activity on the fault. Here again we see the importance of rock type on the profile of the stream. If it was known how long it took the stream to smooth out a tectonically induced nickpoint in Paliza Canyon andesite, then a time-since-last motion on this splay of the fault zone could be calculated. However, the presence of more dramatic nickpoints farther upstream--which do not appear to be related to faults--suggests that this stream simply does not have the requisite power to cut through resistant bedrock. As in Bland Canyon, the Dixon fault and one antithetic fault (see above) mapped by Smith *et al.* (1970) do not produce a nickpoint in the Rio Chiquito at this scale. It should be noted that the Rio Chiquito has been modified by the apple orchards of the Rancho Cañada and (downstream of the confluence with Eagle Canyon) by additions of pumice from mining operations. The most downstream fault mapped crossing Cochiti Canyon does so below a nickpoint created by the streams attempt to cut through a Plio-Pleistocene Basalt flow of the Cerros del Rio field (figure 3,10). This basalt flow is capped by Rio Grande axial stream gravels that are themselves offset across this splay of the fault (the Basalt splay). This surface is apparently correlative with the Cañada terrace (see section on terraces) and has been tentatively dated as 350-240 ka based on calibrated varnish-cation ratios (Dethier and Harrington, 1987). If this age is roughly accurate, then fault motion here since approximately 0.3 Ma can be inferred. A nickpoint is seen in the stream profile above this fault that is more prominent than the nickpoint created where the stream first reaches the basalt flow (figure 10)

This fact indicates motion on this fault since the stream began to incise through the basalt (post-Cañada terrace formation).

Eagle Canyon

Eagle Canyon's long profile has two rather dramatic nickpoints (figure 8d). The one farther upstream coincides with the main splay of the Pajarito Fault Zone. Although Goff *et al.* (1990) map a Canovas Canyon intrusive (symbol on their map indicates Canovas Canyon, but color indicates Paliza Canyon andesite) just above this point, field observations show that the nickpoint is actually cut into Upper Bandelier Tuff. The stream also takes a prominent southerly dogleg at this point (figure 2,4). Although the bend in the stream suggests right-lateral offset, no other indication of this type of motion is seen anywhere else in the area and it is known that streams will sometimes follow fault zones regardless of the sense of displacement (Pazzaglia, personal communication 1995). An even more prominent nickpoint downstream in this canyon is found entirely within Bandelier Tuff and does not seem to be related to intra-Bandelier contacts. This nickpoint does, however, coincide with the projection of the Dixon fault (figure 2). It is hypothesized here that this nickpoint is preserved due to the smaller drainage basin size of Eagle Canyon as compared to Cochiti or Bland Canyons (table 1). This nickpoint (if strictly fault-generated) shows that this fault is affecting the *modern* stream. If this is the case, then the place to look for evidence of fault motion is in the smaller streams crossing fault scarps. This suggestion is in agreement with the conclusions drawn from hypsometric analysis.

Sanchez Canyon

The most prominent nickpoint in Sanchez Canyon forms a vertical waterfall in the stream. The nickpoint is cutting through a Quaternary landslide of andesitic breccias derived from Cochiti Formation (Goff *et al.*, 1990) and also coincides with the main splay of the fault zone (figure 8e). Since Cochiti and Sanchez canyons are so close to each other, and since Cochiti Formation is less resistant than Paliza Canyon andesite to erosion, it would appear that the variation in nickpoint height between the two canyons must be related to a relatively recent increase in landslide activity. It is tempting to speculate that the landslide was earthquake-generated and therefore emphasizes or preserves a nickpoint that would be there anyway. Unfortunately, no independent data exist to confirm this hypothesis. One nickpoint upstream of the fault is coincident with a major tributary confluence and one near the mouth of Sanchez Canyon is coincident with the basalt contact mentioned in the above discussion of Cochiti Canyon (figure 3). Besides these two other small nickpoints, the profile of Sanchez Canyon is remarkably smooth considering that the number of lithologic changes shown by Goff *et al.* (1990) is comparable to that shown for Cochiti Canyon--which is quite irregular in its upper reaches. Either the specific rock-type changes in this canyon are different than those in Cochiti Canyon or Sanchez Canyon is more capable of 'smoothing' nickpoints by erosion of bedrock and/or deposition of alluvium.

Summary

Most of the nickpoints seen in long-profiles of streams along the eastern flank of the Jemez Mountains can be attributed to rock-type changes. Fault-related nickpoints are observed in the profiles of three streams: Bland Canyon (Main splay), Cochiti Canyon (Basalt splay), and Eagle Canyon (Main splay and Dixon fault). All of the canyons examined in detail have nickpoints where the main splay of the Pajarito Fault Zone is crossed, but this is clearly due to rock-type changes in most cases. Although the rock-type changes are ultimately related to fault-offset, these nickpoints may be ‘relict’ features related to ancient offsets. From these profiles alone it is not possible to determine how ancient these offsets are, since no information on rates of nickpoint degradation are available for these streams/rock-types.

Other stream profiles have been created (figure 8f-n), but I am not as familiar with the geology of these canyons as I am with those discussed above. From the map of Smith *et al.* (1970) it appears that most of the nickpoints seen in these other stream profiles can be attributed either to lithologic changes or to tributary confluences. It is interesting to note that some of these streams have markedly concave-up profiles (e.g. Pajarito Canyon). Since this is the classic 'graded' profile of a stream, it seems likely that streams in this region *do* approach a graded condition *if* conditions (lithology, discharge, offset rate of faults, ...) permit.

Sinuosity

Schumm *et al.*, (1987) has shown that, before an alluvial stream that is in a quasi-equilibrium state will change its mode of operation (aggrading or incising) in response to some external factor change, it will first adjust its gradient via changes in channel shape and/or sinuosity (e.g. Schumm, *et al.*, 1987). If this is accepted, then it is apparent that sinuosity can be used to suggest tectonic activity on even shorter time scales than hypsometry or stream-profile analysis. Sinuosity is defined here as the ratio of a stream’s channel length to the length of the stream’s valley axis along the same reach (a perfectly straight reach will have a sinuosity of 1). In order to see if tectonic activity is affecting modern streams, sinuosity was calculated from air photos for reaches of streams just upstream and just downstream of the main splay of the Pajarito Fault Zone in the study area. Reaches were between one and two-and-a-half kilometers long and reaches of comparable length upstream and downstream were chosen for each stream. This analysis is hypothesized to be sensitive to the most recent tectonic activity (if any such activity exists), and should also have the shortest temporal range of any of the morphometric techniques used so far, since stream channels are modified on a year-to-year basis. Stream reaches downstream of the fault (and thus theoretically backtilted to lower gradient) are consistently *more* sinuous than reaches upstream of the fault (figure 13). This is opposite to the effect active tectonics will have on a stream reach according to Schumm, *et al.* (1987). They hypothesize that reaches with gradients that have been increased by tectonic activity (in this case the reaches on the upthrown block) should compensate for this increase in valley gradient by increasing their sinuosity and therefore maintaining a constant stream gradient. However, the experiments mentioned in their monograph deal exclusively with alluvial channels affected by uplift of whole stream reaches, not increases in gradient caused by offset along a discrete fault. If uplift of one reach is accomplished by offset along a discreet fault, then the uplifted reach can be expected to expend the bulk of its energy in vertical incision. This vertical incision will be accomplished primarily by the migration of a nickpoint upstream from the fault scarp. The fastest way for the stream to 'smooth' the nickpoint is for it to cut a straight channel through its bed, and such incision will have the effect of decreasing sinuosity. This situation is also opposite that discussed by Schumm *et al.* (1987) in that the sinuosity of one reach is *decreasing* relative to the adjacent reach rather than sinuosity *increasing* for one reach relative

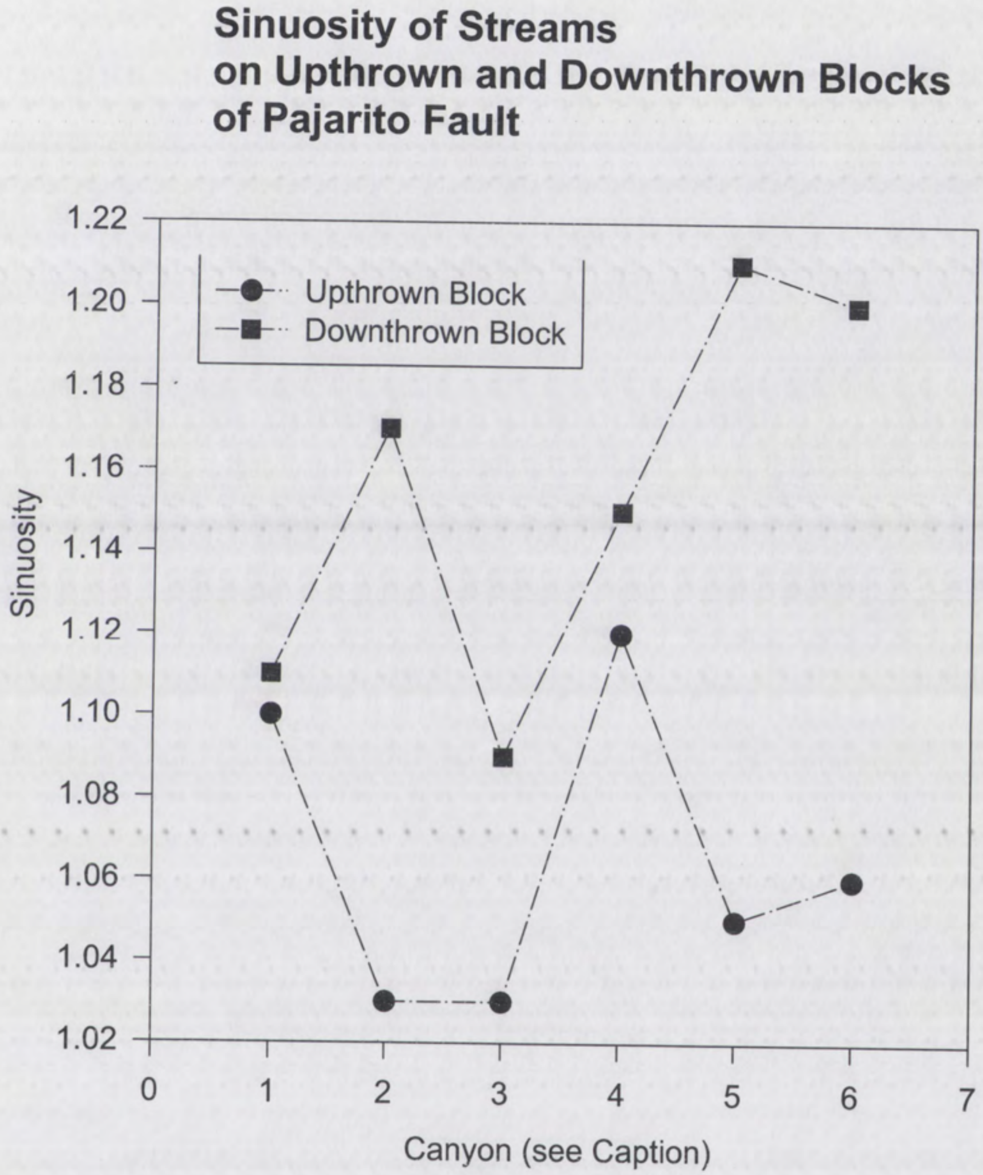


Figure 11. Plots of sinuosity of 1 km stream segments on the upthrown and downthrown block of the main splay of the Pajarito Fault Zone. 1=Peralta, 2=Bland, 3=Cochiti, 4=Bland, 5=Sanchez, and 6=Medio.

to another. It is also likely that the reach downstream of a fault will lie on a thicker alluvial fill than the upstream reach. In many cases, the reach upstream of a fault will be a bedrock reach while downstream reaches will be alluvial (e.g. at many range-front faults). Under these circumstances the type of channel also favors higher sinuosity on the downthrown block (since bedrock reaches are almost always less sinuous than alluvial reaches in the eastern Jemez mountains). This is exactly the case for Cochiti Canyon. It might also be expected that bedrock reaches incising into relatively more resistant formations will be less sinuous than adjacent bedrock reaches incising softer materials. This situation can be used to explain the sinuosity changes seen in Sanchez Canyon, (although here the upstream and downstream reaches of the stream adjacent to the fault are also separated by a 15 m high waterfall) and in the tributary of the main channel in Medio Canyon (figure 4). The remainder of the streams (in Peralta, Bland, and Eagle Canyons) flow either on a thin alluvial fill on both the upstream and downstream reaches (Peralta and Bland) or entirely within the Bandelier Tuff (Eagle). However, as can be seen in the graph (figure 14), sinuosity for these streams varies by less than ten percent on upthrown and downthrown blocks of the fault. It is unknown how large a variation in sinuosity should be expected if the Pajarito fault has been active in this area in the last few thousand years. Ultimately, it is unclear whether 'noise' created by lithologic changes can be separated from a tectonic influence on sinuosity, but tectonic activity is not precluded by the observed sinuosity patterns.

Conclusions

Like any 'state-factor' approach to analysis, morphometric analysis is hindered by the complicated nature of natural systems. No variable in stream channel or drainage basin evolution is simple, and their ultimate influence on landscape evolution can generally only be hypothesized. Further complications arise from the lack of knowledge of process rates and which processes (e.g. rainsplash, mass movement, sheetwash, ...) are active on different parts of a drainage basin (i.e. hillslope, stream channel, mesa top) during different periods in geologic time. With these limitations in mind, several general points can be made about stream systems on the eastern flank of the Jemez Mountains with respect to the presence of the Pajarito Fault Zone.

1) Characteristics resolvable on large and small scale (1:100,000 and 1:24,000) maps follow most of the "rules" of stream morphometry (Horton, 1947), but no tectonic signature can be completely isolated due to problems of scale and the influence of variable controlling factors..

2) Drainage basins are notably elongate (figure 4), probably due to their relative youth (all basins have evolved since eruption of the upper Bandelier Tuff 1.22 Ma), and sapping processes and jointing within the Bandelier Tuff, and active headward erosion precipitated by downcutting of the Rio Grande (Wells, *et al*, 1987).

3) Hypsometric curves are markedly more convex within the study area than in other areas along the fault zone, and smaller basins appear to record this convexity best. This can be taken to indicate that the Pajarito Fault Zone has been most active within the study area in the recent geologic past, although no age control can be placed on tectonism using this method.

4) Long profiles of some streams created from 7.5 minute topographic maps show pronounced nickpoints on reaches that cross various splays of the fault system. If it is assumed that all streams in this area were affected by the same

tectonic events, then smaller streams apparently preserve fault-generated nickpoints for longer periods of time.

5) Changes in hypsometry that are hypothesized to reflect relatively recent tectonism, and nickpoints in stream profiles related to faults are both preserved best in smaller basins. This observation should be used as a guide by future students of stream morphometry interested in resolving tectonic signatures.

6) Streams are consistently more sinuous downstream of fault splays within the study area. This observation is opposite that made for entirely alluvial reaches undergoing regional uplift (Schumm, *et al.*, 1987) where uplifted reaches have greater sinuosity. To my knowledge, this observation has not been discussed previously. It would be worthwhile to calculate sinuosity for reaches of alluvial streams that are offset by discrete, active faults (such as those in northern California), but such a survey is beyond the scope of this study.

Examination of Terraces

Introduction

Stream terraces are commonly used as indicators of tectonic movements (e.g. Merritts, *et al.* 1994; Pazzaglia and Gardner, 1994). In the simplest case, a deformed longitudinal terrace profile may reveal the presence of active geologic structures. Even this simple example, however, contains hidden assumptions and complexity. For example: How is a longitudinal profile constructed when only remnants of a terrace remain?; What does an undeformed terrace look like? Did the profile of a terrace originally mimic the modern stream profile?

Fundamental guidelines for the use of terraces as tectonic indicators have never been laid out in a systematic manner, and the student of terraces is left to deduce the fundamental concepts/schools of thought from geomorphology texts and studies that have attempted to use terraces in this way. In this study of the moderately active Pajarito Fault Zone in the Jemez Mountains, I am concerned with determining in what way relatively small watersheds respond to tectonic movements, and where the record of this response is preserved within fluvial deposits/terraces. Since I have found no systematic treatment of this subject, I will begin by outlining the elements necessary for a tectonic study of fluvial deposits and defining important terms. I will then examine the fundamental principles that control the formation of different types of terraces, outline several theoretical frameworks within which terraces have been viewed, and then attempt to apply what I have learned to the terraces of Cochiti Canyon.

Elements Necessary for a Tectonic Study of Stream Terraces

The first thing that must be established before a terrace study can begin is the existence of both terraces and some exposure of associated sediments. Exposures of sediments under or overlying terrace surfaces must be available in order to distinguish between various types of terraces. The next step is to attempt to correlate terraces within and between individual drainage basins. This can be accomplished in one of several ways.

1) Correlation of terrace remnants by elevation above the modern stream bed. This method of correlation relies on the assumption that the modern stream bed and the stream bed at the time of terrace formation had nearly identical longitudinal profiles. Another way to state this is to say that the stream was graded when the terrace formed. A graded stream is one whos slope is delicately adjusted so that it can transport all sediment supplied from upstream with the available discharge. Such a stream is in a state of equilibrium (is neither aggrading or incising), and will absorb (through changes in hydraulic geomitry) the impact of any change in controlling factors such that the impact of the change on overall stream characteristics is minimized (Mackin, 1948).

2) Soil-stratigraphic correlation. This method utilizes the fact that processes associated with soil development causes progressive, irreversible chemical and physical changes to terrace sediments with time. Individual soil properties that change with time can include: development of soil structure, reddening of deposits due to accumulation of sesquioxides, darkening and thickening of the upper part of the soil profile due to accumulation of organic matter, accumulation of carbonates within the soil profile, translocation and accumulation of clay, translocation and accumulation of calcium carbonate,

and weathering of clasts within deposits. Which property(s) will be appropriate for the description of a soil (and correlation of associated terraces) depends on the specific environment in which soil development has taken place (see Birkeland., 1984). One must also be able to demonstrate that the factors controlling soil development are the same for all terraces or that they have varied in a known way in time and space.

Potential problems with this method include: Truncation of soil profiles by erosion; increasing complexity of soil genesis in older soils; convergence of profile development with time due to the exponentially decaying rate of development for many soil properties; "retrograde" soil development due to degradation of properties in oldest soils, local variability in soil-forming factors (parent material, climate, relief, organisms, ...) ; and inability to properly characterize the parent material from (or within) which a soil formed.

It must be emphasized that soil development does not provide absolute ages for deposits/terraces. Even when radiometric ages of some deposits can be obtained, soil development rates cannot be *perfectly* calibrated because soil development does not necessarily begin at the time of deposition of the dated material and because soil properties do not develop in a linear fashion.

3) Multi-Parameter methods of correlation. This method is analogous to the use of soil development for terrace correlation, in that both use the progressive development of some property as a proxy for the relative age of deposits. Commonly used properties include: degree of clast weathering, clast weathering-rind thickness, destruction of primary depositional features (e.g. bar-and-swale topography), development of desert pavements, and development of varnish on surficial clasts. This method is useful only when

the processes that produce these features are understood and when the features do not reach a "steady-state" too rapidly.

4) <u>Correlation of terraces based on radiometric age-dating.</u> Radiometric age-control is highly desirable due to its assumed objectivity and accuracy, but it is not commonly possible.

The third step in a terrace study is to establish that variations in lithology are not being exploited by erosional processes to create flat surfaces. Such surfaces are often referred to as 'structural benches'. These benches are of little utility in tectonic studies because they are the result of changes in bedrock properties, and because their original orientation is a function of geologic structure, rather than some fluvial process.

As alluded to earlier, one other parameter that must be considered in any terrace/tectonic study is the temporal scale which the study will address. This is a vitally important consideration because the rates and processes which can be considered "possible" on one scale may be unreasonable on another (e.g. you cannot invoke migrating nickpoints traveling tens of kilometers if you referring to Holocene, strath terraces cut in resistant bedrock). It is also wise to remember that the rates of some processes are poorly constrained (e.g. bedrock erosion rates).

The objective of any terrace study ought to be the construction of "timelines" through a landscape, and in this respect any type of terrace will suffice. Once these timelines are established, and *if* their original longitudinal profiles/relationships can be determined, *then* inferences about the evolution of landscapes can be made. The construction of timelines implicitly assumes that terrace treads are not diachronous. The validity of this assumption is only testable where very good radiometric age control is available. This assumption has been shown to be invalid for terraces of Cajon Creek by Weldon (1989). However, Weldon's study concerns a large system (~35 km from headwaters to mountain-front) and small segments (1-2 km) of stream terraces may reasonably be assumed to be syncronous.

Genesis of Different Types of Terraces.

Although the terms strath, fill, and cut terrace do not have a direct genetic significance, some inferences can be made about the conditions necessary for their formation. In the following section of this paper, I attempt to create a comprehensive list of the variables that affect the creation of different types of terraces. This list is taken from published articles that present models for terrace genesis and from consideration of the fundamental variables within a fluvial system and the ways in which these variables could be changed. It is important to remember, however, that different types of terraces are produced by different processes and therefore may have different characteristics.

Fill Terraces

In order for a fill terrace to form, the sediment supply rate must exceed the sediment transport rate for some reach of a stream. Such a condition can occur by adjustments to either of these principle variables, and such adjustments can be caused by the interaction of a plethora of dependent and theoretically independent variables.

Sediment supply rate can be increased by:

Periglacial processes carrying sediment to the stream. These processes include:

Frost shattering (also increases roughness of stream bottom)

Solifluction

Glacial erosion

Increases in the rate of hillslope erosion processes including:

Increases in the rate or number of slumping/mass movements by:

Increasing precipitation relative to vegetation cover

Increasing earthquake frequency

Influxes of eolian sediment from outside the drainage basin

Unroofing of progressively less resistant rocks(?)

Sediment transport rate can be reduced by:

Decreasing the gradient of a river by

Increasing sinuosity

Raising base level by tectonic or eustatic/local base level changes

Increasing roughness of the stream channel

Decreasing discharge

Many of the changes in sediment transport rate can be viewed simply as changes to the variables within the Manning equation for stream velocity:

$$V=1.49/n\ S^{.5}R^{.75}$$

where V=velocity, S=slope, R=hydraulic radius, and n=mannings roughness coefficient (Ritter, 1978, p. 223).

It is interesting to note further that most of the variables controlling fill terrace formation are most easily affected by changes in climate. The exceptions to this generalization include changes in gradient initiated by tectonic movements (ponding) and increases in frequency of earthquakes that cause mass movements.

Strath Terraces

Strath terrace genesis is (in my mind at least) more complicated than fill terrace genesis, as more complex interactions affect the behavior of streams incising bedrock, and because fundamental questions regarding strath genesis appear to be unresolved. Some of these "fundamental questions" will be discussed below. To produce a strath terrace, a stream must erode laterally into bedrock to create a strath, and then incise vertically to create a strath terrace. Vertical incision must be rapid relative to lateral incision, or occur at a location in the stream valley that is removed from a portion of the strath if a strath terrace is to be preserved.

The various ways in which strath terrace genesis might be initiated do not lend themselves to an outline-form discussion as was given for fill terraces above. I will therefore present models of strath terrace genesis within the context of a discussion of models of terrace formation in general.

Models of Terrace Genesis

General Paradigm

The fundamental concepts/observations used to constrain models of terrace genesis are: 1) The existence of a local or regional base level of erosion (Bull, 1991) that all streams are attempting to reach. 2) The idea that a river can become graded (Mackin, 1948). By "graded" it is meant that a stream becomes precisely adjusted to the variables influencing it such that its slope is just steep enough to transport the sediment provided to it. To be graded, then, a stream must be at its base level of erosion. 3) The concept of equilibrium (taken directly from Le Chatelier's principle as used by chemists). The idea of a river as a dynamic system of energy and mass transfer allows inferences to be made about how a river will adjust when it is taken out of equilibrium by a change in some external variable (e.g. climate, tectonics,...)

It is commonly presumed that a stream will not erode laterally if it is above its base level of erosion. This implies that the base level of erosion must first be reached before a strath can be formed by lateral incision. In order for a strath terrace to be formed, then, a stream must be raised above its base level of erosion (by tectonic movement) or the base level of erosion must be lowered by baselevel fall. The later mechanism for creation of a strath is thought to be relatively ineffectual on a basin-wide scale, as shown by the work of Leopold and Bull (1979). Their studies have shown that base level change was transmitted a relatively short way upstream from the point of baselevel change. However, their field work was limited to short time scales, small basins, and sedimentary bedrock. It may be that baselevel change is transmitted throughout a basin given enough time and/or appropriate bedrock characteristics (e.g. Pazzaglia and Gardner, 1994).

Bull (1991)

Bull (1991) proposes that a differentiation between major and minor terraces can be made. Major terraces are relatively continuous along a valley and are usually paired across a valley. Unpaired, discontinuous terrace remnants are termed minor terraces. Minor terraces are the result of adjustments between dependent variables and intrinsic thresholds of sediment transport (Schumm, *et al*, 1987). For example, armoring of stream beds may momentarily "force" lateral incision in a valley that is incising vertically through a fill or through bedrock. Infrequent, high discharge events will break up this armor and rejuvenate vertical incision, producing a cut terrace or minor strath terrace.

In Bull's model, major fills are the result of climate-driven changes in sediment supply and/or transport capacity. Major strath terraces are the fundamental indicator of long-term, net tectonic uplift. The reasoning for this designation is as follows: For a major strath to form, a stream must incise laterally along a substantial reach. This is equivalent to saying that the stream must reach its base level of erosion, because a stream which is out of equilibrium (away from its base level of erosion) will expend the bulk of its available power in vertical incision. Indeed, the rate of vertical incision will be proportional to the degree of disequilibrium in the system. Now, once the stream has reached the base level of erosion, the only way for a strath *terrace* to form is for vertical incision to begin anew. The only way that the vertical space needed for incision can be produced is by uplifting the stream away from its base level of erosion. Thus, major straths are direct evidence of net uplift. Implicit in this argument are the ideas that: 1) uplift is episodic or the rate of uplift is variable in time, 2) incision rates are greater than long-term uplift rates, and 3) that baselevel change does not induce strath terrace formation far from the point of base level change.

Merritts and others (1994)

In a recent article, Merritts *et al* .(1994) attempt to "..provide the earth scientist considering the use of terraces as tectonic indicators with guidelines to their analysis and interpretation." (Merritts *et al*, 1994, p. 14,033). As previous workers have, they begin by reviewing the concept of the graded river, which is fundamental to all modern conceptual frameworks of fluvial system change (e.g. Mackin, 1948; Bull, 1990). They then describe the climate, lithology, and tectonic setting of their chosen study area, the Mattole River Basin of Northern California. The most important observations they draw from their investigation are the following:

1) The age of deposits overlying a strath is time-transgressive, and the time of formation of the surface overlying these deposits (and any soil developed on that surface) may therefore be different (by up to 3,000 years in this case) from formation of the underlying strath terrace.

2) Eustatic sea level (and thus baselevel) change has caused valley backfilling for tens of kilometers upstream from the mouth of the Mattole. Most importantly, eustatic changes have not affected the fluvial system upstream of this depositional wedge. The gradient of the fill terraces associated with this wedge is approximately one-half the gradient of the modern river; an observation that agrees with base level change studies conducted on smaller, artificially-dammed drainages in the southwestern U.S. (e.g. Leopold and Bull, 1979; Leopold, 1978).

3) Upstream of the eustatically-driven depositional wedge, but far enough downstream so that stream power is greater than some threshold level, the Mattole is constantly cutting strath terraces. The model developed to explain this scenario relies on the assumptions that incision rate is directly proportional to uplift rate and that a stream will expend all of its energy in vertical incision unless it has "excess" energy to expend in lateral incision into bedrock. In this model, the middle reach of the Mattole is constantly incising *both* laterally and vertically because it is unaffected by base level change but has more stream power than is required to keep up with uplift. However, the amplitude of lateral migration constantly decreases if strath terraces are to be preserved. The decreasing amplitude of migration allows a

portion of strath cut during one lateral swing across the river valley to be preserved and abandoned as a strath terrace upon further incision (see Merritts *et al.* , 1994 figure 8).

The idea that a balance between the rate of lateral vs. vertical incision should be considered is an important contribution, but the mechanism by which a stream segment "feels" its uplift rate is unclear. Taking the model proposed by Gilbert (1877) that each stream segment acts as the local baselevel for the adjoining upstream segment, then the only way that the incision rate of an entire stream can respond to uplift is if uplift rate increases in an upstream direction. This appears to be the case in the Mattole River Basin, but this point is not specifically addressed by Merritts *et al.* (1994).

Conceptual Model Developed from Consideration of these Studies

Considering the fundamental changes that must take place in a river system in order for a strath terrace (net vertical incision after some lateral incision) or a fill terrace (net aggradation of sediment) to form, it is possible to construct alternative methods of terrace formation. Although these scenarios are not constrained by data I believe they demonstrate the limitations of any particular terrace-genesis model that is chosen. For example:

As a stream incised through a rock sequence, could changes in rock type initiate the formation of terraces? If so, a strath could be formed due to a change in the lateral:vertical incision ratio even if climate and tectonic influences were perfectly constant. A fill terrace could be formed if the sediment entering the stream through erosion of a ‘new’ type of bedrock was harder to transport (i.e. contained more gravel) than that previously provided. This could occur even if the rate of sediment supply remained constant.

Although it seems evident that major strath terraces cannot form without net incision, a climatically driven change in discharge could change the uplift to incision ratio of a stream and thus allow a stream to reach its base level of erosion and create a strath. Although the record of this change would only be preserved if net uplift caused the creation of a strath terrace, the *genesis* of the strath itself would be fundamentally controlled by climate. For example, after eruption of the Bandelier tuff, streams attempted to cut through this deposit to reach their base level of erosion. As they incised, climate-driven changes in discharge could have initiated the formation of terraces. This same type of complex interaction has been pointed out by Bull (1991), but in a slightly different way. Bull notes that a stream can only cut a strath when sediment supply rate (climatically controlled in most cases) does not exceed transport rate. It is also possible to envision purely intrinsic controls (those that are not controlled by external variables, but by naturally occurring thresholds within the stream system) initiating terrace formation as streams cut through the Bandelier Tuff.

It is apparent that stream terrace genesis is a highly complex process, and that multiple "pathways" can lead to the creation of otherwise identical terraces. Complexity and equifinality do not preclude the use of terraces as indicators of tectonic movements, however. Even if the genesis of a specific terrace is shrouded in the mists of the geologic past, its original gradient may be reasonably inferred if the terace deposit is similar to the modern stream deposits. Even if a distinction cannot be made between terraces produced by climate-driven or tectonically-driven sediment supply rate increase, the relationship between the profiles of successive terraces can reveal otherwise hidden faults or areas of uplift or subsidence.

From the above discussion it can be seen that for terraces to be of use in tectonic studies several conditions must be established:

First, terraces must be properly correlated. If terraces are improperly correlated, then any inferences made from the reconstructed profiles of these terraces will have no bearing on the actual geologic history of a region (e.g. Meritts, *et al.,* 1994).

Second, the time scale under consideration must be explicitly defined, and some information about rates of processes on that time scale must be used to constrain hypothesis/judge conclusions.

Third, differences in rock type, climatic history, and basin size must be avoided or corrected for. Rock type should ideally be homogeneous within a basin or changes in rock type should be well-known and accounted for when examining terrace profiles.

Questions that Remain Unanswered

After long consideration and a review of a large number of published articles that profess to use terraces to interpret tectonics, I have concluded that there are many assumptions that are made about their genesis that are not backed up by objective evidence. A partial list of questions that I have formulated about specifics of terrace genesis, or about the theoretical frameworks from which terraces are commonly viewed follows. Fundamental questions necessary for understanding terrace formation and using terraces as indicators of tectonic activity include:

1. Is uplift episodic or continuous?

2. Does the nature of uplift depend on the specific study area or the time scale under consideration?

3. With question two in mind, are average rates of real value?

4. What type of terrace has a higher gradient. Is one type always steeper? Does this change across climatic gradients?

5. Are streams ever perfectly graded throughout their length? If yes, on what time and space scales is this true?

6. How is (or *is*) baselevel change transmitted to upstream reaches?

7. Are incision rates truly proportional to uplift rates?

8. Are incision rates constant or are they a function of degree of disequilibrium (i.e. do they decay exponentially in time after an uplift 'episode')?

9. How are changes in stream power with time reflected in incision rates?

10. What is the three-dimensional shape of a strath--how steep are they perpendicular and parallel to flow?

11. What fundamentally controls strath formation--uplift alone? Or is it some combination of attainment of the base level of erosion, lateral vs. vertical incision rate, and sediment supply vs. stream power, ...

12. Does Le Chatelie's principle really have any place in our thinking about the fiendishly complex *open* systems of interest to geomorphologists?

13. Why do some valleys have terraces while others do not?

Application

Until now the discussion of terraces as indicators of tectonic movements has been somewhat theoretical. Let us now turn back toward the area which is the specific focus of this study. As mentioned before, this site was chosen because it possesses some of the only preserved terraces along the entire length of the Pajarito Fault System (figure 12, 13, 14).

Description and Correlation of Terraces

A stream terrace can be defined as a valley floor that has been abandoned due to stream incision (Bull, 1991). In Cochiti Canyon, ancient valley floors have been eroded such that they are preserved only locally. The preserved sections are referred to here as terrace remnants. Sediments associated with terrace remnants in Cochiti Canyon range from a few square meters to more than a square kilometer in surface area, are dominantly gravel, and are covered by a moderately developed desert pavement. Correlation of these terrace remnants has been carried out using field mapping and by comparing soil development in terrace deposits. Because soil development is not only a tool used by the student of terraces, but is also a subject of interest in itself, the nature of soil development in the terrace deposits of Cochiti Canyon is discussed more fully in another section of this report. Comments on soil development in the section immediately below are confined to those related to correlation of terrace remnants.

The Ridge Terrace

The Ridge terrace is found approximately 100 m above the active channel of Cochiti Canyon, is rarely more than 20 meters wide, and is almost continuously exposed for >2 km down valley along the divide between Cochiti and Bland Canyons (figure 12, 15). Deposits associated with this terrace are composed of 1-3 meters of stream gravels that unconformably overlie lower Bandelier Tuff. Terrace remnants are thus part of a strath terrace as defined by Bull (1991). Channelized flow patterns have been observed on Ridge terrace remnants immediately after rainstorms, and erosion associated with this type of flow has lead to local relief of 1-3 m. A well developed soil, as evidenced by stage III carbonate morphology (after Gile *et al.*, 1966) and a 20-40 cm thick Bt Horizon (table 3) has formed within terrace deposits of the Ridge terrace. Correlation of remnants of the Ridge terrace was fairly straight-forward. The bulk of the remnants lie along the southern margin of Cochiti Canyon and can be correlated based on their juxtaposition alone since they are only separated by small saddles (figure 12, 15), and because the Ridge terrace occupies a fairly unique landscape position. Where there was any doubt about correlation of terrace remnants, a 20-30 cm deep pit was dug and the color and structure of the Bt horizon of the exposed soil was compared with samples from pits 1 and 2 (figure 12, appendix 3). The Ridge terrace is offset across the Dixon fault. Since correlation across this fault was particularly important, complete soil pits were dug as close to the trace of the fault as possible while still choosing a position that promised to afford the most complete, undisturbed soil profile. Soil pit 1 is located on the upthrown side of this fault, and soil pit 2 on the downthrown side (figure 12). These soil profiles were dissimilar in some respects, but this appears to be due primarily to the shallow depth of bedrock in pit 2 (appendix 3). The degree of development of structure and color in both the Bt and Bk horizons in these profiles indicated that they had developed for a similar amount of time and allowed me to estimate offset across the fault at this location with a high degree of confidence.

The Cañada Terrace

This terrace is best preserved just down-valley of the historic townsite of Cañada, but relatively broad remnants occur on both sides of Cochiti Canyon at a level approximately 45 m above the modern stream channel (figure 16). This terrace is a strath cut on both lower Bandelier Tuff and pre-Bandelier fluvial gravel deposits, and overlain by 3-5 meters of fluvial gravel. Along the southwestern side of Cochiti Canyon a colluvial wedge (apparently derived from the Ridge terrace) apparently covers the southwestern (upslope) portions of Cañada terrace remnants. It is unclear weather this colluvium was deposited during a single episode or how old it is. Some wedges of sediment are fine-grained and others are coarse. Some wedges of sediment contain moderately-developed soil profiles both at the surface and buried within the colluvium, while others are presently aggrading (based on observations of recent deposits after storms and the fortuitous recovery of a golf ball from within a reconnaissance soil pit). Since the Cañada terrace is located 60 m below and is partially covered by colluvium from the Ridge terrace, it must be a somewhat younger geomorphic feature. However, soil development (as measured by carbonate and clay content) on this terrace is even more pronounced than on the higher Ridge terrace (table 3). The age of a portion of this surface near the Rio Grande has been estimated using calibrated varnish-cation ratios of surface clasts as between 240-350 ka (Dethier and Harrington, 1987). This dating technique is highly suspect, but this age is used in this report because it is the only numeric age estimate ever obtained from within Cochiti Canyon, and more importantly this age range is resonable based on the degree of soil development observed in Cañada terrace deposits.

Once I had become familiar with the soils of Cochiti Canyon, assignment of remnants to the Cañada terrace was extreamly easy. Complete soil profiles were described on the upthrown and downthrown block of the Dixon fault (pits 5 and 6, figure 12) for the same reason that similar locations were chosen on the Ridge terrace. Both of these soil pits exposed an extreamly clay-rich Bt horizon (table 3). The surface of the Cañada terrace remnants at soil pits 5 and 6 was covered with a pavement of clasts that was more well developed than the pavement on adjacent, lower remnants. Based on both landscape position and the degree of development of this pavement, reconnaissance pits were dug on other terrace remnants that I believed were part of the Cañada terrace. These shallow pits revealed that this horizon was widespread and fairly well-preserved. With time, it became possible to identify a remnant of the Cañada terrace by merely observing the surface and then driving a pick 10-15 cm into the soil. If the pick felt as though is was being driven into a deposit of chewing-gum and came out of the ground with large clumps of red clay stuck to its end, I felt confident in my correlation.

There were two locations, however, that defied this simple means of correlation. Pit 7 is located on a small remnant of a strath terrace on the southwestern side of a tributary to Cochiti Canyon (figure 12). Lower Bandelier Tuff is well exposed in the cutbank of this tributary and the site seemed ideal for a soil pit, but the soil exposed was unlike any of the previously described soils. The lower part of the soil profile is similar to the Cañada soil seen in pits 5 and 6, but the upper portion lacked the clay-rich Bt horizon of these pits and seemed to have more silt. Because the upper part of the soil could easily have been stripped on this narrow remnant, a second pit was dug 100 meters to the southeast. This pit revealed a sandy, quarts-rich deposit with a 1.5 m deep inset channel-fill. Soil development (as measured by clay and carbonate content and development of soil structure) was negligible, and I took the channel as confirmation of recent stripping of this surface. In this context, I was comfortable in assigning this remnant to the Cañada terrace. Surveying of the lower contact of the terrace deposits showed that this remnant was at the correct height above grade to be correlated with the Cañada terrace in this part of the canyon.

The large terrace remnant at the mouth of Cochiti Canyon (figure 12) is covered in many places with a 1-2 m thick layer of eolian(?) silt and, on the downthrown side of the Basalt splay, with a 2-15(?) m thick silty colluvial wedge. This large terrace remnant was the site chosen by Dethier and Harrington (1987) for analysis of varnish-cation ratios, and it was therefore critical that it be correctly correlated. It was clear from its height above grade that this remnant was either a part of the Ridge terrace, the Cañada terrace, or some intermediate terrace. Once I had dug reconnaissance pits in the small, higher remnant to the north (figure 12), I was confident that that remnant was a part of the Ridge terrace. I was not succesful in digging a soil pit through the silt layer *and* within an undisturbed portion of the terrace deposits on the large terrace remnant. I was finally convinced that it was a part of the Cañada terrace by the presence of clay coatings on clasts in shallow soil pits on disturbed portions of this terrace remnant that were similar in color and texture to clay in the Bt horizons of pits 5 and 6, and by the morphology of carbonate in other, similar pits. This remnant has apparently had a complex history of recently active surface processes, yet a soil similar to the Cañada soil was at one time developed here. Surveying of the height of terrace remnants ultimately helped to confirm the assinment of this remnant to the Cañada terrace.

<u>The Rio Terrace</u>

The Rio terrace is found primarily as a strath terrace developed on pre-Bandelier gravels on the northeastern side of Cochiti Canyon (figure 17). Two to five meters of gravel overlie the lower bounding surface, and the terrace lies approximately 20 m above modern grade. Soil developed on this terrace has a moderately well developed Bt and Bk Horizon, but laboratory analysis show that less total carbonate and clay has accumulated here than on the two higher surfaces (table 3).

Correlation of this terrace was largely a matter of eliminating the posibility that a terrace remnant was a part of the Cañada terrace based on height above grade, or that it was a part of the Ash terrace. The later was easily done by digging a reconnaissance pit in any suspect deposit. If a reddened, relatively clay-rich horizon was encountered in such a pit then the terrace remnant could be confidently assigned to the Rio terrace. The possibility still remained that several terraces of similar height above grade and degree of soil development were being mapped as one, distinct terrace. The resolution of the later surveying did not allow this possibility to be eliminated, but where this terrace is offset by the Dixon fault the degree of soil development in cut-bank exposures is very much like that seen in pits 3 and 8 (figure 12), and a single terrace can be confidently traced on either side of the fault.

<u>The Ash Terrace</u>

Although not as extensive as the other three terraces, the Ash terrace is important to this study because 0.5-1 meter of ash associated with the El Cajete

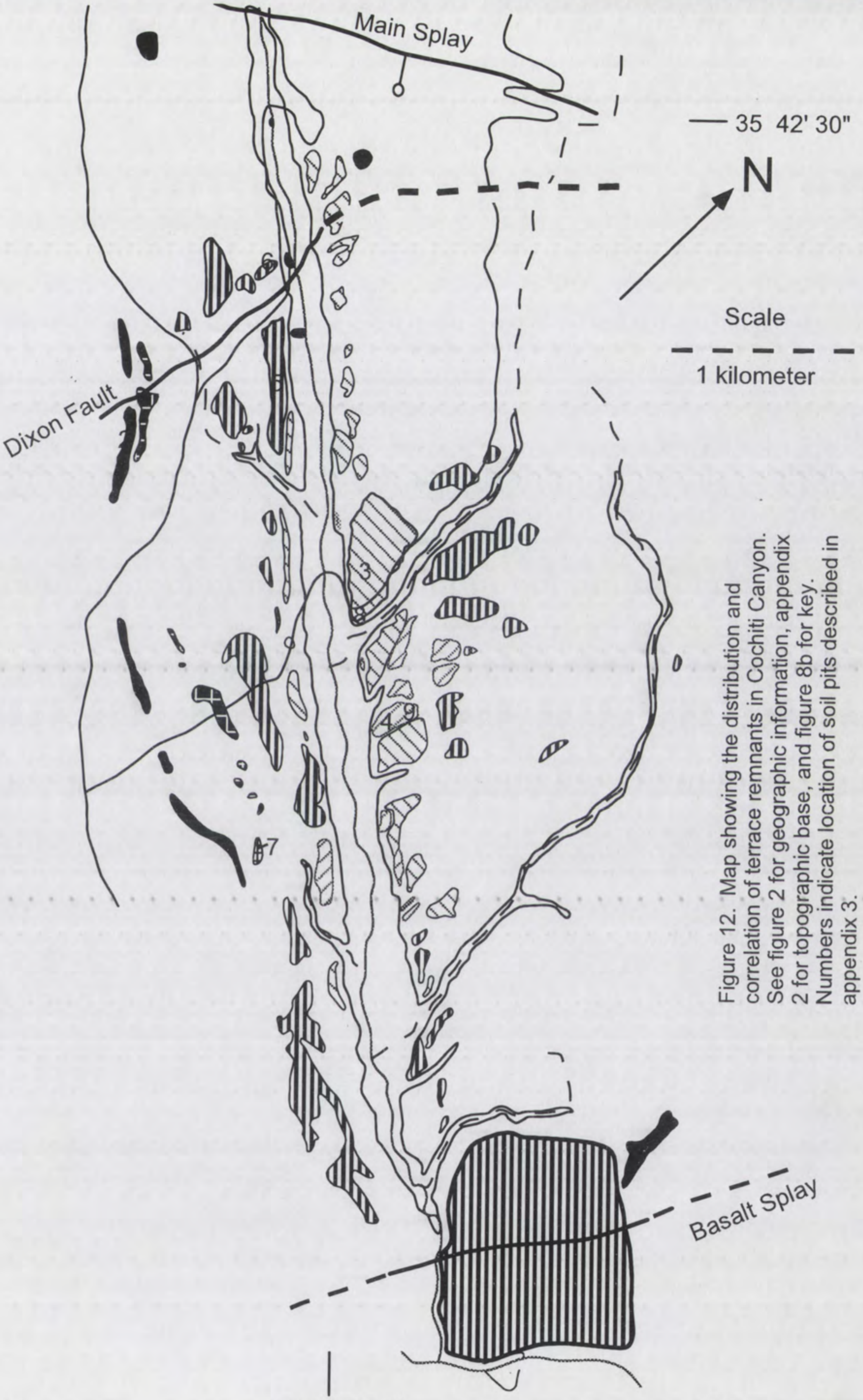


Figure 12. Map showing the distribution and correlation of terrace remnants in Cochiti Canyon. See figure 2 for geographic information, appendix 2 for topographic base, and figure 8b for key. Numbers indicate location of soil pits described in appendix 3.

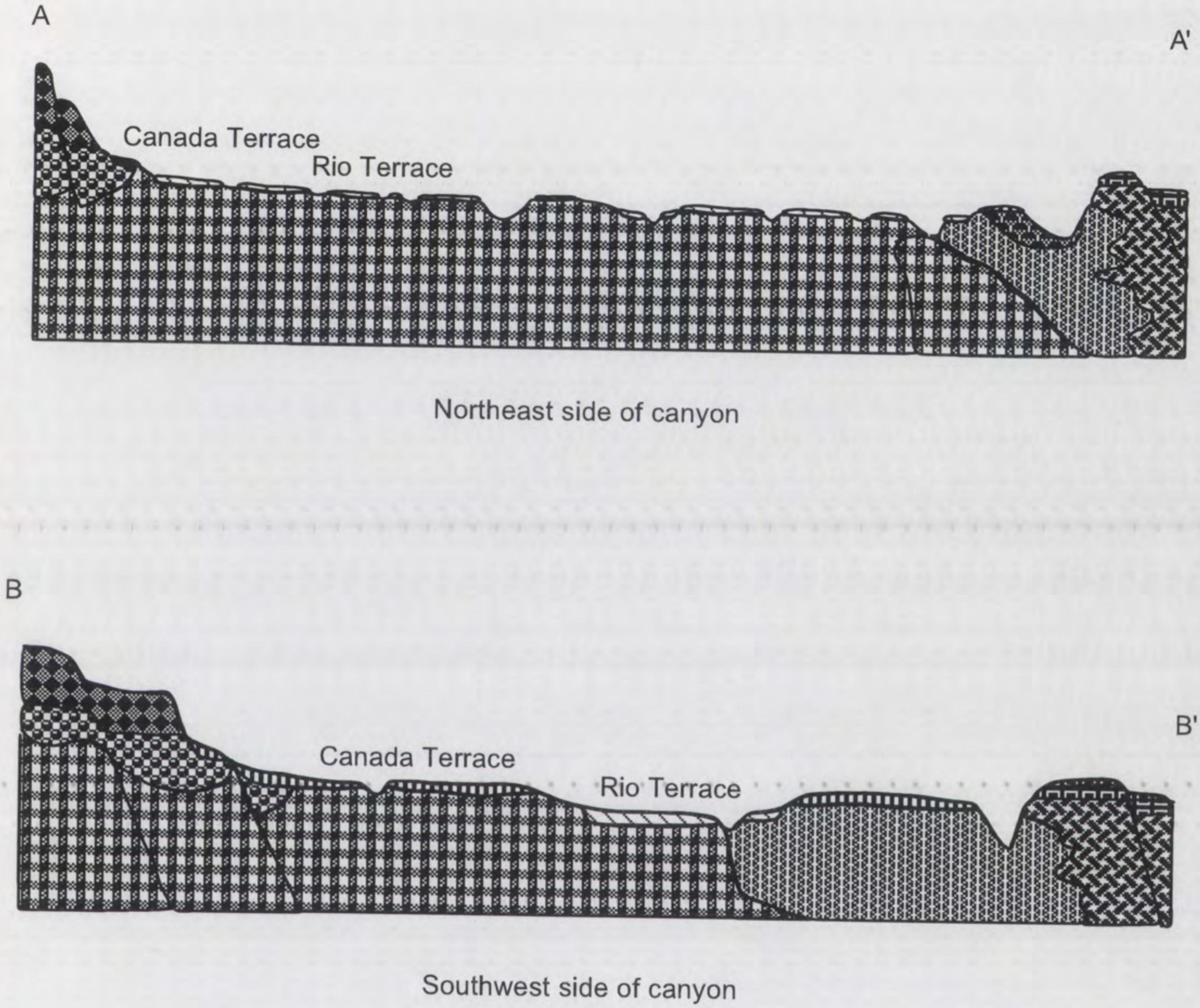


Figure 13. Schematic cross sections along either side of Cochiti Canyon showing relations of units underlying terrace gravels. For Key see figure 8b, and figure 2 for approximate location of sections.

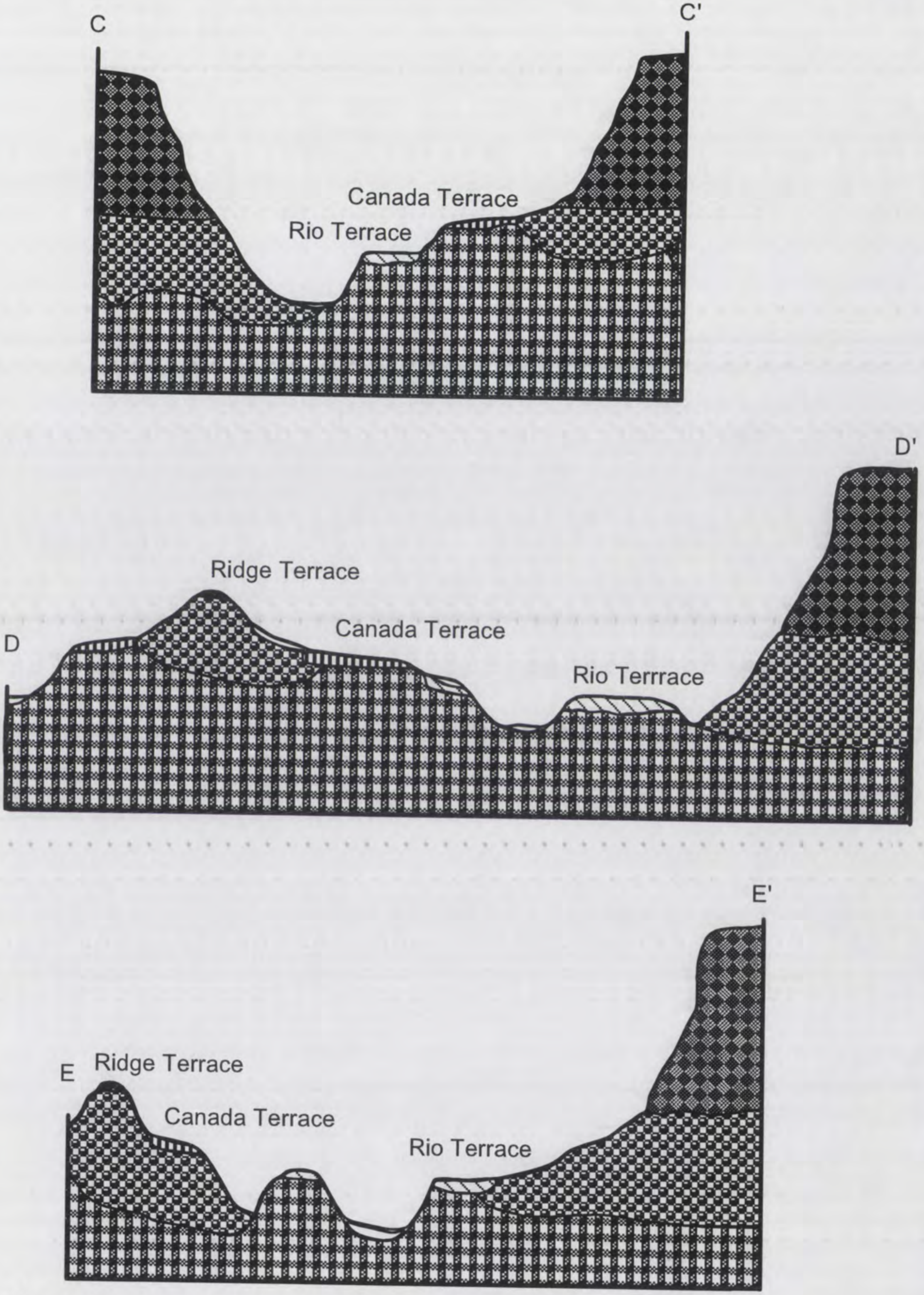


Figure 14. Schematic cross sections perpendicular to Cochiti Canyon. See figure 2 for approximate location of sections and figure 8b for key.

A

B

Figure 15. Photographs of the Ridge terrace. A) View of the Ridge terrace surface near soil pit 1 (figure 12). Gravel pavement visible in this view is underlain by a silty Av horivon. White clasts in foreground are coated with pedogenic calcium carbonate and have apparently been brought to the surface by burrowing animals. B) . The Ridge terrace makes up the skyline in this photograph. This view is from the Ash terrace near pit 4 looking southwest across Cochiti Canyon. The Cañada terrace can be faintly discerned in the middle ground, on the southwest side of the canyon.

A

B

Figure 16. Photographs of the Cañada terrace. A) This view is looking from the site of soil pit 6 (see figure 12), on an upthrown portion of the Cañada terrace, across the Dixon fault to the townsite of Cañada on the downthrown block. Note building of townsite in light-colored area near right side of photograph. B) This view is 180 degrees from 16A, looking from soil pit 5 toward soil pit 6. The scarp of the Dixon fault forms the tree-covered rise in the middle ground. Faint trail visible along the center of the photograph leads to the townsite of Cañada.

A

B

Figure 17. Photographs of the Rio terrace. A) Surface of Rio terrace looking toward soil pit 8 (note mound of soil in center of photograph). Surface in foreground is mantled by a thin colluvial(?) layer that pinches out away from viewer. B) View from the Cañada terrace near soil pit 5 across Cochiti Canyon to Rio terrace in middle ground, Cañada terrace further back, and Mesas of Bandelier Tuff on skyline and on right side of view. This photograph shows a typical view of the terraces.

A

B

Figure 18. Photographs of the Ash terrace. A) Surface of Ash terrace at soil pit 9. Note pedogenic carbonate coatings on clasts in mound of soil next to soil pit. This view exaggerates the pavement cover to some extent. B) View from near soil pits 4, 9 to northwest across a small tributary of Cochiti Canyon. Small white patch in center of photograph is El Cajete ash in Ash terrace fill. Ash terrace is inset here (and everywhere) into the Rio terrace, which forms the green horizon in the middle of this view.

Table 3: Laboratory Data for Soils of Cochiti Canyon

Soil* and Horizon	Horizon Thickness (cm)	Sand %	Silt %	Clay %	Percent Carbonate in <2mm fraction	Percent Carbonate in >2mm fraction	Percent of Sample <2mm
Ridge Soil							
Pit 1							
Av	5.0	42.35	54.3	3.4	0.5	2.4	30.7
Bw	10.0	48.95	34.3	16.8	0.5	0.6	61.4
Bt	20.0	56.65	8.1	35.3	1.1	1.0	38.6
Btk	18.0	40.75	40.2	19.1	18.3	14.9	46.9
Bk	20+	13.85	83.4	2.8	20.9	23.7	62.4
Pit 2							
Av	10.0	59.4	28.5	12.2	0.6	0.8	55.1
Bt1	25.0	54.4	9.5	36.2	1.4	1.0	22.8
Bt2	15.0	40.5	23.5	36.1	0.9	1.0	42.1
Bk	10.0	23.8	55.3	20.9	18.3	23.6	39.8
R	20+	7.35	89.6	3.1	2.5	2.7	51.6
Canada Soil							
Pit 5							
Av	6.0	55.1	31.8	13.1	0.4	1.0	78.2
Bt1	11.0	14.05	15.4	70.6	0.6	1.0	94.7
Bt2	19.0	20.95	16.5	62.6	1.3	0.7	75.0
Btk	7.0	35.4	12.7	51.9	13.3	14.4	34.9
Bk	27+	30.65	51.0	18.4	38.7	32.0	24.3
Pit 6							
Av	8.0	49.75	37.6	12.7	0.7	0.4	43.7
Bt1	25.0	22.6	10.9	66.6	0.5	0.6	65.7
Bt2	15.0	38.1	7.7	54.2	0.4	0.6	26.8
Btk1	12.0	54.2	10.1	35.8	5.1	1.0	25.0
Btk2	18.0	53.45	9.7	36.9	7.5	1.9	24.3
Bk1	19.0	72.6	10.0	17.4	27.1	4.1	18.4
Bk2	23+	84.55	6.1	9.3	11.1	4.4	12.8
Rio Soil							
Pit 3							
Av	10.0	63.05	24.6	12.4	0.5	0.5	90.7
Bw	18.0	33.1	36.3	30.7	0.6	0.7	27.5
Bt1	17.0	29.6	21.2	49.3	0.7	0.8	25.1
Bt2	10.0	49.05	21.8	29.2	0.4	1.2	58.6
Bk	40.0	31.9	52.9	15.3	13.4	14.1	25.4
C	10+	13.7	84.6	1.7	1.2	1	27.9
Pit 8							
Av	5.0	54.9	35.3	9.9	0.6	1.0	53.4
Bw	18.0	34.05	38.6	27.4	0.4	1.0	91.0
Bt	17.0	21	34.8	44.2	0.4	0.9	96.6
Btk1	12.0	28.05	22.0	50.0	0.8	1.0	49.9
Btk2	11.0	51.2	15.0	33.8	7.6	1.8	28.0
Bk	37+	53.65	32.0	14.4	32.8	13.4	9.1

Table 3 Continued

Ash Soil							
Pit 4							
Av	5.0	30.8	65.4	3.9	0.8	1.2	57.0
A	11.0	35.15	56.8	8.1	8.1	14.9	57.5
Bk1	49.0	22.6	74.6	2.8	19.8	18.3	31.9
Bk2	20.0	8.8	90.6	0.6	14.0	3.7	18.4
Bk3	30+	16.75	82.7	0.5	2.5	3.2	76.3
Pit 9							
Av	3.0	59.9	33.6	6.5	1.7	7.1	78.4
A	12.0	68.85	20.0	11.2	5.5	9.8	87.5
Bk1	23.0	73.9	20.0	6.1	10.9	5.4	63.4
Bk2	35.0	78.45	18.3	3.3	11.2	6.8	68.6
Bk3	52.0	86.1	13.7	0.2	5.8	1.3	28.4
C	15+	98.55	1.2	0.2	1.3	0.8	21.1
Miscellaneous							
Pit 10							
btk	15.0	82.9	4.50	12.60	N.A	N.A	94.50

* See figure 12 for location of pits

Pumice is found within the fill beneath the terrace tread (figure 18). This fine-grained ash is found within a 5 meter thick, pebbly/gravely deposit containing no unconformities. No other deposits of ash from the El Cajete eruption are preserved in this part of the Jemez Mountains. Preservation of ash here suggests that the stream was actively aggrading at the time of eruption, since any ash left on the ground surface elsewhere was completely removed from the landscape. This eruption has been dated at approximately 60 ka (Reneau, *et al.*, 1994). The Ash terrace is a fill terrace that grades to an elevation 2-4 meters below the Rio terrace. The fill itself is more pebbly than the gravels overlying the other surfaces, probably because the streams which deposited it were small tributaries of Cochiti Canyon. This lower stream energy may also explain the preservation of the fine-grained ash. Soil development is markedly weaker than on other surfaces (table 3). Stage I+ carbonate is found in a horizon approximately 1.3 meters thick, but no appreciable clay accumulation is observed. At present, this surface provides the only numeric age control with which to calibrate rates of soil development in this area.

Correlation of Ash terrace remnants was relatively easy because the terrace is confined to a small part of Cochiti Canyon, and has a soil that is distinct in its lack of clay accumulation and universally light color (appendix 3). The correlation of terrace remnants across Cochiti Canyon was additionally confirmed by excellent exposures of a 10-15 m thick fill beneath all terrace remnants assigned to the Ash terrace.

Genesis of the Terraces

Taking the above discussion of terraces in general as a starting point, several hypothesis for the generation of terraces in Cochiti Canyon are possible. The three higher terraces of Cochiti Canyon are straths underlain by 1-5 meters of coarse gravel overlying either lower Bandelier Tuff or pre-Bandelier Gravel (figure 13, 14). Considered in this light, and applying the reasoning of Bull (1991) these terraces would be interpreted as fundamentally tectonic in origin. Episodic uplift is mandated by the mere presence of these terraces as each terrace tread represents a period when the stream first reached its base level of erosion, incised laterally for a time, and then incised vertically due to renewed uplift. However, all of these terraces have formed after the obliteration of the previous fluvial system by deposition of the upper Bandelier Tuff. The time of terrace formation may then have been influenced by climate-induced changes in the rate of lateral vs. vertical incision.

Alternatively, climate-modulated intrinsic thresholds may have controlled terrace formation. If the Rio Grande incised through the Bandelier Tuff at a constant rate, but the valley floor of Cochiti Canyon was periodically armored with cobbles that were too large to be moved by anything but extremely large floods, then terrace formation would represent times when the ancestral Rio Chiquitio simply downcut rapidly after this armor was stripped.

Any model, in order to account for the present distribution of terraces (figure 12), must incorporate a decreasing ratio of lateral to vertical incision rate through time. This is the only way in which successive *paired* terraces can be preserved. Indeed, all studies of paired terrace sequences are biased toward environments in which this has been the case, since terraces are not preserved otherwise (see discussion of Merritts *et al.* (1994), above).

It is not entirely clear which of these models (or any other model) for terrace genesis is most appropriate. The fact that terraces are unique to

Cochiti Canyon (among streams incising into Bandelier Tuff in this area) suggests that discharge (as controlled by basin size) must exert some control on terrace formation. This influence is most likely exerted by the stream's ability to attain the base level of erosion and subsequently incise laterally. Eruption of the Bandelier Tuff has obviously had a strong influence on basin evolution. These two facts lead me to favor a model of terrace formation that combines overall incision due to eruption of the tuff with climate-driven changes in discharge. Changes in baselevel controlled by incision of the Rio Grande (itself possibly climate driven) may have additionally controlled the timing of terrace formation, as the Rio Grande seems to have been incising more rapidly during the last 600 Ka (Wells, *et al.*, 1987). An additional control on baselevel is the presence of a basalt flow at the mouth of Cochiti Canyon that was extruded prior to eruption of the lower Bandelier Tuff (figure 13).

The Ash Terrace is different than the other three terraces discussed. First, this terrace was created at the mouths of tributaries of the Rio Chiquitio, and secondly it is an inset fill terrace. The sequence of events leading to its formation are as follows: 1) The Rio Chiquito incised vertically through the level at which the Rio Terrace had formed and attained a level near its present grade. 2) The tributaries that were to eventually form the Ash Terrace incised vertically and then laterally. 3) The Rio Chiquitio and its tributaries backfilled to a level several meters lower than the Rio Terrace. During backfilling, the El Cajete Pumice was erupted and ash from this eruption contributed to filling of the small side canyon. 4) The Rio Chiquito incised vertically again, as did the tributaries, leading to the present landscape configuration. Since no faults are present near Ash Terrace remnants, it can be presumed that this sequence of events occurred in response to climatic change.

I should mention at this point that small terrace remnants do exist below the level of the Ash Terrace. These terraces range from <1 meter to ~5 meters above present grade, are unpaired, and are either small fill or fill-cut terraces. No systematic correlation of these terraces has been attempted because they do not cross fault splays and therefore cannot be used to indicate tectonic movements. Their origin is almost certainly related to either intrinsic processes or climate variation since the last glacial age.

Soils

Introduction

As noted above, soil development was used in the correlation of terrace remnants in Cochiti Canyon. After examination of two 1-2 m deep soil pits on each terrrace, it was determined that several specific properties of texture and color could be used to differentiate the terraces by excavating shallow (20-50 cm) soil pits in terrace deposits. Because soil analysis plays an integral role in the correlation of terraces in this study, I offer a short discussion of soil development and the theory of soil/geomorphic analysis before discussing the specifics of soil development in Cochiti Canyon.

Soils (or specific soil properties) can be usefully seen as systems whos development at any specific point in time is a function of the local climate, relief, soil organisms, parent material, the amount of time since soil formation began, and other local conditions, such as dust influx (Jenny, 1980). When each of these variables except time is reasonably believed to have been constant (or to have varied in a systematic way through time), then soil development can be taken as an indication of relative age. The utility of soil development as a tool for relative dating of geomorphic surfaces has been greatly advanced during the last three decades. Many studies of well-dated chronosequences have shown that soils in any one area develop in a generally systematic way (e.g. Gile and Hawley, 1966; Birkeland, 1984, Burke and Birkeland, 1979;). The most important aspects of soil development *relevant to this study* are the way in which clay and carbonate content of soils in arid regions change with time. In many cases where it can be demonstrated that the parent material of a soil contains little or no clay and/or calcium carbonate, the amount of each increases in progressively older soils (e.g. Gile and Hawley, 1966; Birkeland, 1984). Analysis of mineral grains in soil parent material and the same constituents in overlying soils has demonstrated that not all the clay found in soils can come from alteration of the parent material, and isotopic evidence indicates that the calcium in soil carbonate does not come primarily from the mineral grains in the parent material (Birkeland, 1984). These observations led to the hypothesis that relatively large additions of material are accomplished by influx of eolian dust. This hypothesis is consistent with measurements of modern dust-influx rates made throughout the American southwest (Reheis, *et al.*, 1989).

Not only do pedogenic clay and carbonate content increase with time, but the morphology of the soil as a whole changes with increasing amounts of these constituents. As clay content increases, soil structure (development of peds) also increases. As carbonate content increases, different stages of carbonate morphology are observed (Gile, *et al.*, 1966). These well-documented changes in soil morphology can be used to assign relative ages to geomorphic surfaces in an area which has similar conditions of parent material, climate, dust influx, soil organisms, and relief. These conditions for the use of soils as a relative dating tool are met by the soils found in Cochiti Canyon. Climate can be assumed to be spacially (but not temporally) constant since all of the soils are within a few kilometers of each other, and relief between the highest and lowest soil is on the order of 100 m. The present vegetation is similar on all terraces and the local relief is minimal and comparable between terraces.

Since it is presumed that terraces in the study area have been deformed by faulting, their height above grade alone is insufficient for purposes of correlation (although in places where more than one terrace is present their relative position is useful). Soils were found to be the best method by which terrace remnants could be correlated. This was surprising, since many remnants are small and no surface is free of active surface erosion at present. This situation *suggests* that much of the dissection of surfaces has occurred in the geologically recent past, since soil profiles would not be preserved indefinitely on narrow terrace remnants.

Soil Forming Factors in the Cochiti Canyon Area

The Jemez Mountains, like much of New Mexico, have a semi-arid to subhumid, seasonally hot **climate**. Precipitation comes in the form of rain during the summer/fall monsoon season and mostly as snow during October-March. Average annual rainfall in the region is between 9 and 17 inches depending on elevation (Gonzalez, 1993). Precipitation and temperature regimes also vary markedly with elevation, and these changes are reflected in the **plant communities** that inhabit different parts of the mountains. In narrow parts of canyons (particularly on the upthrown blocks of the Pajarito Fault System) an "inverted" plant community is observed, where drier/hotter communities (e.g. Piñon-Juniper grassland) are found on mesa tops and wetter/cooler communities (e.g. Ponderosa Pine forest and Riparian vegetation) are found in the valleys. A whole range of plant communities from Desert Scrub to Boreal Spruce Forest can be found within the drainages studied here. The area of intense study, however, is entirely Piñon-Juniper Grassland with occasional Ponderosa Pine and minor riparian zones near perennial or near-perennial streams. Therefore, for purposes of soil analyses, climate can be considered constant (or to be changing in the same way for all soils). No gauging stations or climate stations are available within the study area itself.

Relief in the Jemez Mountains can be dramatic (e.g. mesa top to canyon bottom relief up to 1000'), but within the study area it is rather subdued. Local relief on terraces is 0-3 meters over distances of 100 meters or less, and the slope of terrace remnants is on the order of 1-2 meters per 100 meters. The **parent material** for all of the soils is entirely derived from volcanic units of the Jemez Mountains, and is dominated by gravel.

Soils Developed on Terraces in Cochiti Canyon

At least two complete soil pits were described and samples from these pits were analyzed for each terrace in Cochiti Canyon (table 3, appendix 3). Soil samples were analyzed for soil carbonate content (using a Chittik apparatus), and particle size distribution using the pipette method.

Ridge Soil

Soil developed on the Ridge terrace is *clay and carbonate rich* (in the manner of some botanical keys, I am italicizing characteristics of these soils that I find most helpful in distinguishing them). A thin Av horizon overlies well-developed Bt horizons underlain by a Bk or K horizon (Figure 19). Bk horizons are most commonly observed, but zones of laminated carbonate are occasionally observed. Bt horizons contain up to 52% clay (table 3) but have *noticeably more coarse sand* than Bt horizons of the Cañada soil (see below) The Bk horizon is developed in gravelly deposits and in some cases is observed to extend to the contact of this deposit with lower Bandelier Tuff. Where this is the case, a 2-5 cm thick laminated layer of calcium carbonate is often observed. In areas where soil has been stripped off of hillsides, this carbonate layer can be seen to have entirely covered the bedrock/alluvium contact (figure 19). Laminated carbonate can often be found as surface float on the terrace in areas where some degradation is shown by local relief, but this same

Figure 19. Photographs of the Ridge soil at soil pit 1. A thin Av horizon makes a light colored band at the surface. This silty layer is underlain directly by a Bt horizon-- indicating stripping of the surface at some time before deposition of the Av. The Bk horizon is clearly visible at the base of the profile and extends to the base of this pit. Note knife with brass-tipped handle at top of profile for scale.

A

Figure 20. Photographs of the Cañada soil. A) Soil pit 5 on the downthrown block of the Dixon fault. Note strongly developed Bt horizon above stage III carbonate horizon. Note camera case for scale. B) Soil pit 6 on the upthrown block of the Dixon fault. In this photograph the Av horizon can be seen at the top of the profile as a light colored band extending from the surface to the point where the knife enters the soil. Knife is approximately 10 cm long.

A

B

Figure 21. Photographs of the Rio soil. A) Soil pit 3 on the Rio terrace. This pit exposed relatively unaltered parent material of post-upper Bandelier stream gravels overlain by a stage III carbonate, a ~20 cm thick Bt horizon, and a thin Av horizon.. In this photograph the soil looks similar to the Cañada soil (figure 20B), but soil structure is much less developed and laboratory analysis shows that this soil has lower clay and carbonate percentages (table 4). B) Soil pit 8 on Rio terrace. This shallow pit exposes an Av horizon underlain by a 20-30 cm thick Bt horizon and a Bk horizon. Note knife for scale.

laminated material is seen only in 'pockets' between or beneath large clasts in soil pits. Bk horizons in soil pits are, in fact, notably chalky and the percent carbonate is less than that calculated for Bk horizons of the Cañada Soil (see below), particularly in pits where bedrock has been exposed. There are two possible explanations for this phenomena. First, instrumentation of the contact between the Alluvium/tuff contact near Los Alamos has demonstrated lateral flow of groundwater in this zone (McFadden, personal communication, 1996). Groundwater flowing in this manner could dissolve soil carbonate and carry it to the margins of the terrace remnant where it would then be flushed from the system as surface flow. Such a process would be particularly effective on the Ridge terrace because remnants are thin. Alternatively, the lack of an ultra-rich clay horizon (table 3) above the Bk horizon in the Ridge soil (as opposed to the Cañada soil) could lead to throughflow of water and leaching of carbonate regardless of any lateral flow. Since a period of destabilization of surfaces seems to have occurred throughout the study area (see above), it is likely that both of these processes are acting to degrade both the Bt and Bk horizons of this soil. Bk horizons throughout the study area seem to have been degraded to some extent. This observation is discussed further below.

<u>Cañada Soil</u>

The Cañada soil is the most well-developed soil preserved within the study area (table 3, figure 20). An Av horizon between 0 and 10 cm thick is underlain by a Bt horizon containing *up to 70% clay* and usually exhibiting 7.5 YR 4/4 color. This horizon is *notably darker* (higher value) than the Bt horizon of other soils, and is so clay rich that *dry peds are extremely hard.*

The Bt horizon also has a notably *low percentage of gravel*. This lack of gravel suggests that either overbank deposits have been preserved on this terrace and altered to clay with time or that eolian additions of fine material have been 'trapped' on this surface more effectively than on the Ridge terrace. Based on observations of the way that samples react to 30% hydrogen peroxide, the darkness of the Bt seems to be caused disseminated organic material. Bt horizons of this soil are underlain by a *well cemented* Bk horizon. Although sandy patches within the Bk horizon are occasionally carbonate-free and hand-samples are somewhat chalky, laminated carbonate is seen on this terrace where digging animals have left small tailings-piles. Apparently a K horizon was present within this soil but has been degraded in the same manner as the Bk (formerly K?) horizon of the Ridge soil.

Rio Soil

The Av horizon of this soil is immediately underlain by a light-orangish, Bw/Bt horizon containing up to 49% clay (table 3, figure 21). Although the clay content of the <2mm fraction is very high, this horizon is predominantly gravel (table 3), so the overall volume of clay in the soil profile is lower than for soils found on higher terraces. Bt horizons within this soil are underlain by a Bk horizon containing considerably less carbonate than older soils (table 3). In one soil pit, however, an intact K horizon was found. This horizon is developed in a bouldery portion of the terrace deposits. I speculate that channelization of downward-moving, carbonate-bearing water around these boulders caused rapid plugging of the soil and subsequent deposition of laminated carbonate.

Figure 22. Photographs of the Ash soil. A) Soil pit 9 on the Ash terrace. The only organic-rich A horizon observed in any of the soil pits is seen at the top of this profile beneath a 5 cm thick Av horizon. The entire profile effervecces in dilute hydrochloric acid, but percent carbonate is relatively low (table 4). Graduations on tape are 10 cm apart. B) Soil pit 4 on the Ash surface. This view exposes a profile much like that in 22A, and also shows the weakly cemented Bk3 horizon at the base of this pit. Note knife for scale.

Figure 23. El Cajete Ash incorporated into fill beneath Ash terrace. Exposure in cut-bank of tributary to Rio Chiquito below Pits 4 and 9 (see figure 12). Note how lower contact dips down near left edge of white layer. The material in this lower portion is coarser than the rest of the layer and appears to be a channel fill.

Ash Soil

The ash soil is much less developed than other soils in the area, as evidenced by a *lack of clay accumulation* (table 3, figure 22). This soil is developed in a relatively *pebbly/sandy* fill deposit, and contains *the only organic-matter rich A horizon* found in the study area. I attribute A horizon development to the increased infiltration capacity generated by *lack of a Bt horizon* and the sandy/silty nature of the upper part of this soil. Below the A horizon of this soil are three Bk subhorizons distinguishable by varying carbonate morphology. The C horizon reached in one pit dug in this soil is a sandy gravel.

Age of Soils and Associated Terraces

A thorough discussion of chronosequences and the use of soils to give approximate numerical ages to deposits or surfaces is beyond the scope of this report simply because the subject is so complex. In brief, where calibration of the rates of a soil property's development can be achieved, then the degree of development of that property can be used as a proxy for the age of the deposit or surface in which the soil has developed. For example, if numeric age control is available for the oldest and youngest (based on landscape position) soils in an area, and those soils have been accumulating calcium carbonate *at a constant rate,* then carbonate content of intermediate-aged soil profiles can be used to calculate an age for those soils. At least three assumptions are implicit in this technique. First, it is assumed that soils began to develop immediately after deposition ceased. Second, that no (or very little) erosion of the soil has taken place. Third, that no (or very little) carbonate has been leached from the soil. It is very rare for all of these assumptions to be entirely true. In

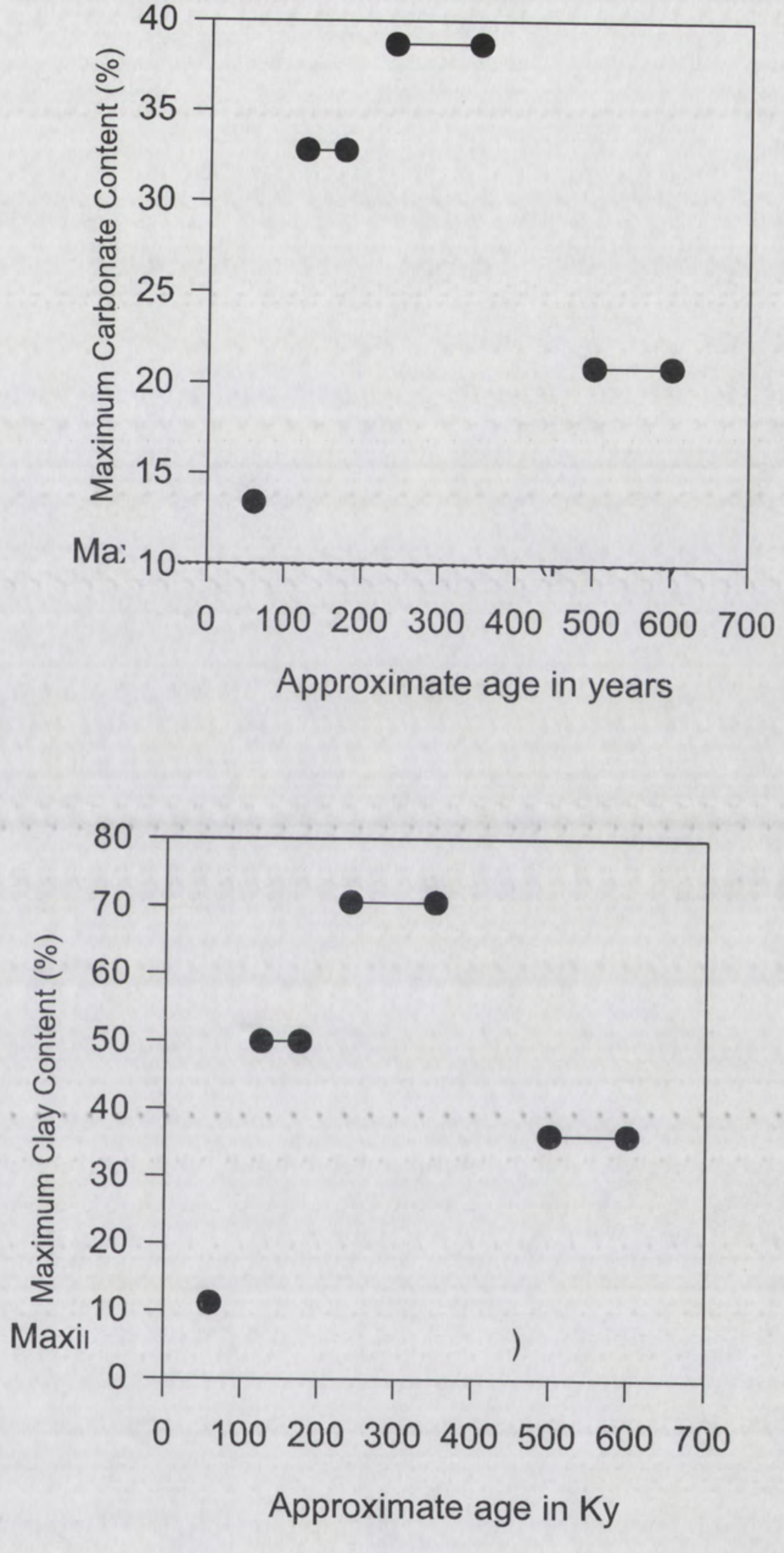


Figure 24. Plots of estimated age versus maximum clay and carbonate percentagesof soils in Cochiti Canyon. Horizontal lines indicate uncertainties in age estimates (see text for discussion)

Cochiti Canyon, for example, the upper portion of most soil profiles appears to be truncated and carbonate seems to have leached out of soil profiles where Bandelier Tuff underlies terrace gravels near the present ground surface (see discussion above). Values of soil carbonate were calculated (based on the method of Machette, 1985) in the hope that they might provide a means by which to confirm age estimates, but values varied more between different soil profiles on the same surface than they did between soils on different surfaces.

I am fortunate to have at least some age control on soil development in Cochiti Canyon. Ash from eruption of the ~60 ka El Cajete Pumice is preserved within an aggrading sequence of gravels beneath the Ash terrace (figure 23), and a portion of the Cañada terrace near the mouth of Cochiti Canyon was dated at between 240-350 ka by Dethier and Harrington (1987) based on calibrated varnish-cation ratios and amino acid racemization. Although more recent age estimates have been obtained for correlative terraces in the Rio Chama drainage, I am using this age estimate because the analysis was performed on a sample from within Cochiti Canyon. Based on these calibration points and comparison of soil development with dated chronosequences in the Southwest (e.g. Gile and Hawley,1966) it is possible to estimate that the Ridge soil has taken 0.5-0.6 My to develop to its present state, and that the Rio soil has taken between 125 and 175 ky to develop. Using these estimated and calibrated ages results in plots that show exponential rates of soil-property development for the first ~400 ka and a decrease in the rate of soil-property development after this time (figure 24). These observations agree, in general, with observations of soil development throughout the western United States (e.g. Birkeland, 1984).

I am aware that the age estimates given here could be easily debated, and that the *possible* age range of each surface is wider than that given. Since my purpose was to attempt to constrain the timing of motion on splays of the Pajarito Fault I have chosen to "stick my neck out" and give narrow estimates. As is the case in any study , my rate calculations are dependant on these age estimates and (as is often the case in such studies) I have no independent age control with which to confirm my estimates. I had originally intended to construct a graph comparing degree of soil development to age of associated deposts using data from all of the dated soil chronosequenses in the Southwest and to compare my age estimates to this graph. I found this impractical for several reasons. First, it is unlikely that the rate of development of any soil property is the same in all parts of the southwest. Indeed, the rate of accumulation of soil carbonate seems to vary widely even within the Jemez Mountains. For example, Rodgers (1996) found virtually no accumulation of carbonate on terraces along the Jemez River and Rio Guadalupe, while terraces in Cochiti Canyon have substantial accumulations. I am not suggesting that the terraces are of the same age, but there is undoubtedly some overlap in their ages. Second, I found that many terrace sequences were imprecicely dated. Ages are often assigned based on some model (preconcieved notion) of when a deposit formed, such as the assignment of terraces to marine oxygen isotope stages. Third, any such graph would be constructed using a numeric value for soil development when some properties do not easily allow numeric description. For example, the degree of development of carbonate morphology can be given a stage rank (Gile, *et al*, 1966) or may be expressed as a percentage. The first approach does not allow for fine distinctions to be made, and the second is complicated by the variability in rate of carbonate accumulation in different settings. Despite the complications listed above, I am reasonably confident of the age estimates given because I do have some radiometric age control. Until some technique is developed to date deposits that do not contain radiometrically datable materials, soil development will continue to serve as our best means of delimiting the age range of many alluvial deposits.

Summary of Soil Data

The three oldest soils developed in Cochiti Canyon have strong Bt and Bk horizon development, but are easily distinguished based on their landscape position and the color and structure of their Bt horizons. The youngest soil is less extensive than the three older ones, has no Bt horizon, but does have appreciable amounts of calcium carbonate accumulation (table 3). Several lower, discontinuous surfaces are found at the margins of some modern channels, but these are always much lower in the landscape than the major surfaces described above, and soil development consists of A horizon development and thin, weak accumulations of carbonate. Soil development allowed terrace remnants to be confidently distinguished from one another and therefore correlated.

As mentioned above, the upper surface of most soil profiles seem to have been truncated. This truncation is most clearly demonstrated by the presence of Bt horizons near the surface below thin Av horizons (by the absence of an organic A). Erosion on terrace surfaces is evidenced by observations of flow patterns following a large storm in 1995, and 'pedestalled' trees throughout the study area (figure 25). The recent incision that has left trees stranded on islands of fine sediment is likely related to a period of arroyo development initiated in the late nineteenth and early twentieth century in many locations across the Southwest.

Soil development on these terraces seems to follow a pattern of increasing soil clay and carbonate content until some maximum value is reached, whereupon clay and soil carbonate content decrease due to degradation of terrace surfaces. This pattern has been observed in other soil chronosequences (e.g. Birkeland, 1984) and is apparently related to the decreasing infiltration capacity of older (clay and carbonate rich) soils.

Figure 25. Photograph showing a tree that has been left on a pedestal after surrounding regolith was eroded. The recent onset of erosion in fine-grained, unconsolidated sediments seen here is consistent with an episode of arroyo cutting observed in the southwest during the last 125 years.

Construction of Paleo-Long Profiles

Introduction

Paleo-long profiles can be constructed for the terraces discussed above. Since these terraces are old valley floors created in the same way as the modern one, it is reasonable to assume that they originally had profiles similar to the modern floodplain of the Rio Chiquitio. This assumption is supported by the similar grain size and composition observed in the terrace gravels and the modern floodplain (where it has not been covered by pumice due to mining operations or modified by the planting of apple orchards). Any offset that has occurred on the Pajarito Fault Zone can be expected to have deformed the long profiles of the terraces and can therefore be used (with the approximate ages obtained from the study of soil development on the terrace deposits) to directly constrain the timing of faulting and to develop an incision history for the Rio Chiquitio.

Application

Long profiles of the various terraces and the modern floodplain of the Rio Chiquito were constructed by surveying of the floodplain, the contact between terrace gravels and bedrock, and the upper surface of the gravel (figure 26) with a stadia rod, compass, and tape. The modern floodplain of the Rio Chiquito has been modified by the planting of apple orchards on the first reach of the stream flowing on the hanging wall block, and by an influx of pumice created by mining. Despite this disturbance, a smooth profile was constructed by surveying of undisturbed floodplain sections and then joining these segments with the aid of topographic maps (figure 27).

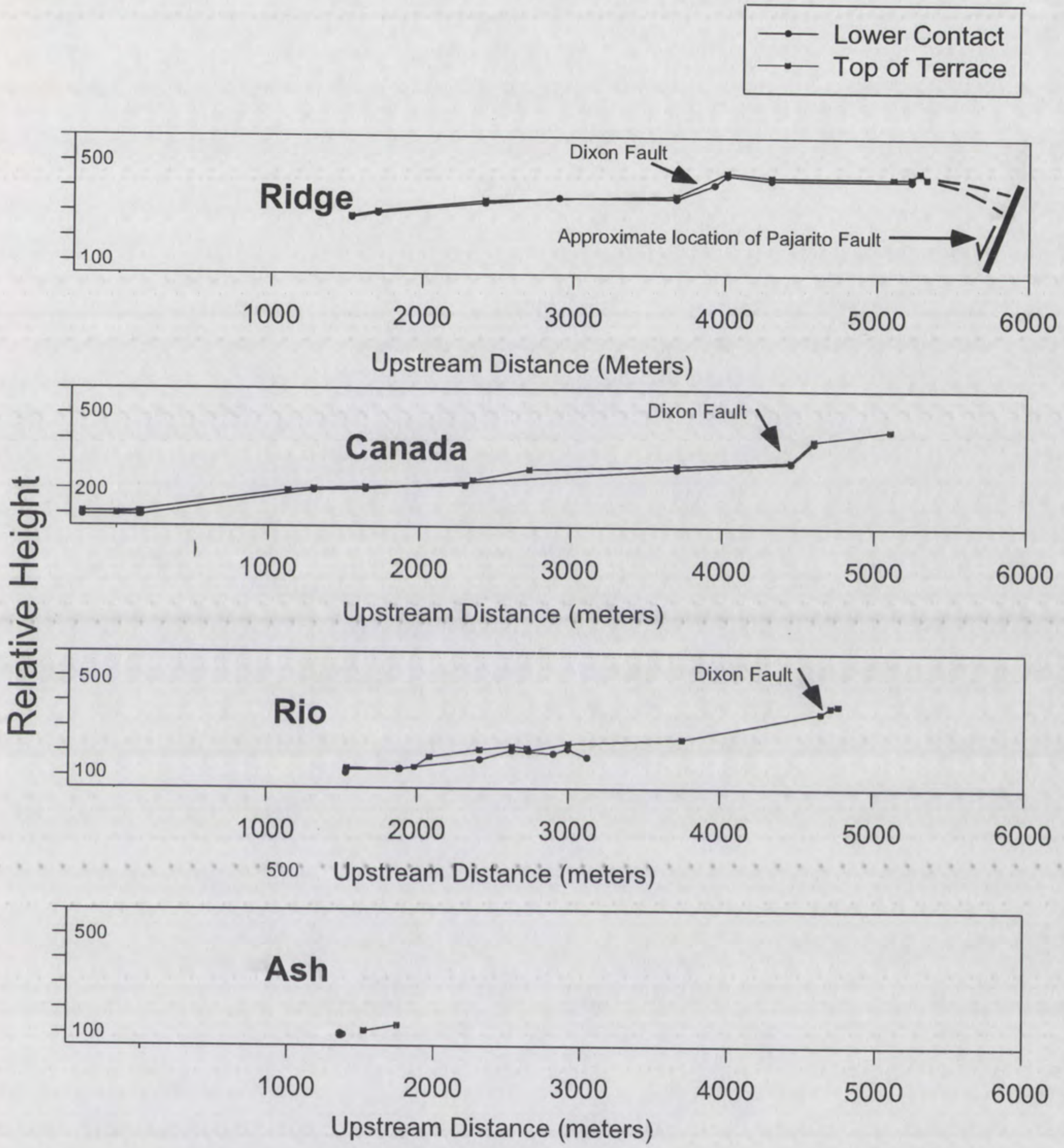


Figure 26.
Long profiles of terraces constructed from surveys of terrace remnants. Upper line of each profile indicates the top of terrace gravels (the terrace tread) while lower line shows the profile of the underlying strath. Upstream distance and relative height are measured from an arbitrary zero point at the intersection of the Rio Chiquito and the Rancho Canada property line. All elevation points were projected to a plane running down the axis of the modern valley.

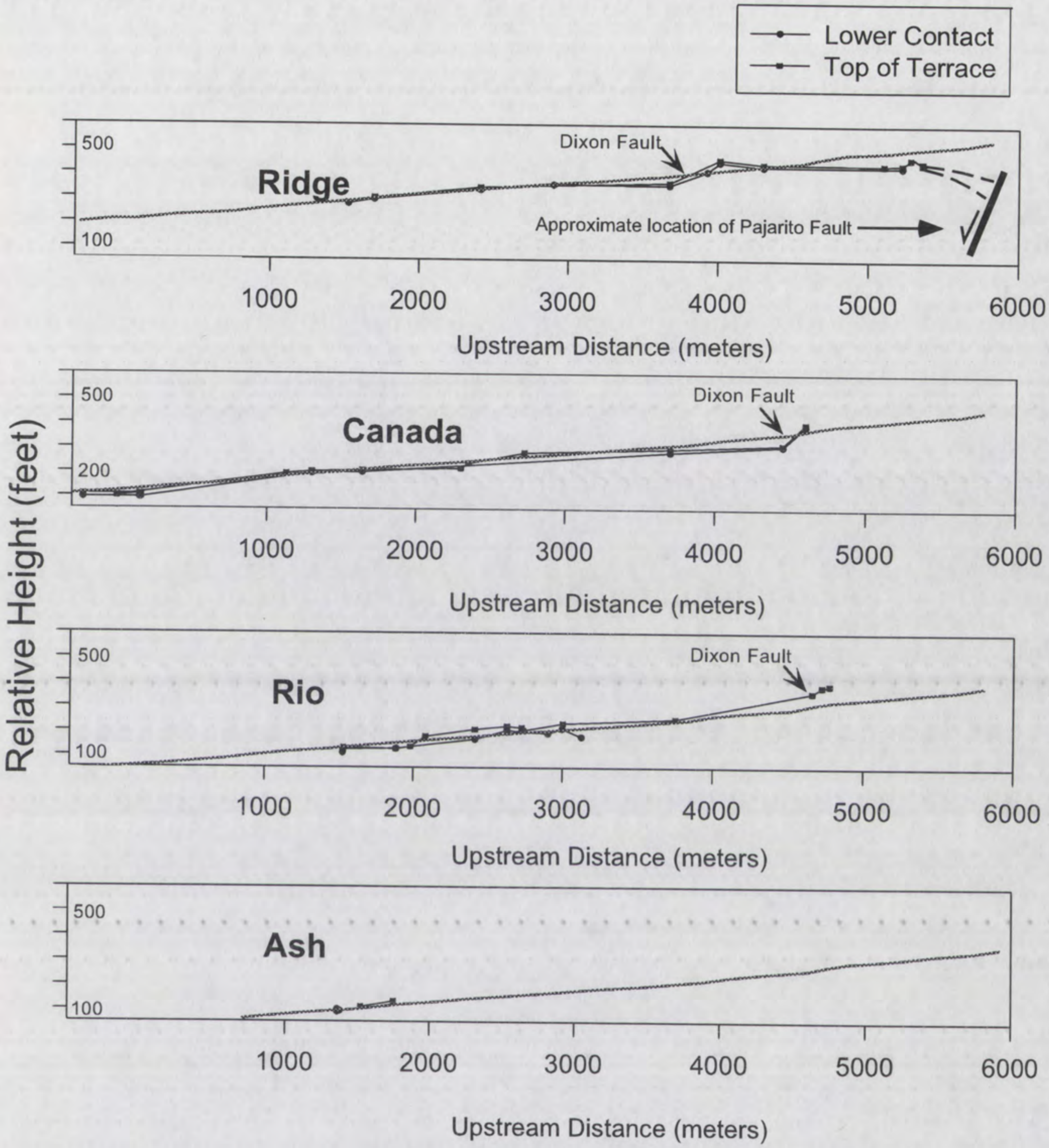


Figure 27.
Long profiles of terraces constructed as in figure 26, with the modern stream profile (stippled line) superimposed for reference. Modern profile was shifted vertically, but not horizontally, so that a best fit was obtained for the surveyed points on the terrace profile. This procedure allows discontinuities in the terrace profile (in this case fault-offsets) to be more easily seen., and also illustrates that offset of the profiles is confined to areas adjacent to fault splays.

Deformation of Terraces

Comparison of the modern profile to the profiles created for the terraces (figure 26) reveals that the Ridge terrace has clearly been backtilted towards the main splay of the Pajarito Fault. Thirty-three meters of offset seen along the trace of the Dixon fault (figure 2) is also clearly seen on this profile. Backtilting of the terrace on the hanging wall of the Dixon fault is also apparent. The Cañada terrace is offset 25 meters by the Dixon fault, and the profile is backtilted to a lesser degree than the Ridge terrace. Unfortunately no remnants of the Cañada terrace are exposed near the main splay of the fault zone. The Cañada terrace is graded to the surface of a basalt flow mantled with gravel near the mouth of Cochiti Canyon (figure 13). Offset of this basalt flow and the overlying gravel is approximately 30 meters, and offset on this fault creates a nickpoint in the profile of the Rio Chiquito (figure 8). The Rio terrace is offset by the Dixon fault by only 5 meters. The Dixon fault seems to join the main splay of the fault zone just to the north of Eagle Canyon, so offset can be expected to decrease to the north (as does offset along the main fault). In fact, upper Bandelier Tuff is offset only 21 meters by the Dixon fault in the north wall of Cochiti Canyon. Since this offset is measured north of remnants of the Rio terrace, at least 16 meters of post 1.22 Ma/pre 125-175 Ka offset is indicated for the Dixon fault north of the Rio Chiquito, and at least 5 meters of offset in the last 125-175 Ky.

The Ridge terrace is the only surface that is demonstrably deformed by movement on the main splay of the Pajarito fault zone in Cochiti Canyon (figure 27). If it is assumed that the profile of the ridge terrace was originally identical to that of the modern valley floor and that backtilting would increase toward the fault in an exponential manner, then 35 to 50 meters of offset on the main splay of the Pajarito Fault Zone since Ridge terrace formation is indicated. Offset of the upper Bandelier Tuff in the same area is 110 m (figure 28, table 4). The main splay of the Pajarito fault zone is exposed in a cut-bank exposure in Bland Canyon. The fault juxtaposes Peralta Tuff against terrace gravels that show soil development intermediate between the Ash and Rio Terraces (table 3). An age of approximately 85 ka is estimated for this soil, based on the distribution of points in figure 24.

Based on the offset listed above and the previous estimates of soil age, rates of offset were calculated for the various splays of the fault zone (table 4). Offset rate on the Main splay and the Dixon fault has remained fairly constant through time (figure 28). There is not enough data for the Basalt splay to determine any trend in offset rate. It is also not possible to definitively determine if the Dixon fault and Basalt splay were active prior to creation of the Canada terrace. However, the fact that they do not offset the Bandelier Tuff to the north of Cochiti Canyon indicates that they are localized structures, and may indicate that the total offset across the fault zone in this area has increased in the last few hundred ky as these faults became active.

Stream Incision

The relatively undeformed portions of the terrace profiles (between 1 and 3 kilometers from the zero point used in this survey) allow the calculation of incision rates for the Rio Chiquito from 1.61 Ma to the present (figure 29, table 4). Rates were calculated by subtracting the elevation of several points on the terrace profiles from the elevation of equivalent points on the profile of the Rio Chiquito. Calculated rates range from 0.1 to 0.30 mm/yr. Rates of incision were fairly constant from 1.22 Ma until 60 ka, and then increased

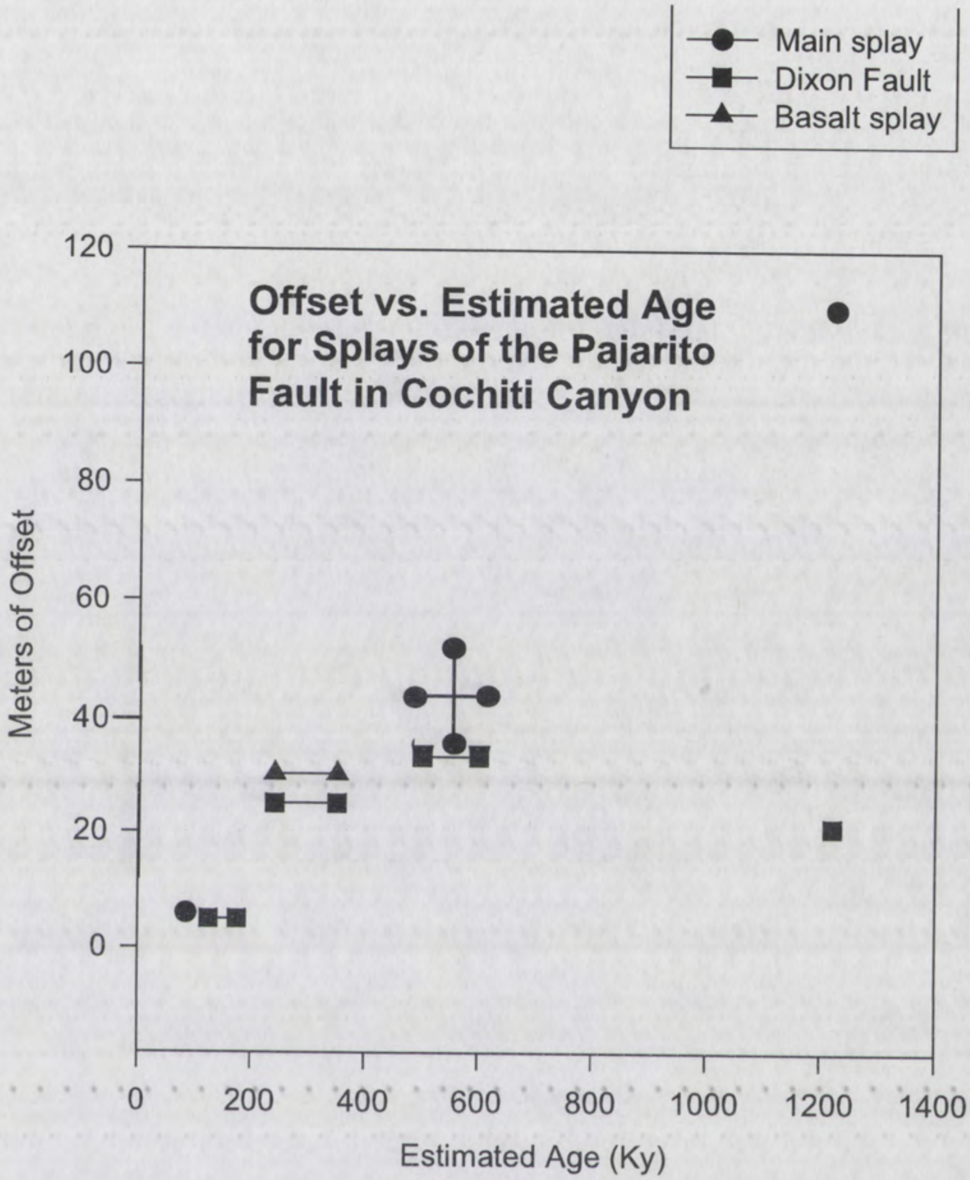


Figure 28. Plot of apparent offset of terraces as measured from long profiles (figure 27) versus estimated ages of terraces for various splays of the Pajarito Fault Zone. Horizontal lines indicate uncertainties in age estimates for terraces, vertical lines indicate uncertainties in amount of offset. Please see text for a discussion of age estimates.

OFFSET RATES

Fault Splay	Estimated age of Offset Surface (ky)	Amount of Offset (meters)	Offset Rate (mm/yr)
Main Splay			
	85	6	0.07
	500	35	0.07
	500	50	0.10
	600	35	0.06
	600	50	0.08
	1220	110	0.09
Dixon Fault			
	500	33	0.07
	600	33	0.06
	350	25	0.07
	240	25	0.10
	175	5	0.03
	125	5	0.05
Basalt Splay			
	240	30	0.13
	350	30	0.09
Average of calculated offset rates =			0.08

INCISION RATES

Terrace/ Rock Unit	Estimated age of Surface (ky)	Incision (meters)	Incision rate (mm/yr)
Bandelier (lower)	400*	110	0.3
Bandelier (upper)	1220	125	0.1
Ridge	600	87	0.15
	500	87	0.17
Canada	350	30	0.09
	240	30	0.13
Rio	175	18	0.1
	125	18	0.14
Ash	60	15	0.25
Average of Calculated incision rates=			0.16
Average of Post upper-Bandelier only=			0.13

Table 4. Offset and incision rates calculated from heights of terraces, estimated ages, and measured offset of terrace surfaces

*Amount of time between upper and lower Bandelier eruptions.

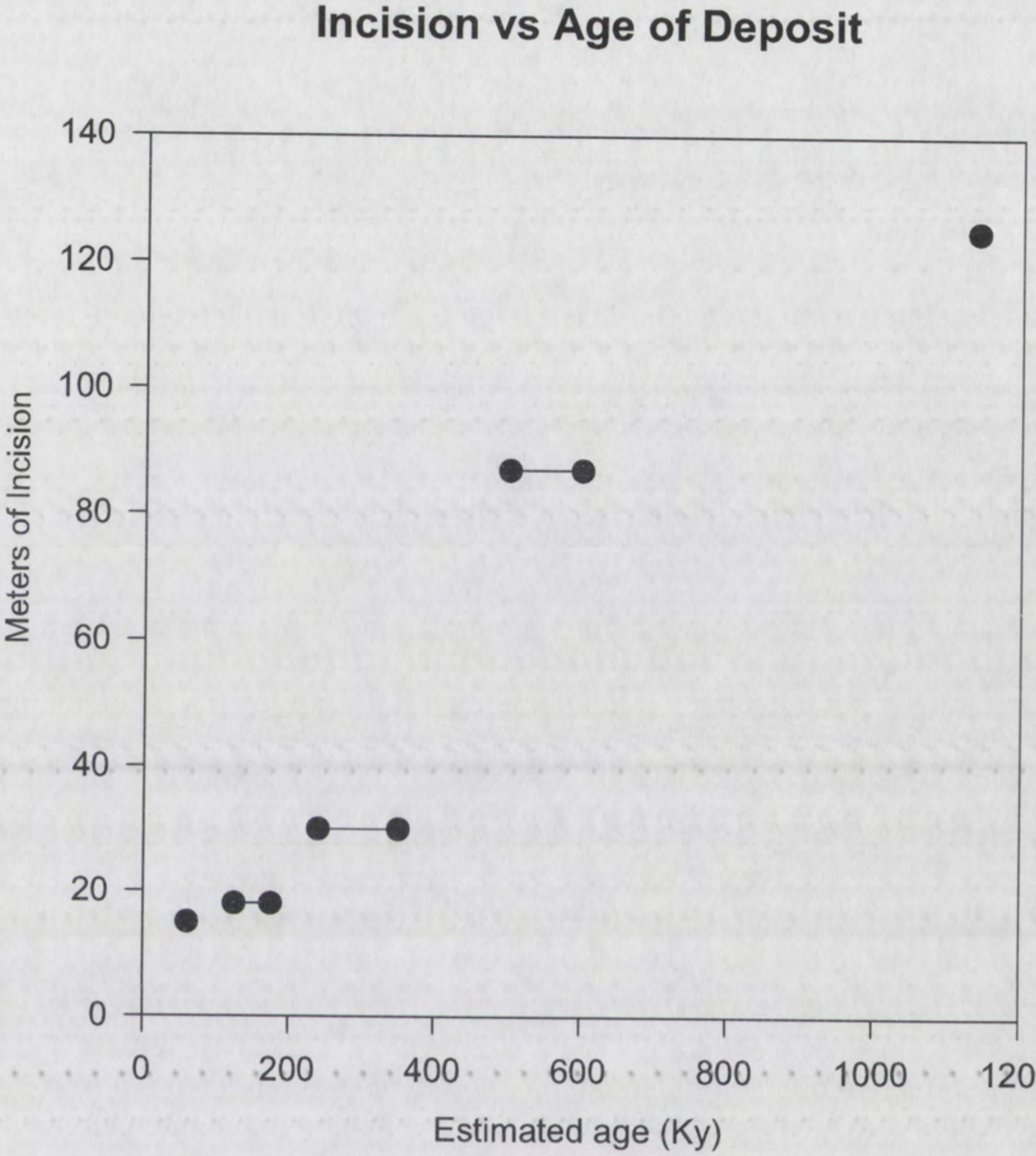


Figure 29. Plot of estimated age of terrraces and age of the upper Bandelier Tuff versus amount of incision (present height above grade). Horizontal lines indicate uncertainties in age estimates of terraces. Incision data are averages of 5 measurements of height above grade on relatively undeformed portions of long profiles shown in figure 27. Please see text for a discussion of the ages used in this plot.

from 60 ka to the present. This observation either reflects a period of relatively rapid incision through the Ash terrace fill, or the incorporation of periods of aggradation or equilibrium into the average rates for older terraces. The latter effect has been documented for geomorphic and tectonic rates in general by Gardner *et al.* (1987).

It is also possible to calculate a minimum amount of incision in the lower reaches of the Rio Chiquito between the time of the two Bandelier Tuff eruptions (figure 13,14). Measuring from the top of the present outcrop of the lower/upper Bandelier Tuff contact to the bottom of tributaries of the Rio Chiquito flowing on consolidated, lower Bandelier-derived sediments yields a minimum amount of incision of 110 meters in the 370 ky between the two eruptions. These figures give a rate of incision of 0.30 mm/yr. This rate is three times the average incision rate from 1.22 Ma to the present, and indicates that much more stream power was available for incision during this period, or that the lower Bandelier Tuff is very easily eroded relative to the upper Bandelier Tuff.

Conclusions

The above data allows for the construction of a history of motion on splays of the Pajarito Fault Zone near Cochiti Canyon and fluvial incision since eruption of the lower Bandelier Tuff. Important observations include:

1) Between one-third and one-half of the total post-Bandelier displacement on the main splay of the fault zone has taken place since creation of the Ridge terrace. Since the Ridge terrace formed roughly halfway between the upper Bandelier eruption and the present, it would seem that displacement on the main splay of the fault zone has been evenly distributed through time.

2) Since many reaches of the terrace profiles appear to be undeformed, it seems that most of the deformation has occurred within 1 kilometer of specific faults (figure 27). This observation contradicts the common practice of depicting normal faulting as a plunger-like displacement of a hanging-wall block relative to the footwall.

3) The Dixon fault displaces successively younger terraces by successively smaller amounts.

4) The rate of displacement on the Dixon fault is relatively constant except for displacement calculated for the Bandelier Tuff on the north side of Cochiti Canyon, indicating that displacement decreases rapidly to the north.

5) Displacement rates are between 0.01 and 0.13 mm/yr.

6) Displacement of the Cañada terrace is roughly equal across the Dixon Fault and Basalt splay (~33 m). If a line is drawn between the two data points showing offset on the main splay of the Ridge terrace and the ~85 ka terrace in Bland Canyon, it is seen that the predicted displacement of a Cañada-age terrace would be approximately 33 meters. In other words, displacement across the fault *zone* during the last 240-350 ka has been three times the average rate for the main splay alone (if the Dixon and Basalt splays truly became active since creation of the Cañada terrace).

7) Incision by the Rio Chiquito proceeded at a relatively constant rate from the time of deposition of the upper Bandelier Tuff until formation of the Ash terrace and was perhaps then slightly more rapid from the time of Ash terrace formation to the present.

8) Incision in the lower reaches of Cochiti Canyon was most rapid during the period between eruption of the lower and upper Bandelier Tuff than at any time since eruption of the upper tuff.

Comparisson to Previous Terrace Studies

To date, two other terrace sequences have been examined in detail in the Jemez Mountains. These are the terraces near the confluence of the Rio Grande and Rio Chama (Dethier and Harrington, 1987; Dethier *et al.*, 1988; Dethier and McCoy,1993; Dethier and Reneau, 1995; Gonzalez, 1993; Gonzalez and Dethier, 1991), and a terrace sequence near the confluence of the Jemez River and the Rio Guadalupe in the southwestern Jemez Mountains (Rogers, 1996). The former sequence has been dated and correlated using numerous amino-acid racimization ratios calibrated with radiometric ages of tephras associated with terrace deposits and radiocarbon ages of young terraces. Rogers obtained ages for his terraces by archaeologic and radiometric means and correlated undated terraces with those of the Rio Grande/Rio Chama area based on amino acid ratio comparisons and height above grade. Comparison of the age and height of these terrace sequences to those in Cochiti Canyon, reveals that terraces of the same estimated age are consistently found at lower elevations in Cochiti Canyon than elsewhere in the Jemez Mountains (table 5, figure 30). This observation, if based on acurate age estimates, indicates that a regional baselevel is not the only factor controlling the height of terraces at a given time.

Incision rates were calculated by Dethier *et al.* (1988) and Dethier and Reneau (1995) for the northwestern Española Basin using ages obtained from both varnish-cation ratios of surface-clast rinds and amino acid racemization ratios in shell material calibrated with uranium/thorium ages of soil carbonate. The rates of stream incision they calculate are consistently higher than those calculated for the Rio Chiquito (figure 30). For example, surfaces

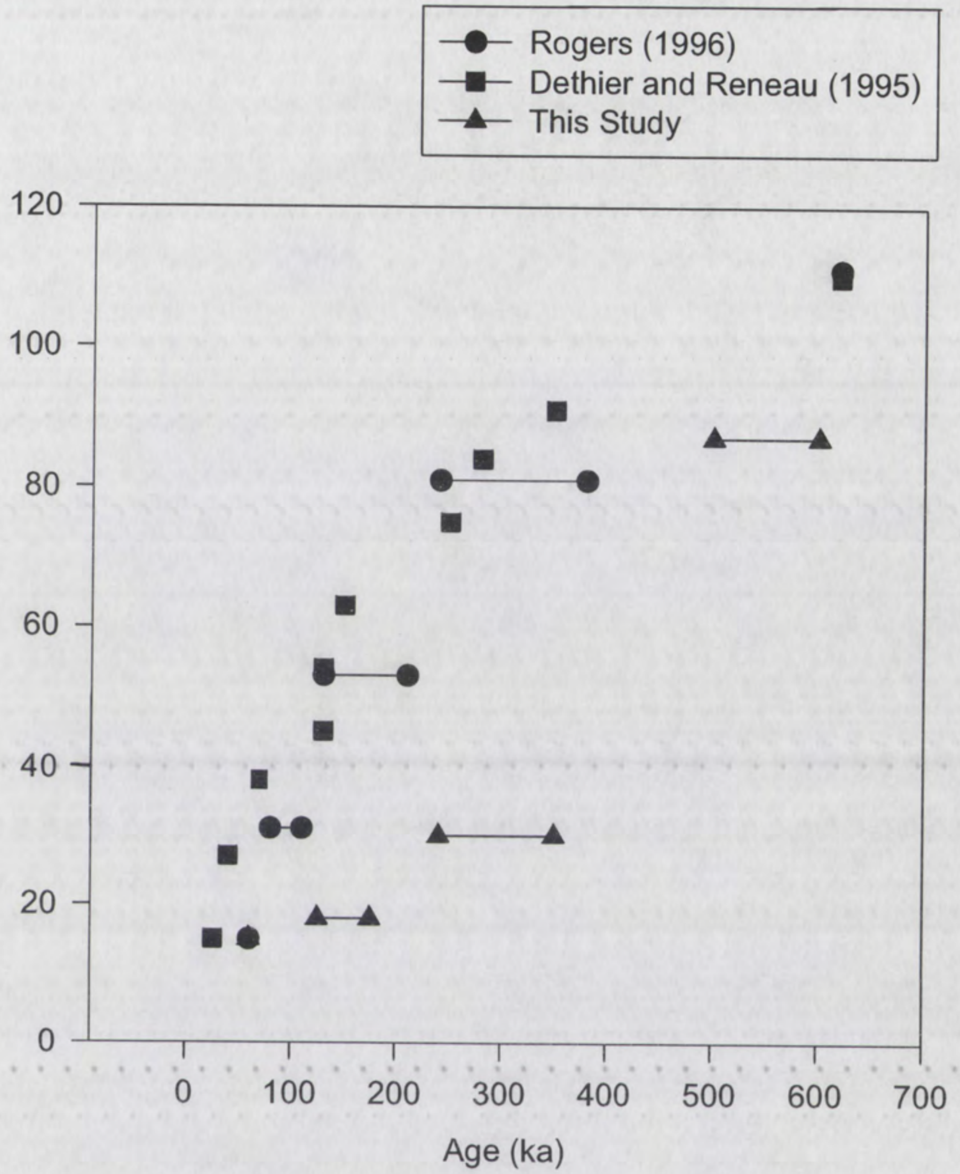


Figure 30. Comparisson of height above grade vs.age relation for terraces in the Jemez Mountains. Age and height data are taken from Rogers (1996), Dethier and Reneau (1995) and this study. Note that terraces in Cochiti Canyon are consistently lower for a given age. Horizontal lines indicate uncertainties in age estimates. Uncertainties wre not reported by Dethier and Reneau (1995).

they have correlated to the Cañada terrace of this study, are 90-120 meters above grade in the northwestern Española Basin, whereas this terrace is only 30 meters above the Rio Chiquito. This difference may be attributed to the presence of basalt flows near the mouth of the Rio Chiquito. If this inter-basin correlation is correct, it suggests that the timing of terrace formation is climate-driven (since terraces formed in all canyons regardless of what geologic unit they were incising through or what tectonic regime they were subjected to and that regional correlation of terraces based on height above present grade is an uncertain proposition at best.

I do not believe that the age control obtained for these terrace sequences is precise enough to allow comparison of the timing of terrace formation on a regional scale. Although Rogers (1996) and those studying terraces in the Espanola Basin (e.g. Gonzalez and Dethier, 1991) have suggested that terraces form in these locations during the transition from a cooler/wetter climate to a warmer/drier climate, the uncertainties reported by Rogers (1996) and presumably applicable to the ages reported by Dethier and Reneau (1995), are larger than the duration of a climatic transition or, indeed, a glacial/interglacial *cycle* (table 5).

It is my further contention that a correlation of the terraces in Cochiti Canyon to those found elsewhere in the Jemez Mountains based on height above grade would lead to erroneous conclusions. Such a correlation would rely on the principle that regional base level was controlling the height of all terraces in a region and that all streams in that region were graded at the time of regional stream incision (floodplain abandonment). Such an assumption is not warranted in the case of Cochiti Canyon because the height of terraces here is controlled by not only regional baselevel (the lowest level that they might be found) and by the hydrologic factors unique to this stream system, but by the local baselevel control exerted by the basalts found at the mouth of the canyon (figure 8). Correlation of the terraces of Cochiti Canyon to others in the Jemez Mountains would further require the correlation of fill terraces (most common in other parts of the Jemez) to strath terraces (most common in Cochiti Canyon). The fill underlying a fill terrace commonly has a variable height above grade at different points along a stream (e.g. Bull, 1991), and it is therefore difficult to know at which point along a stream to measure the characteristic "height above grade".

Table 5. Ages and heights above grade for terraces in the Jemez Mountains. See figure 30 for a graphical representation of this data.

Rogers, 1996

Terrace	Height above grade (meters)	Type of age control	Inferred age (ka)
Qt1	111	radiometric*	620
Qt2	81	amino acid**	310+/-70
Qt3	53	amino acid**	170+/-40
Qt4	31	amino acid**	95+/-15
Qt5	15	Carbon 14***	~60

Dethier and Reneau, 1995

Terrace	Height above grade (meters)	Type of age control	Inferred age (ka) (no range given)
	110	radiometric*	620
Names	91	amino acid**	350
not	84	amino acid**	280
assigned	75	amino acid**	250
	63	amino acid**	150
	54	amino acid**	130
	45	amino acid**	130
	38	amino acid**	70
	33	amino acid**	<100
	27	amino acid**	40
	15	Carbon 14***	>26

This study

Terrace	Height above grade (meters)	Type of age control	Inferred age (ka)
Ridge	87	estimate****	500-600
Canada	30	VCR*****	240-350
Rio	18	estimate****	125-175
Ash	15	Carbon 14***	~60

*Age of the Lava Creek B ash Dethier *et al*, *(* 1990).

**Based on calibrated amino acid racemization curve of Dethier and McCoy (1993 and unpublished data). Roger's ages are based on correlation to this curve using both amino acid ratios and height above grade.

***Age of El Cajete Pumice eruption and Bonco Bonito lave flow based on age of trees killed by eruption of pumice (Reneau, *et al*, 1994).

****Estimates based on degree of soil development. See discusssion of age estimates in the text.

*****Calibrated Varnish Cation Ratio age from Dethier and Harrington (1987). See discussion of age assignments in text.

SUMMARY OF INFORMATION OBTAINED DURING EXAMINATION OF THE TIMING OF MOTION ON FAULTS OF THE PAJARITO FAULT ZONE

Introduction

Up to this point I have attempted to include not only objective observations and the logical conclusions that could be drawn from them, but also the assumptions, opinions, and philosophical framework that guided my investigations. This style of reporting my work is intended to give a rounded picture of the way in which conclusions have been drawn and to suggest alternative interpretations. The summary that follows continues in this same vein while trying to combine all of the above information into a cohesive interpretation of the data I collected during this study.

Discussion and Conclusions

The methods used to investigate the timing of faulting within the Pajarito Fault zone lead to many interesting conclusions about the Quaternary geologic history of Cochiti Canyon. However, it was the process of applying these methods itself that lead to the greatest gains in my own knowledge and skills. By trying to elucidate the timing of faulting in the Pajarito Fault Zone through morphometric analysis, I was forced to completely examine the methodology and the assumptions that are this technique's foundation. Foremost among these assumptions is the idea that some connection can be made between a measured morphometric parameter and some landscape-shaping *process*. For example, if the hypsometric integral reflects the relative timing of faulting in a basin, then changes in baselevel of the stream must initiate changes in the distribution of slope on the hillsides of that same basin. How exactly does this occur? First, faulting offsets a stream bed and the resultant nickpoint propagates upstream--but how far? There is no clear consensus on this point. Leopold and Bull (1979) suggest that base level changes affect only those parts of streams immediately adjacent to the baselevel perturbation for basins tens of kilometers in size and over time scales of thousands of years. Leopold (1978) also reports that baselevel change initiated by damming of small arroyos was transmitted upstream a short distance in a few years and then no further in subsequent years. Bull (1991) apparently holds to these same ideas a decade later. Merrits *et al* (1995), on the other hand, implicitly call on nickpoints migrating tens of miles to headwater reaches in response to changes in sea-level (when Merrits has worked ***with*** Bull in the area where this study was conducted!), and also suggest that little is known about bedrock erosion at present. Experimental studies indicate that baselevel change is sometimes transmitted to headwaters reaches, but one must keep in mind that the processes operating on the scale of a flume and those operating in a real drainage basin are not the same. Computer modeling studies suggest that the response of headwaters are dependent on the rate of baselevel change (Bonneau and Snow, 1992), but such models are dependent on the conceptual framework that go into their creation (i.e. computer models are only as good as the field geology that provides their "boundary conditions"). This short discussion is not intended to belittle the state of geomorphology, but to point out that some conclusions must be drawn from these conflicting studies in order to apply the concept of hypsometry. I assumed that nickpoints generated by faulting would migrate at least 1 kilometer upstream. Once this baselevel change has been transmitted to the base of a hillslope, some change in hillslope processes must be initiated. No studies I have seen directly address the question of how this occurs. It is possible that rills could be initiated in the footslope area of a valley and that these could propagate upslope, but the same question of how far nickpoints can travel upslope would apply. It is also possible that no change would occur in the hillslopes if they were isolated from the stream by a floodplain or flat footslope area. I assumed that, over the course of several thousand years, baselevel changes would be transmitted far enough up the hillslope (10-15% of the total length ?) to be reflected in the hypsometry of the basin. Only by constructing such a model was I able to apply hypsometric analysis in the way that I did. Had I drawn some other conclusion from the geologic literature, I would have either not attempted this technique or would have interpreted my results in other ways. Each of the methods applied have a similar thought process behind them. A thorough discussion of the theory behind each technique will not be included here because I believe much of this can be drawn from the section on morphometry. However, several underlying principles of morphometric analysis revealed themselves during the application of the techniques:

1) Different techniques are more or less applicable to different time and space scales.

Example: While drainage density can be used to infer changes in the long-term controls on basin evolution (e.g. climate, rate of uplift, changes in lithologic properties) or reveal differences in basins with different histories, nickpoints in stream profiles reflect either changes in critical power, changes in stream power, or faulting in the recent past (figure 8, Table 1).

2) A technique can be applied in various ways depending on the time or space scale of interest.

Example: Hypsometric analysis of entire basins is best suited to revealing long-term changes in basin evolution, while the hypsometry of basin segments immediately adjacent to faults seems to reveal information about the relative timing or magnitude of faulting in that basin on time scales of several thousands of years.

3) Unless process rates are constrained by experimental, observational, or geologic evidence, no unambiguous conclusions can be drawn from morphometric parameters and no rates can be calculated.

Example: Even though apparently fault-generated nickpoints are seen in several streams studied here (figure 8), no rate of offset can be calculated because the rate at which the nickpoint would be degraded, and the manner in which it would retreat (by parallel retreat, nickpoint replacement, or 'rollback') are not known.

4) Lithology is the overriding control on the value or character of morphometric parameters.

For numerous examples, see the Morphometry section.

5) The value of a morphometric parameter is dependent not only on geologic factors but on the scale of the map used to measure it.

Example: Drainage density calculated from 1:100,000 scale maps will always be lower than those calculated from 1:24,000 scale maps since fewer streams can be 'seen' at this scale (and drainage density is .

6) Morphometry is best utilized as a prelude to field investigations or to augment conclusions drawn from field work. When it is not clear what part of a basin should be concentrated on, morphometry can suggest avenues of investigation; and when conclusions drawn from field investigations are unclear, morphometric analysis can lend corroborating or contradictory evidence.

In the case of this study, all of the above principles were applied profitably. Each technique I applied led me to apply another that was more narrowly focused on determining the timing of faulting. General morphometric analysis (drainage area, drainage density, etc.) revealed that Cochiti Canyon was of a unique size within the Jemez Mountains (table 1) and that this was one possible reason for the existence of terraces here. This observation both encouraged me to continue my investigation in general and directed my efforts toward smaller-scale analysis. Hypsometric analysis of basin segments immediately adjacent to fault splays offered the first evidence to suggest that faulting had occurred relatively recently in Cochiti Canyon (figure 7). Hypsometric curves for canyons in the study area were markedly more convex than those in other parts of the Jemez Mountains, even when lithology appeared to be the same. Hypsometric curves also became more convex to the north *within* the study area, and this is interpreted as showing that faulting began in the vicinity of Bland Canyon and has migrated northward. The fact that the measured offset of geomorphic surfaces decreases to the north (figure 28) lends support to this conclusion. Hypsometry is inferred to reveal the influence of faulting on time scales of many thousands of years since changes in baselevel must be transmitted through the basin system before they are reflected in the hypsometric integral. Nickpoints seen in the long profiles of streams, on the other hand, reflect recent faulting events. Cochiti, Eagle, and Bland Canyons have nickpoints near fault splays that cannot be attributed solely to lithologic variations (figure 9 and 10). Although the rate of nickpoint obliteration is not known (see above) this observation still suggests that faulting has occurred in the geologically recent past on the Main splay (in Bland canyon) Dixon fault (in Eagle Canyon) and Basalt Splay (in Cochiti Canyon). Sinuosity is thought to respond closely to tectonics because it is sensitively related to slope. Sinuosity changes in the study area are opposite those seen for other tectonically active areas (*Schumm et al*, 1987), where sinuosity increases on uplifted areas. For the streams examined here, sinuosity is consistently lower on the upthrown block of the main splay of the fault zone (figure 11). This observation is related to changes in lithology for some canyons, but may also reflect the influence of fault-generated nickpoints. Since segments of a stream immediately upstream of a nickpoint are actively downcutting through their beds, their slope should be at a maximum--which leads to a low sinuosity. The true value of the morphometric analysis performed here was as a compliment to the field investigation.

The offset of stream terraces measured in the field area is the most solid evidence with which to constrain the timing of faulting on various fault splays (figure 28). Offset could be measured directly where the Bandelier Tuff and the Ridge, Cañada, and Rio terraces are offset by the Dixon fault; and where the Basalt splay offsets the Cañada terrace. Offset could be inferred for the main splay by estimating the amount of backtilting of the Ridge terrace (figure 27). Although the conclusions drawn from this study are simple ones (amount and estimated timing of offset), such data has not existed up to this point. This type of data is important, considering that one of the nation's leading nuclear research facilities is located within 15 miles of the field area. Faulting on the Pajarito Fault Zone has been distributed evenly in time since the eruption of the upper Bandelier Tuff and as such can be expected to continue in the future. Faulting since creation of the Cañada terrace has been evenly distributed on three splays of the fault zone. Although little or no data exists with which to constrain the offset on two of these faults during other periods, this observation may indicate that the mechanics of offset within the fault zone has changed in the last few hundred ky .

QUATERNARY GEOLOGIC HISTORY OF COCHITI CANYON

A geologic history is the sum total of the knowledge of a particular area. As such, I have chosen to conclude with a discussion of the Quaternary geologic history of Cochiti Canyon as revealed by my investigation of the Pajarito Fault Zone. In the early Quaternary the area near Cochiti Canyon seems to have been covered entirely with a volcanically-derived alluvial apron of pre-Bandelier gravels (figure 13, 14). Relief on this landscape, as revealed by the present exposures of the lower Bandelier/pre-Bandelier gravel contact, was not as dramatic as the relief present today. The Pajarito Fault Zone was apparently active prior to eruption of the lower Bandelier Tuff, as offset of pre-Bandelier deposits is greater than offset of the Bandelier Tuff itself (Gardner and House, 1987). Eruptions that lead to deposition of the San Diego Canyon Tuffs (G.A. Smith, written communication, 1996) in the southwestern Jemez Mountains also lead to the deposition of 50+ m of pumice-rich fluvial deposits prior to eruption of the Bandelier Tuff. These deposits were buried and preserved by a basalt flow of the Cerros del Rio field near the present mouth of Cochiti Canyon. Eruption of the lower Bandelier Tuff filled in much of the pre-Bandelier topography that was developed on the alluvial deposits. Streams developed on the surface of the lower Bandelier Tuff and carried lithic and pumice-rich sands and gravels to the mouth of an ancestral Cochiti Canyon and periodically deposited them in interfingering tongues with ancestral Rio-Grande deposits (figure 13, 14). Some thin gravel deposits (presumably derived from volcanic highlands to the west) were preserved above the lower Bandelier Tuff by eruption of the upper Bandelier, but no substantial constructional landforms are exposed between the two tuff deposits. This is not surprising since most stream energy was probably expended in the creation of stream networks and incision through the tuff. However, relatively rapid erosion of the lower Bandelier Tuff lead to the deposition of pumice and lithic-rich sand near the present mouth of Cochiti Canyon (figure 13). These deposits seem to be interbedded with quartzite-rich Rio Grande axial deposits, and seem to indicate episodic pulses of pumice-rich sediment entering the Rio Grande during this time. These deposits also show that the Rio Grande flowed to the west of its present position sometime after eruption of the lower Bandelier Tuff. Rapid incision rates calculated in the lower reaches of Cochiti Canyon for the time between the eruption of the two members of the Bandelier Tuff (see discussion of paleo-long profiles) may be further evidence that the Rio Grande flowed in this area during this time. Eruption of the upper Bandelier Tuff effectively erased pre-existing topography. Stream networks developed on the upper surface of the tuff and began to incise through it in an effort to once again attain their baselevel of erosion. Curiously, the present bed of many streams found on the hanging wall of the fault system are at the same stratigraphic level that they were at just before the eruption of the lower Bandelier Tuff. It would seem that down-dropping on the hanging wall and lowering of over-all baselevel have proceeded at the same rate throughout this time. However, long-profile analysis indicates that deformation has been restricted to zones immediately adjacent to fault splays. It would therefore seem that regional baselevel for these streams has not changed more than a few tens of meters in the last million and a half years!

Eruption of the upper Bandelier Tuff also displaced the Rio Grande to the east by an unknown amount. When the Rio Grande incised through the upper Bandelier Tuff, it did so ~2 km to the east of its former location. While incising through the upper Bandelier Tuff, the Rio Grande exhumed and beveled the basalt flow near the mouth of Cochiti Canyon, leaving a quartzite-rich gravel sheet and eventually incised through the basalt to establish its present course.

As the Rio Grande incised through the upper Bandelier Tuff, it continually increased the power of the Rio Chiquito by lowering the local baselevel. The Ridge terrace was formed before the Rio Grande had beveled the basalt flow at the mouth of Cochiti Canyon (figure 12). This terrace forms the present divide between Cochiti and Bland Canyons and reconnaissance indicates that it is present in other nearby areas. This evidence suggests that the Ridge terrace may have been part of an extensive pediment, or (if it was originally confined to valley-bottoms) that the positions of the canyons in this area have shifted laterally more than one (present) valley width during incision.

The Ridge terrace is <150 meters lower than the upper surface of the upper Bandelier Tuff, but is >700,000 years younger. Although the Rio Grande (and an ancestral Rio Chiquito?) apparently incised through the entire thickness of the lower Bandelier Tuff in 100-200 ky, the river has not incised as rapidly since eruption of the upper Bandelier Tuff. It seems unlikely that this decrease in incision rate is due solely to the welding of the upper Bandelier Tuff. If it is correct that the Rio Grande flowed to the west of its present location, it is possible that the baselevel control exerted by the basalt flows at the mouth of Cochiti Canyon was not felt prior to eruption of the upper Bandelier Tuff. The Cañada terrace is graded to a surface beveled(?) on this same basalt flow by an ancestral Rio Grande (figure 12). Dethier and Harrington (1987) dated this surface with the varnish-cation method (calibrated with uranium/thorium ages of soil-carbonate) at 270-350 ka.

At some time subsequent to formation of the Cañada terrace, north-facing hillslopes on the southwestern side of Cochiti Canyon developed colluvial wedges graded to this terrace (figure 12). The lower parts of these colluvial wedges are gravel rich, but the upper parts contain 10-99% El Cajete Pumice, indicating that colluviation was episodic and protracted. After 25 meters of incision into the Cañada terrace, another period of baselevel stability and lateral stream incision formed the Rio terrace. The Rio Chiquito and its tributaries then incised to near their present level and began to backfill. During this period of aggradation, the El Cajete pumice was erupted and ash and pumice from this eruption was incorporated into valley-filling alluvium in some small tributaries of the Rio Chiquito. Subsequent incision of this fill lead to the present stream configuration.

Approximately 10 ka, the forerunners of the Pueblo people may have (i.e. I have no evidence) hunted mammoths, ground sloths, and camels in Douglas Fir forests in Cochiti Canyon. Five or six hundred years ago, Chacoan people migrated *en mass* to a site in Cochiti Canyon (on the Rio and Ash terraces) that would become the largest pre-Columbian settlement in the Southwest. The Cochiti people, as those who lived here would later be known, encountered Spanish explorers in the 1500's, were eventually subjugated by them, revolted during a period of severe drought in the early 1680's and moved to defensive positions at the top of Horn mesa between the Dixon fault and the main splay of the Pajarito fault zone. Here they stockpiled volcanic clasts transported by the Rio Chiquito and Bland Creek to throw down on any Spanish people who might return. The Spanish returned with guns and swords and the Cochiti were re-subjugated and moved out of Cochiti Canyon to their present site. Erosion of hillslopes initiated during this century lead to the formation of gullies and pedestalled trees and reworked many of the artifacts and buildings left behind. The walls of former dwellings collapsed, sheetwash moved large volumes of El Cajete Pumice during storms, and the Rio Chiquito rolled (damn-near perennially) on.

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Appendix 2

Definition of morphometric parameters used in this report.

Morphometric Parameter	Definition
Basin Area	A
Basin Perimeter	P
Lenght of Long Axis of Basin	Bl
Number of streams of nth Order	Nn
Length of Streams of nth Order	Ln
Length of All Streams in a Basin	Lt
Drainage Density	D= Lt/A
Basin Shape	B=A/Bl
Bifurcation Ratio	Nn/N(n+1)

Appendix 3
Field Descriptions of Soil Profiles in Cochiti Canyon
(see figure 13 for location of soil pits)

Soil Pit 1: Ridge Soil
Ridge on upthrown side of Dixon fault
Described on 8/31/95
Vegetation: Pinon/Juniper forest with sparce grass, cactus, and desert shrubs.
Mood: Real good, Lluvia up here with me
Percent pavement cover: 90

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-10	10YR 5/3	SL	v1,f,br	a/s	1,f,sbk	V.WEAK FIZZ
Bt1 10-28	10YR 4/2	SiCL	3,d,co	c/s	1,f, pl/sbk	NO FIZZ
Bt2 28-42	7.5YR 3/4	SCL	3,d,br,po,pf	g/w	2, c,sbk/col	NO FIZZ
Btk 42-47	7.5YR 6/6	SL	2,d,br,po,pf	g/b-w	sg/1,m,sbk	STAGE I+
Bk 47-115+	10YR 7/2	LS	--		cemented	STAGE III

Impressions after completing description: Pit shows a variable soil as far as thickness of horizons (for example, Btk is 0-25 cm thick). Some clasts are shattered due to weathering even though upper Bandelier clasts are found. Bk is composed of chalky pockets in an indurated mass with laminated clast coatings up to 1 cm thick and some sandy, reddened pockets with only moderate fizz.

Soil Pit 2: Ridge Soil
Ridge on downthrown side of Dixon fault
Described on 8/31/95
Vegetation: Pinon/Juniper forest with sparce grass, cactus, and desert shrubs. /sparce Ponderosa pine
Mood: Racing the sun, still good
Percent pavement cover: 90

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-6	10yr 5/3	SL	v1,f,br	a/w	1,v1sbk	No Fizz
Bt1 6-50	7.5yr 3/3	SCL	3,d,co	g/s -small pit	2,m,sbk	NO FIZZ
Btk 50-60	7.5yr 4/6	SiL	2,d,br,po,pf	a/s	2,c,sbk	STAGE I
Bk/K 60-75	7.5yr 7/2	SL	--	a/s		STAGE II/III
R 75-90+	7.5yr 6/3	Si (after sampling)	--			MOD. FIZZ

Impressions after completing description: Similar to pit 1 except for shallow depth to bedrock. Carbonate is well indurated to chalky and shows some lamination on clast coatings. Possibly dissplacive carbonate. Basalts in Bt have rinds up to 3 cm thick and some clasts are shattered with carbonate in fractures. Bedrock is lightly fizzing and reddened, but still as coherent as fresh Bandelier

Explaination of abbreviations used in this appendix
Texture: Si=silt, L=loam, S=sand, C=clay. (e.g. SiL =a silty loam).
Clay Films: 1=weak, 2=moderate, 3=strong; f=faint, d=distinct, p=prominant; po=pores, br=bridges, pf=ped faces. (e.g. 3,p, po,br,pf indicated strong,prominent clay films as bridges, on pores, and on ped faces).
Boundaries: A=abrupt, c=clear, g=gradual, d=diffuse; s=smooth, w=wavy, b=broken (e.g. a/w=an abrupt, wavy boundary).
Structure: 1=weak, 2=moderate, 3=strong; s=small, m=medium; sg=single grain, g=granular, pl=platy, sbk=subangular blocky, abk=angular blocky, col=collumnar (e.g. 3,m, col/sbk= strongly developed, medium sized collumnar and subangular block structure)

Appendix 3
Continued

Soil Pit 3: Rio Soil
Just off dome road where it comes out of Rio Chiquito
Described on 9/1/95
Vegetation: Pinon/Juniper forest with sparce grass, cactus, and desert shrubs. /sparce Ponderosa pine
Mood: Nice night, good meal, good morning.
Percent pavement cover:20-90

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-10	10YR 5/3	SiL	v1,f,br	a/s	sg/1,f.sbk	NO FIZZ
Bw 10-28	10YR 5/4	SL	1,br,po,pf	g/s	2,m,sbk	NO FIZZ
Bt1 28-45	7.5YR 4/3	SCL	2,d,br,po,pf	c/s	2,c,sbk	NO FIZZ
Bt2 45-55	7.5YR 7/2	SCL	2,d,br,po,pf	g/s	2,c,sbk	NO FIZZ
Bk 55-95	7.5YR 4/6	LS	1,d,co	c/w	1,c,sbk	STAGE II+
C 95-100+	10YR 6/1	S	--		sg	NO FIZZ

Impressions after completing description: Moderately well developed soil here with some rhyolite clasts shattering. Carbonate is not laminated, but is working its way up through upper horizons.

Soil Pit 4: Ash Soil
North side of Rio chiquito, second trib downstream of dome road, in inset fill with el Cajete ash
Described on 9/1/95
Vegetation: Pinon/Juniper forest with sparce grass, cactus, and desert shrubs. /sparce Ponderosa pine
Mood: Good, it's hotter than hell.
Percent pavement cover: 75-90

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-5	10YR 5/3	SiL	v1,f,co	a/s	sg	NO FIZZ
A 5-16	10YR 5/2	SiL	1,f,co	c/s	sg/1,f,sbk.	NO FIZZ
Bk1 16-65	7.5YR 6/2	SiL	1,f,co	c/s	sg/cement	STAGE I
Bk2 65-85	10YR 7/2	carbonate	--	g/s	cemented	STAGE II
Bk3 85-125+	7.5YR 7/2	S	--		sg	STAGE I

Impressions after completing description: Only evidence of soil development is carbonate accumulation. upper portion of soil may be partially colluvial in origin. Perhapse the Av is same as other locations and A is sheetwash-derived. This seems to be developed on a thin fill inset into surface that pit 3 soil is developed on.

Explaination of abbreviations used in this appendix
Texture: Si=silt, L=loam, S=sand, C=clay. (e.g. SiL =a silty loam).
Clay Films: 1=weak, 2=moderate, 3=strong; f=faint, d=distinct, p=prominant; po=pores, br=bridges, pf=ped faces. (e.g. 3,p, po,br,pf indicated strong,prominent clay films as bridges, on pores, and on ped faces).
Boundaries: A=abrupt, c=clear, g=gradual, d=diffuse; s=smooth, w=wavy, b=broken (e.g. a/w=an abrupt, wavy boundary).
Structure: 1=weak, 2=moderate, 3=strong; s=small, m=medium; sg=single grain, g=granular, pl=platy, sbk=subangular blocky, abk=angular blocky, col=collumnar (e.g. 3,m, col/sbk= strongly developed, medium sized collumnar and subangular block structure)

Appendix 3
Continued

Soil Pit 9: Ash Soil
Just upslope from pit 4 (~50m)
Described on 2-19-96
Vegetation: P.J./sparce ponderosa
Mood: good/ok
Pavement cover: 60-90%

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-3	10YR 5/3	SiL	v1,f,co	a/s	sg	MOD.FIZZ
A 3-15	10YR 5/2	SiL	1,f,co	c/s	1,f,g	STAGE I
Bk1 15-38	7.5YR 6/2	SiL	1,f,co	g/s	1,m,sbk	STAGEI+ /II
Bk2 38-73	10YR 7/2	SiS	--	c/s	1,m,sbk	STAGE II
Bk3 73-125	10YR 7/1	S	--	c/s	sg	STAGE II
C/Bk4 125-140+	7.5YR 7/2	S	--		sg	STAGE I-

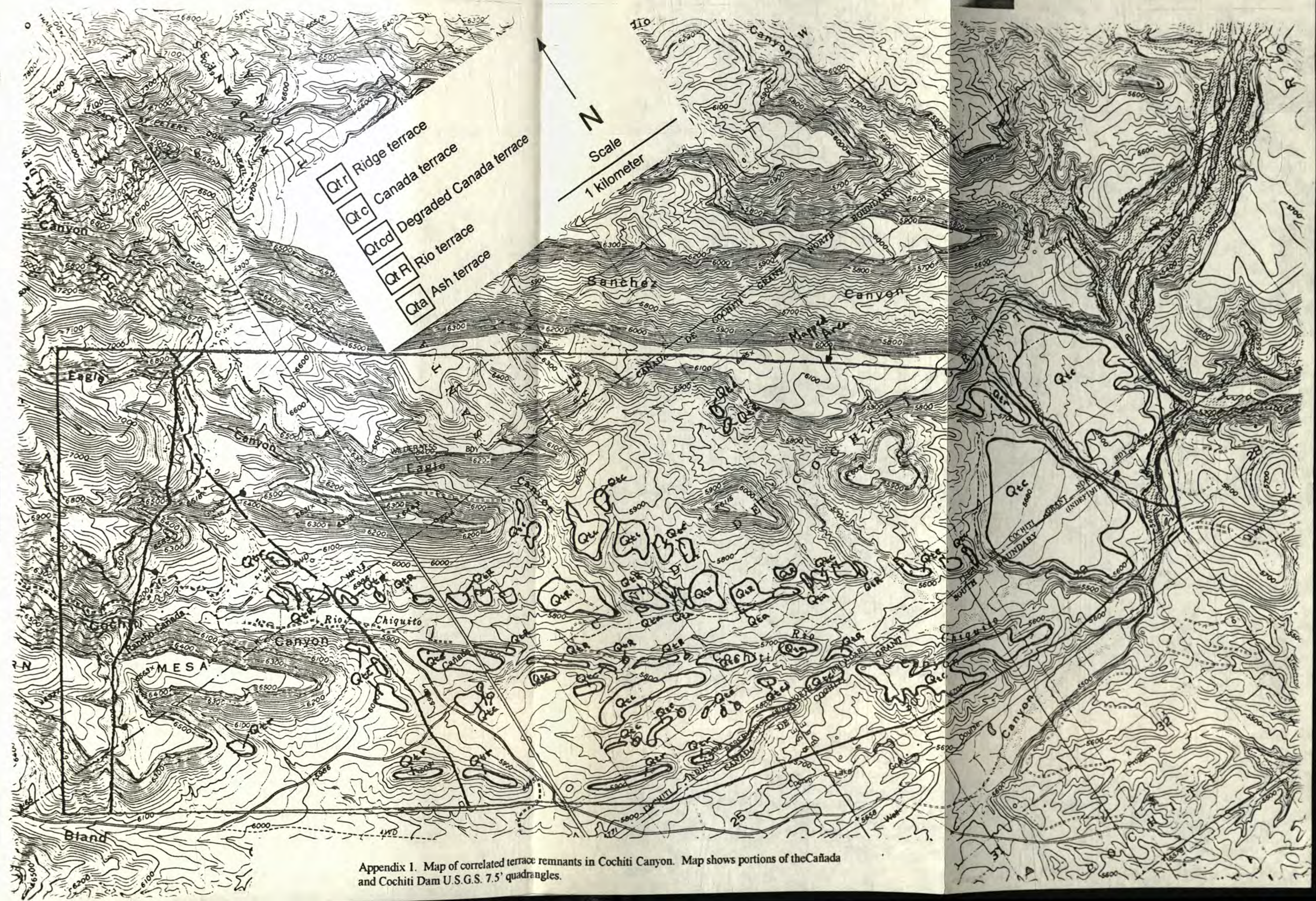
Impressions after completing description: Finer grained than pit 4, carbonate rinds on clasts often look degraded, as in pit 4. Not sure how much of upper part is colluvial here but would appear at surface like upper 30 cm ought to be. Some remnants of a carbonate layer in float on surface but could be from near-by Canada equivalent.

Soil Pit 10 Bland Canyon Soil
Described on 2-19-96
Vegetation: P.J./sparce ponderosa
Mood: good/ok
Pavement cover: 60-90%

Horizon and depth (cm)	Color (dry)	Texture	Clay films	Boundaries	Structure	Carbonate morphology
Av 0-5	10YR 5/3	SiL	v1,f,co	a/s	sg	MOD.FIZZ
A 5-14	10YR 5/2	SiL	1,f,br	c/s	1,f,g	STAGE I
Bw 14-38	7.5YR 6/3	SiL	1+,f,br,po	g/s	1,m,sbk	STAGEI+ /II
Bt 38-58	10YR 5/4	CL	2,d,po,br.pf	c/s	1,m,sbk	STAGE II
2Bw 58-90+	10YR 6/4	SiL	1,f,br,po	c/s	sg	STAGE II

Impressions after completing description: This soil seems to be somewhere in between the Ash pits and the Rio pits. There is more Bt horizon development thatn ash, but not quite as 'solid' a Bt as Rio.

Explaination of abbreviations used in this appendix
Texture: Si=silt, L=loam, S=sand, C=clay. (e.g. SiL =a silty loam).
Clay Films: 1=weak, 2=moderate, 3=strong; f=faint, d=distinct, p=prominant; po=pores, br=bridges, pf=ped faces. (e.g. 3,p, po,br,pf indicated strong,prominent clay films as bridges, on pores, and on ped faces).
Boundaries: A=abrupt, c=clear, g=gradual, d=diffuse; s=smooth, w=wavy, b=broken (e.g. a/w=an abrupt, wavy boundary).
Structure: 1=weak, 2=moderate, 3=strong; s=small, m=medium; sg=single grain, g=granular, pl=platy, sbk=subangular blocky, abk=angular blocky, col=collumnar (e.g. 3,m, col/sbk= strongly developed, medium sized collumnar and subangular block structure)



Appendix 1. Map of correlated terrace remnants in Cochiti Canyon. Map shows portions of theCañada and Cochiti Dam U.S.G.S. 7.5' quadrangles.