ASSOCIATED MIOCENE LACCOLITHS, GRAVITY SLIDES, AND VOLCANIC ROCKS, PINE VALLEY MOUNTAINS AND IRON AXIS REGION, SOUTHWESTERN UTAH

Geological Society of America 2002 Rocky Mountain Section Annual Meeting
Cedar City, Utah
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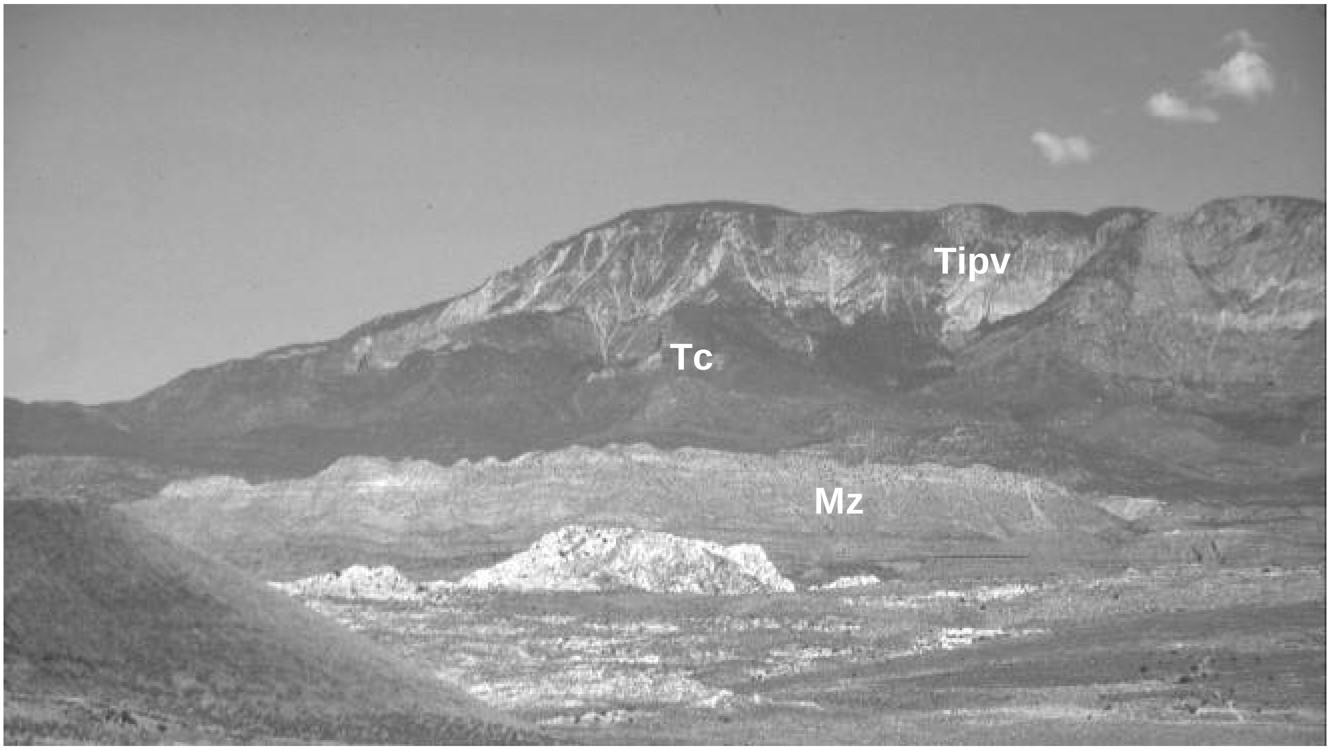

The unroofed Pine Valley laccolith (Tipv) capping the south side of the Pine Valley Mountains looking northwest from the Hurricane Cliffs. The light colored Mesozoic (Mz) rocks at the base of the mountains form the Virgin anticline, and they are overlain by Tertiary sedimentary rocks of the Claron Formation (Tc), into which the laccolith intruded and overlies.

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ABSTRACT

Detailed mapping in the Pine Valley Mountains and adjacent Bull Valley Mountains and Iron Springs mining district of southwestern Utah has delineated large allochthonous masses consisting of Tertiary volcanic and sedimentary strata laterally bounded by tear faults and soled on low-angle detachments. These masses are interpreted to be gravity slides resulting from catastrophic slope failure on oversteepened topography produced by rapidly inflating, early Miocene quartz monzonite intrusions (Pinto Peak, Iron Mountain, Bull Valley-Big Mountain, Stoddard Mountain, and Pine Valley laccoliths) of the so called "Iron Axis" magmatic province. The largest slide mass covers >150 km^2, is more than 550 m thick, and extends more than 20 km from its parent laccolithic dome. Mappable rock units within slide masses are typically broken and rotated and some are pervasively brecciated and sheared; internal folds and faults as well as disconnection and omission of rock units are common. However, the overall stratigraphic order in many instances is persevered. Deformation is confined to the allochthons, with structures terminating abruptly downward at bounding low-angle faults. These slides have resulted in emplacement both of younger rocks on older rocks and of older rocks on younger rocks.

During or closely following each sliding event, probably due at least in part to concomitant sudden release of overburden pressure, each intrusion (except Iron Mountain) erupted ash flows and (or) lava flows that partly or totally covered the associated slide mass. These deposits help date and separate the slides and identify their sources. The age of all this activity is constrained by ^{40}Ar/^{39}Ar ages of 21.93$\pm$0.07 Ma on the oldest volcanic unit of Iron Axis affinity (from the Pinto Peak intrusion) and 20.54$\pm$0.07 Ma on the youngest unit (from the Pine Valley intrusion), ages that are nearly indistinguishable from the ages of the quartz monzonite sources. On this one-day field trip we will visit and examine several of these representative allochthons and examine structural and volcanic features that bear on their mode of origin.

Evidence from recent studies supports current models of laccolith development, with initial emplacement as sills before the main phase of intrusion and doming. Laccoliths of the study area show continuous growth stages including: (1) initial sill emplacement, (2) vertical growth, (3) gravity sliding of portions of the roof, and (4) eruption. Field relationships show that the large (>200 km^2) Pine Valley laccolith is an intrusive feature (not extrusive as suggested by some earlier investigators) that was emplaced as a very shallow (<200 m) concordant intrusion.

INTRODUCTION

The purpose of this one-day field trip is to examine gravity-slide structures and volcanic deposits in the northern Pine Valley Mountains and adjacent areas of southwestern Utah (figure 1), and to evaluate field evidence bearing on their relations to laccolithic quartz monzonite intrusions. Rocks of the Pine Valley Mountains consist mostly of volcanic and intrusive rocks that range in age from Oligocene to Quaternary that were erupted upon or intruded into Mesozoic and Tertiary sedimentary rocks (figure 2). The laccolithic bodies belong to a group of more than a dozen closely related, early Miocene intrusions that constitute a magmatic province trending northeasterly across the structural transition zone between the Basin and Range and Colorado Plateau in this region, generally along the trend of the Sevier orogenic front. Because laccoliths of the Iron Springs district and eastern Bull Valley Mountains are well aligned within the belt and have produced sizable iron deposits, the belt has known informally as the "Iron Axis" (Toby, 1976; Blank and others, 1992; Rowley and others, 1995; Hacker, 1998) (figure 3). Intrusions of Iron Axis affinity were forcibly emplaced within 3.0 to 0.25 km of the surface as bulbous laccoliths, sills, and other partly concordant bodies, and were emplaced within the axial zones of some of the older, southeast-vergent Sevier thrusts and folds (Mackin, 1960).

The largest Iron Axis intrusion forms the gigantic (>200 km^2) igneous mass capping the Pine Valley Mountains. Previous studies have placed its origin either as an extrusive volcanic dome (Mattison, 1976; Grant, 1991) or as an intrusive laccolith of world-class size areal extent and volume (Cook, 1957, 1960). Recent work has aimed at resolving differences of interpretation of this unique feature. A fruitful approach has proved to be detailed mapping directed towards delineation of and identification of the sources of far-traveled slide masses that involve roof rocks of the several quartz monzonite-cored uplifts in the vicinity, including the Pine Valley Mountains body (Hacker, 1998). Similar slide masses were earlier recognized and mapped in the Iron Springs mining district (Mackin, 1947, Blank and Mackin, 1967) and in the Bull Valley Mountains (Blank, 1959, 1993); more recent work focused on gravity slides in the northern Pine Valley Mountains (Hacker, 1998). It can be inferred from field relations that slides are the result of slope failure on oversteepened topography produced by the rapid rise of laccolithic intrusions. This slide-laccolith association was the focus of an earlier Geological Society of America field trip in the Iron Axis region (Blank and others, 1992). Moreover, sliding was often synchronous with or immediately followed by volcanic eruption from the flank of the associated intrusion, doubtless triggered at least in part by the sudden release of overburden pressure and accompanied by earthquakes. As a result, the slide masses are typically blanketed by ash-flow tuffs, tuff breccias, and lava flows that post-date volcanics of the cover rock succession. This sequence of events (1-doming, 2-gravity sliding, 3-volcanism) has been documented elsewhere for initiating violent eruptions by release of confining pressure on magma chambers within large volcanic cones (e.g., see Lipman and Mullineaux, 1981). The association of gravity sliding and volcanism on the flanks of laccoliths is less well known; therefore, a summary of new evidence pertaining to the geometry, scale, and timing of gravity faulting and their relationships to the growth and eruptive history of laccoliths is presented here. We also discuss a model for the growth of the Iron Axis laccoliths.

Gravity slides are recognized in the field as highly sheared, allochthonous rock masses soled on low-angle faults and laterally bounded by tear faults. A typical mass is pervasively faulted, fractured and locally pulverized, with much internal rotation and interpenetration of constituent formations. Duplication and omission of strata are common although in most places the overall stratigraphic order is preserved. Internal faults terminate downward at the sole fault, which may juxtapose younger over older rocks or vice versa. The slide sole faults of the study area are distinctly different from well known other low-angle fault systems such as the Sevier thrust faults of Cretaceous age and Neogene basin-range detachment faults (see figure 4). They do resemble soles of gravity slides associated with collapse of regional high-angle fault scarps (e.g., Swadley and others, 1994), but candidate fault scarps that pre-date basin-range faults, yet are not laccolith-related, are lacking in the Pine Valley Mountains area. Mapping of the stratigraphy and structures of the displaced slide masses confirms their relation to intrusive doming.

GEOLOGIC SETTING

The Pine Valley Mountains are situated in a structural and stratigraphic transition zone between the Colorado Plateau and the Basin and Range physiographic provinces (figure 5). During the Paleozoic and early Mesozoic, this area occupied the edge of the shelf of the North American continent, and it accumulated a thick wedge of westerly thickening shelf sediments that represented deeper water farther to the west. In the late Cretaceous and Paleocene, the area was affected by the Sevier orogeny where deformation consisted manly of

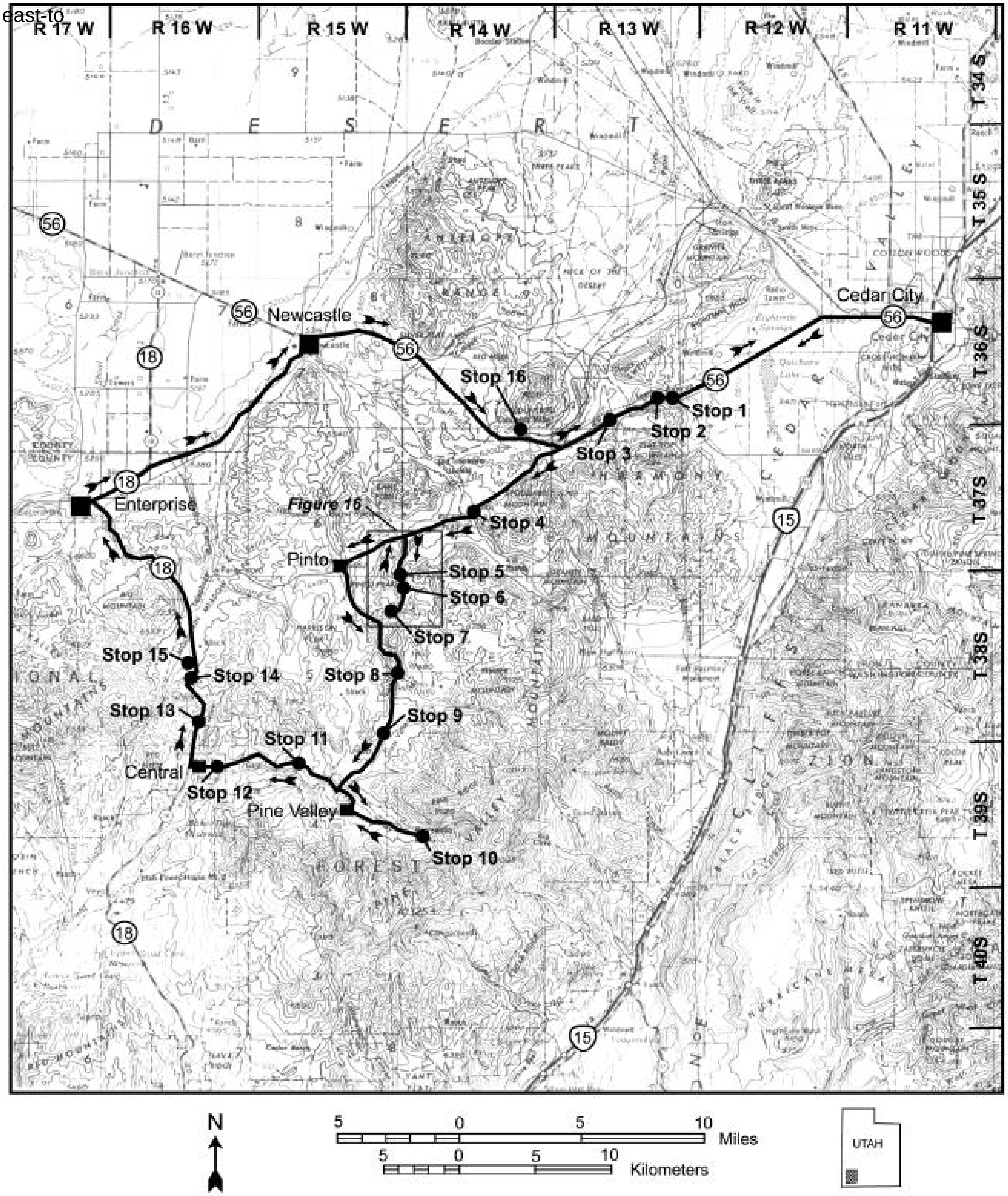


Figure 1. Map of the Pine Valley Mountains area showing location of stops. Based on U.S.G.S. Cedar City 1:250,000 topographic map NJ 12-7, 1971 (revised)

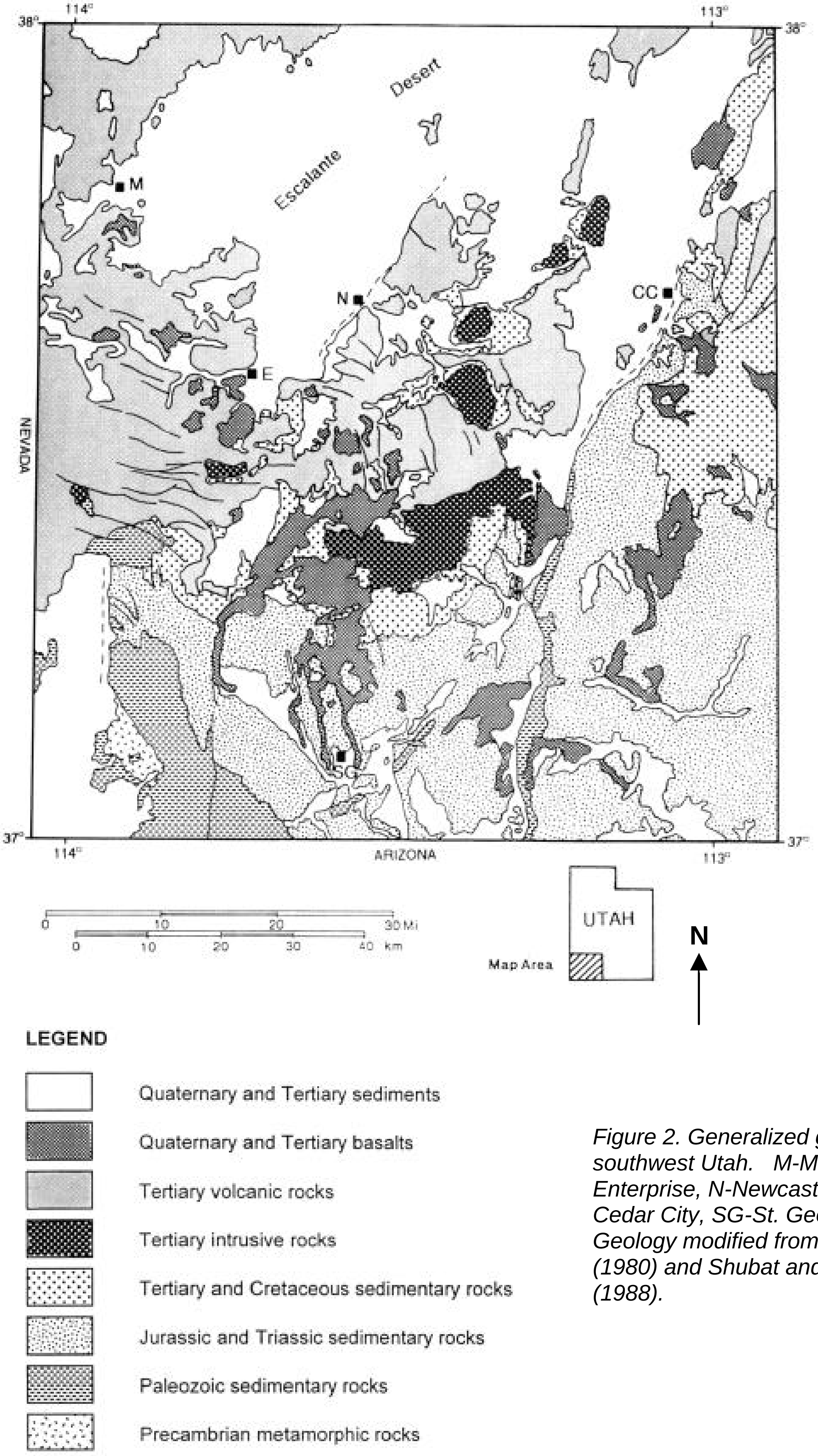


Figure 2. Generalized geology of southwest Utah. M-Modena, E-Enterprise, N-Newcastle, CC-Cedar City, SG-St. George. Geology modified from Hintze (1980) and Shubat and Siders (1988).

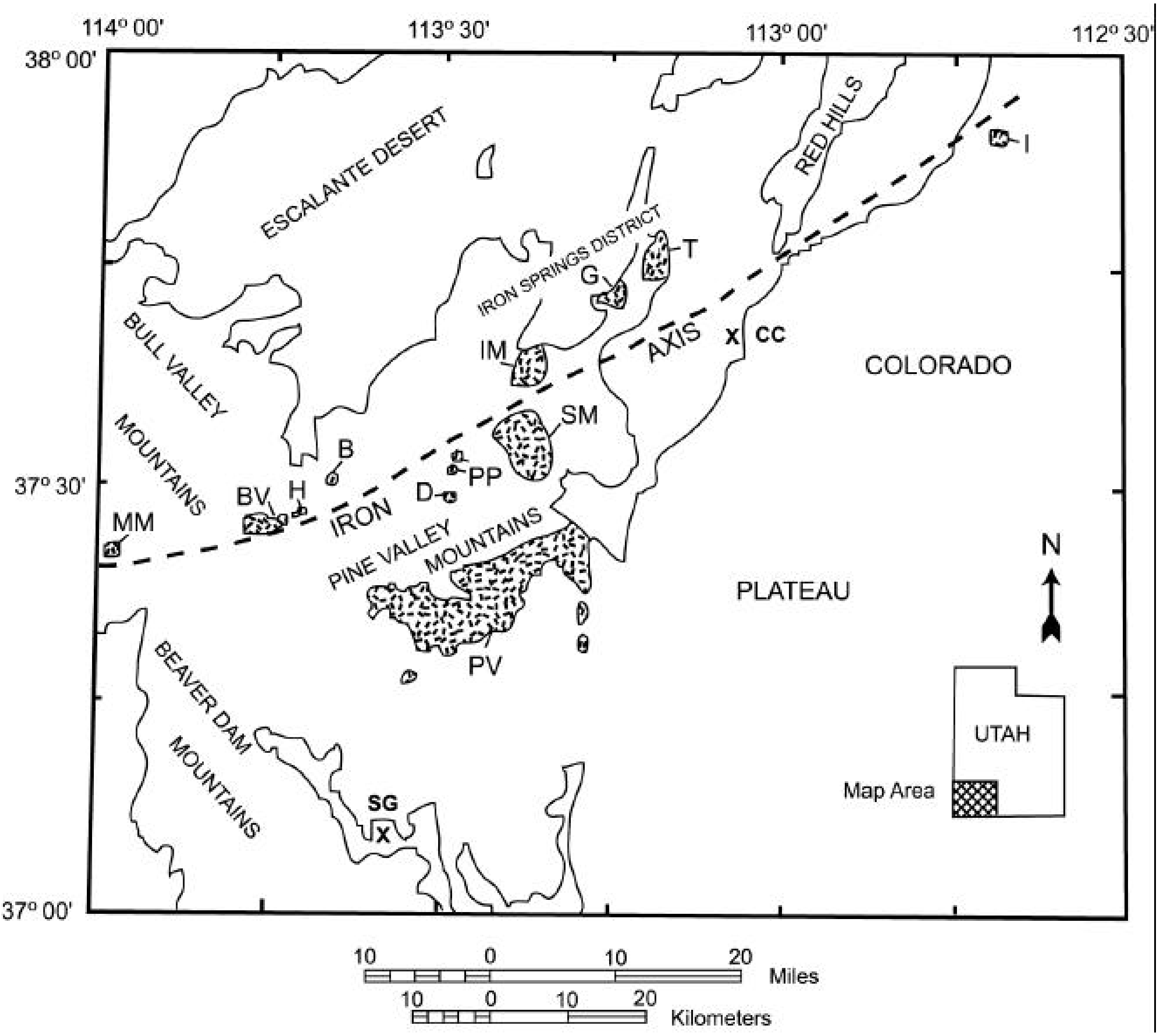


Figure 3. Index map of the Iron Axis region of southwest Utah, showing area of study. Intrusions of and near the Iron Axis are stippled and have the following name symbols: B - Big Mountain; BV - Bull Valley; D - The Dairy; G - Granite Mountain; H -Hardscrabble Hollow; I - Iron Peak; IM - Iron Mountain; LP - Lookout Point; MM - Mineral Mountain; PP - Pinto Peak; PV - Pine Valley; SM - Stoddard Mountain; T - Three Peaks. Towns: CC - Cedar City; SG - St. George.

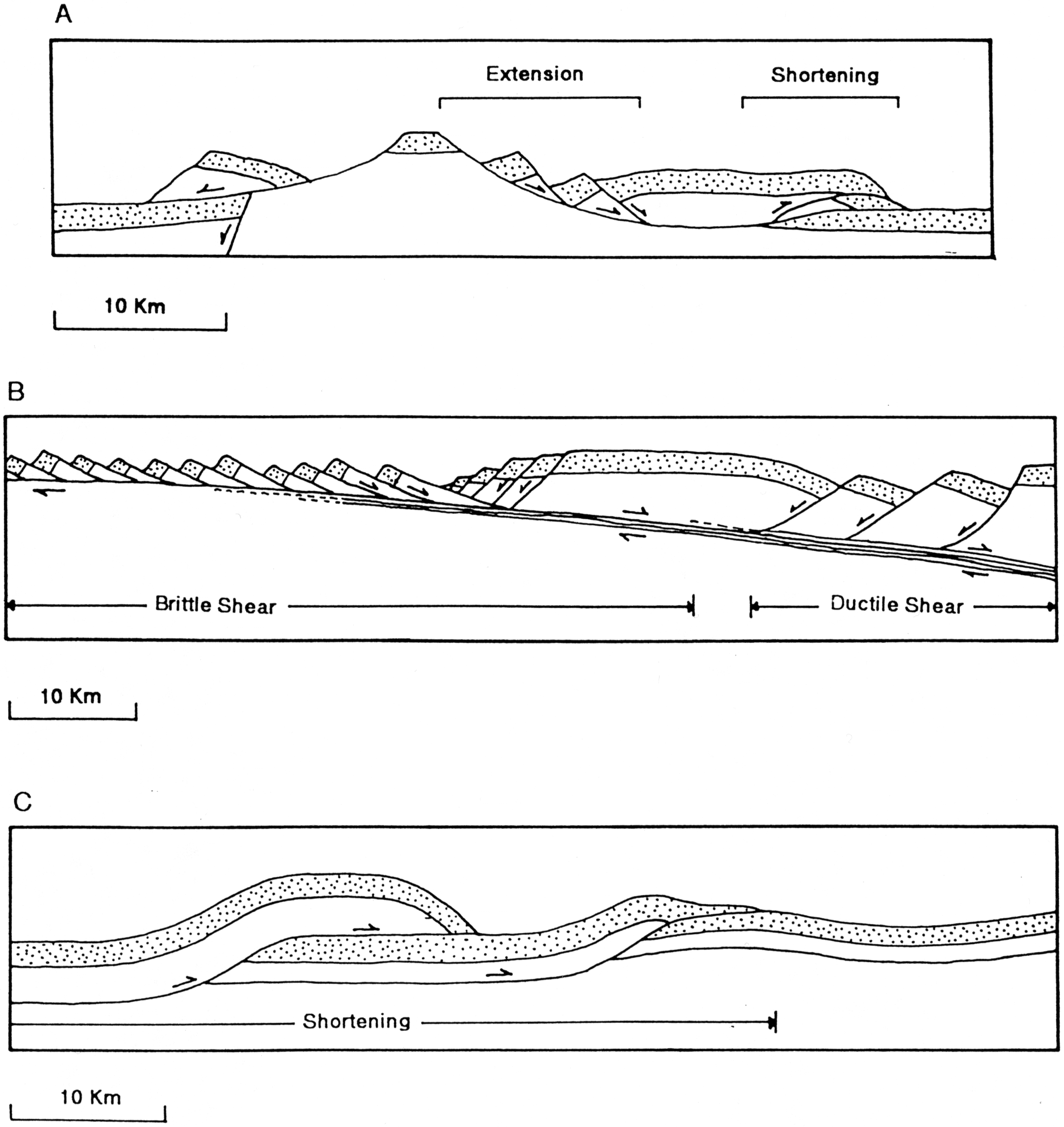


Figure 4. Diagrammatic sketches of three types of low-angle faults found in the Basin and Range. A-Gravity-slides, B-Detachment faults, and C-Thrust faults. Modified from Wernicke (1981) and Van Kooten (1988).

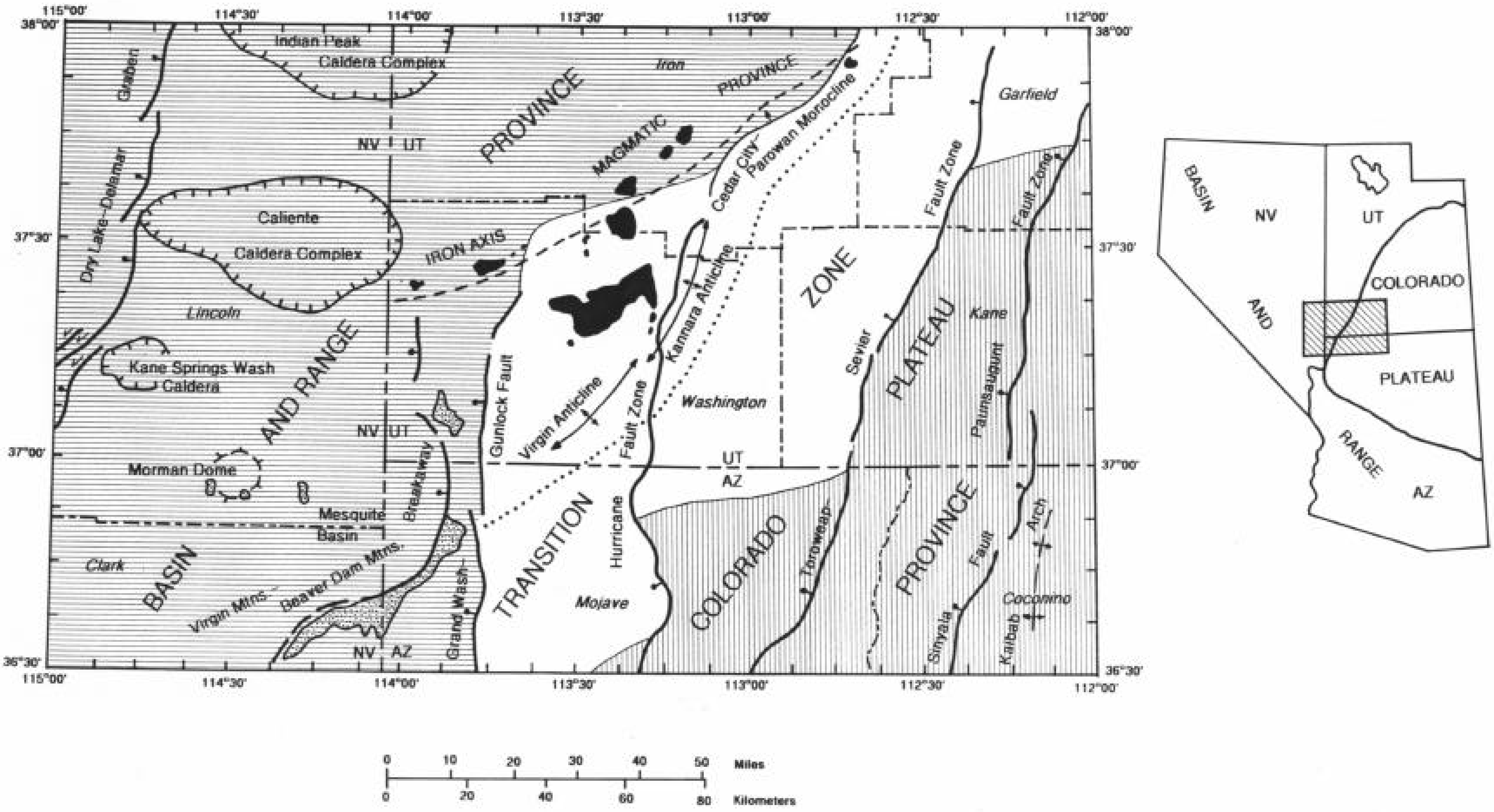


Figure 5. Tectonic map of southwest Utah, southeast Nevada, and northwest Arizona showing trends of major structures (modified from Blank and Kucks [1989] and Scott and Swadley [1995]).

southeast-directed, thin-skinned ramp-decollement style thrusting. Not coincidentally, the easternmost extent of Mesozoic thrusting and folding in Utah corresponds closely to the structural shelf hinge that marks the eastward change from thick “miogeoclinal” shelf strata to thinner strata of cratonic platform facies (Hintze, 1986). In southwestern Utah, the easternmost extent of major exposed Sevier thrusting is represented by the Square Top Mountain thrust fault, located in the northern Beaver Dam Mountains to the west of the Pine Valley Mountains. East of the Square Top Mountain thrust, a Sevier-age northeast-trending anticlinal structure extends mostly in the subsurface from the Bull Valley Mountains through the Iron Springs district to the Red Hills. The Iron Springs thrust, also known as the Iron Springs Gap thrust of Mackin (1947) and the Calumet fault of Lewis (1958), locally displaces the fault. This fault shows about 5.6 km of southeastward displacement and places strata of the Jurassic Carmel Formation over units of the Cretaceous Iron Springs Formation (Van Kooten, 1988). Other related folds include the Virgin anticline-Kanarra fold system, which is east-southeast of the Pine Valley Mountains and extends from southeast of St. George northeast past Cedar City (Threet, 1963; Kurie, 1966).

Following the Sevier orogeny, erosion reduced the structural and topographic forms in the Pine Valley Mountains area during the Late Cretaceous and early Paleocene. A broad basin later developed in the region, beginning in the late Paleocene (Goldstrand, 1994), into which interbedded conglomerate, sandstone, and limestone were deposited within alternating fluvial, deltaic, and lacustrine environments. Conditions in southwest Utah changed dramatically in early Oligocene time with the onset of calc-alkalic igneous activity at about 34 to 33 Ma (Anderson and Rowley, 1975; Best and others, 1989; Rowley and others, 1994; Rowley, 1998; Rowley and Dixon, 2001) and continuing into the early Miocene. The Pine Valley Mountains area was a relatively flat plain during this time, across which regional calc-alkaline ash-flow tuff sheets were deposited from sources outside the area. These sources included the Oligocene Indian Peak caldera complex (Best and others, 1989) to the northwest, and the Caliente caldera complex (Williams, 1967; Noble and McKee, 1972; Ekren and others, 1977; Rowley and others, 1995, 2001) to the west (figure 5). Both caldera complexes consist of numerous nested and partially overlapping calderas produced during catastrophic eruption cycles. Locally the tuffs intertongue with lava flows and laharic breccia from andesitic volcanoes and fissure eruptions.

Structural and topographic relief returned to the Pine Valley Mountains area again during an episode of early Miocene magmatic activity, starting at about 22 Ma, which produced the series of shallow, calc-alkaline laccolithic intrusions of the Iron Axis magmatic province (figure 3 and 5). Other intrusions in the province are known in the subsurface through drilling or have been inferred from their roof rock structures and aeromagnetic signatures. Geophysical data suggest that many of the intrusions are interconnected at depth and are inferred to be underlain and fed by a larger batholith complex (Mackin, 1947; Blank and Mackin, 1967; Blank and Kucks, 1989; Blank and others, 1992; Rowley and others, 1994, 1998, 2001). Most of the Iron Axis intrusions are quartz monzonite to granodiorite porphyries (Blank and others, 1992). However, the Mineral Mountain intrusion at the southwest end of the belt is a granite porphyry (Morris, 1980; Adair, 1986), and the Iron Peak intrusion, at the northeast end, is a gabbro-diorite porphyry (Spurney, 1984). The intrusions were forcibly emplaced into Paleozoic, Mesozoic, and Tertiary sedimentary rocks to form laccoliths, sills, and other partly concordant bodies that strongly deformed their roofs by folding and faulting. The alignment of the intrusions closely coincides with the northeast trend of earlier Sevier orogenic structures (Mackin, 1954, 1960; Blank, 1959; Van Kooten, 1988). The intrusions in the eastern Bull Valley Mountains (Bull Valley, Hardscrabble Hollow, and Big Mountain) and Iron Springs district (Iron Mountain, Granite Mountain, and Three Peaks) intruded into limestone strata of the Jurassic Carmel Formation and produced major replacement iron ore bodies (Mackin, 1968) that led to the term “Iron Axis” (Tobey, 1976). At least four intrusions (Pinto Peak, Bull Valley, Stoddard Mountain, and Pine Valley) broke through their roofs and erupted ash-flow tuffs and (or) lava flows (Blank, 1959; Hacker, 1995, 1998; Hacker and others, 1996). Most of the major structures within the Pine Valley Mountains are associated with the laccolithic intrusions emplaced during the Iron Axis magmatic episode.

Following Iron Axis magmatic activity (post-22 Ma), the area once again received additional regional calc-alkaline ash-flow deposits from the Caliente caldera complex to the west. Beginning at about 14-15 Ma, a change from calc-alkaline to bimodal (basalt and high silica rhyolite) magmatism began in the region and has persisted sporadically throughout the late Cenozoic and was accompanied by regional basin-range extensional faulting. Bimodal magmatism produced the extensive basaltic lava flows and associated cinder cones found in southwest Utah (see figure 2), but their total volume is much less than that of the earlier calc-alkaline volcanic rocks (Rowley and Dixon, 2001). The basin-range style tectonic extension imprinted a dominantly northerly striking fault pattern upon previously formed structures. Regional crustal extension in the Basin and Range occurred through a complex combination of displacements along low-angle normal, high-angle normal, and strike-slip fault systems (Moores and others, 1968; Anderson, 1971; Armstrong, 1972; Wernicke and others, 1988; Anderson and Barnhard, 1993; Rowley and Dixon, 2001). Extension in the region occurred principally during the last 20 Ma (Anderson and Mehnert, 1979; Wernicke and others, 1988; Rowley and Dixon, 2001) and appears to have taken place in two episodes with two slightly different extension directions.

The present topographic relief of this eastern part of the Basin and Range Province is primarily the result of high-angle, basin-range block faulting (horst and graben style) that began in the late Miocene, at about 10 to 8 Ma (Anderson and Mehnert, 1979; Rowley and others, 1979; 1981; Shubat and Siders, 1988; Rowley and Dixon, 2001). Steep, north-to-northeast-trending, high-angle normal faults are abundant. Synextensional sediments eroded from the uplifted blocks were deposited in the adjacent developing basins, and now generally occupy more area than the ranges. Many basins have accumulated very thick sections of basin fill, such as the Newcastle graben west of the Antelope Range in the southern Escalante desert. In contrast, the Colorado Plateau to the east remained an undeformed stable highland. The boundary between the Basin and Range Province and Colorado Plateau is not sharp in southwest Utah; instead they are separated by a structural transition zone that varies in width from about 30 to 100 km (figure 5). The area of transition is characterized by high-angle faults similar to those in the Basin and Range but generally they are less abundant and have less displacement (Anderson and Rowley, 1975; Rowley and others, 1979, 1998; Scott and Swadley, 1995; Rowley 1998; Rowley and Dixon, 2001), and the intervening basins occupy less area than the ranges and are filled with a thinner sequence of alluvial fill or do not receive erosional fill at all.

STRATIGRAPHIC AND IGNEOUS UNITS

The rock section of the Pine Valley Mountains can be divided into four distinct sequences based on their associated origin before, with, or following Iron Axis magmatic activity (figure 6). Pre-Iron Axis rocks include a lower sequence of in-place sedimentary rocks, which hosted the Iron Axis intrusions, and an upper sequence of mostly regionally deposited volcanic rocks that later became part of the allochthonous gravity-slide masses derived from the roofs of the intrusions. The Iron Axis rocks include a sequence of volcanic and intrusive igneous rocks related to magmas that formed laccolithic structures. The post-Iron Axis rocks include regionally and locally derived volcanic rocks and local sedimentary rocks that formed contemporaneously with post-Iron Axis extensional structures.

Pre-Iron Axis Sedimentary Rocks

The pre-Iron Axis sedimentary rocks consist of 2,700 m of Mesozoic and Cenozoic strata, belonging to seven well-known and regionally extensive formations (figure 7). The pre-Iron Axis sedimentary rocks are exposed mostly in the southern part of the range and also in the eroded cores and flanks of intrusive anticlines and gravity-slide masses. The Mesozoic section consists of clastic sedimentary rocks of continental origin along with minor shallow-marine carbonates and evaporates of the Jurassic Kayenta Formation, Navajo Sandstone, Temple Cap Formation, and Carmel Formation, and the Cretaceous Dakota Conglomerate and Iron Springs Formation (Hintze, 1986). The sedimentary sequence is interrupted by at least two regional unconformities, one at the base of the Temple Cap Formation, and the other at the base of the Dakota Conglomerate. The Cretaceous section represents conglomeratic to shaly clastic material shed eastward from the Sevier orogenic belt in eastern Nevada and western Utah into a foreland basin (Hintze, 1986).

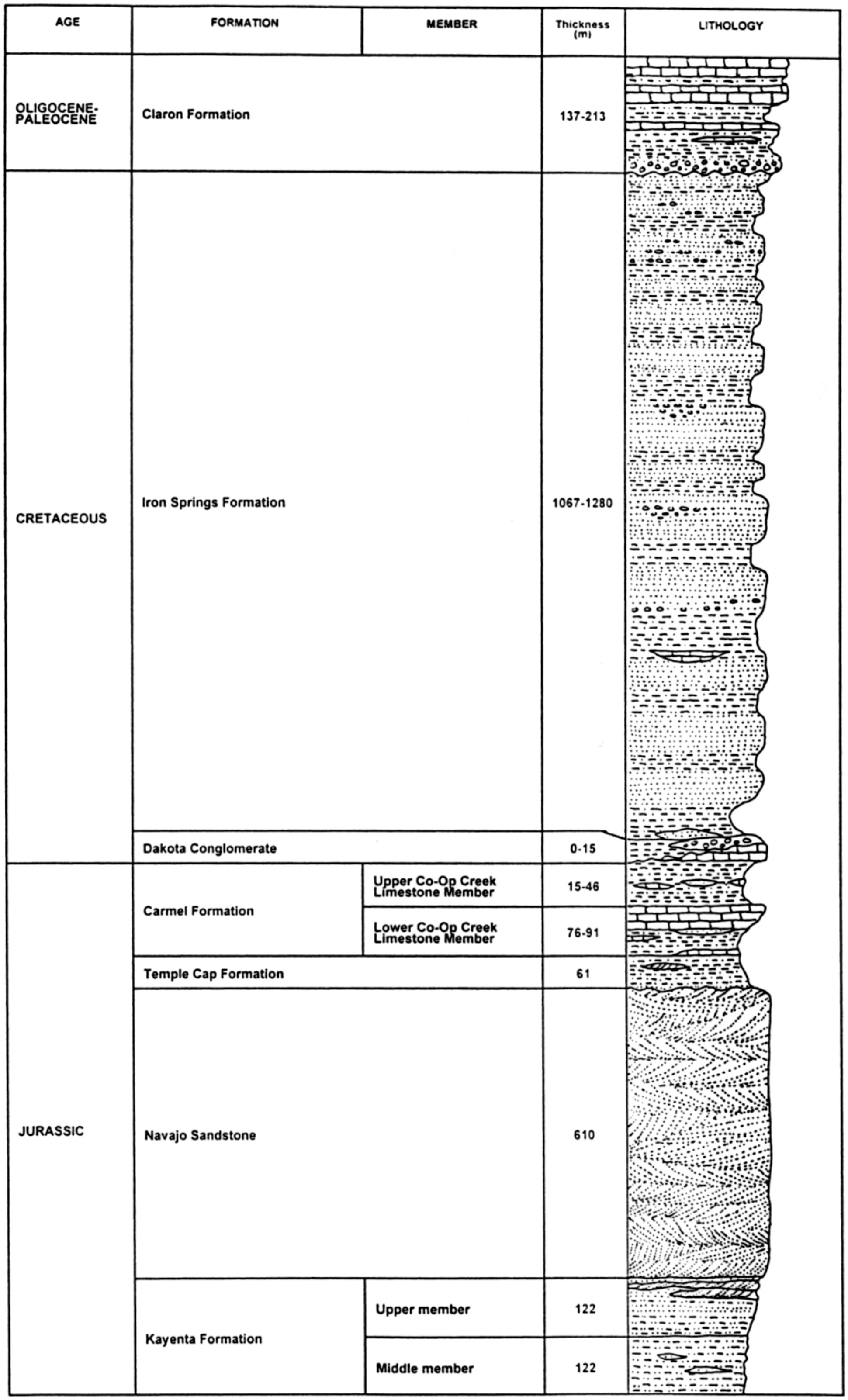
The Jurassic to Cretaceous sequence was thrusted and folded in the Late Cretaceous to form subaerial topographic features later planed down by erosion and overlain unconformably by Paleocene to Oligocene fluvial and lacustrine sedimentary rocks of the Claron Formation (Cook, 1960). The Claron forms the basal Tertiary unit throughout the Pine Valley Mountains as well as over much of southwestern Utah. The Claron and Iron Springs Formations were the host rocks for Iron Axis intrusions in the Pine Valley Mountains (Cook, 1957; Hacker, 1998) while in the Iron Springs district and Bull Valley Mountains, the Temple Cap and Carmel Formations were the host rocks for intrusions (Blank, 1959, 1993; Blank and Mackin, 1967).

Pre-Iron Axis Volcanic Rocks

The pre-Iron Axis volcanic rocks (figure 8) consist of late Oligocene to early Miocene (30 to 22 Ma) regional ash-flow tuff deposits derived from distant sources to the northwest and west. Their ages are known from several published isotopic and fission track dates (see Anderson and Rowley, 1975; Best and others, 1989; Rowley and others, 1995). They include ash-flow tuffs of the Wah Wah Springs Formation (of the Needles Range Group) and the Isom Formation, both derived from the Indian Peak caldera complex to the north; ash-flow tuffs of the Leach Canyon Formation and Condor Canyon Formation (of the Quichipa Group) derived from the Caliente caldera complex to the northwest; and the Harmony Hills Tuff (of the Quichapa Group) from the Bull Valley Mountains (see figure 5 for caldera locations). This volcanic sequence shows an overall thinning from

AGE	(Ma)	FORMATION			MEMBER	Thickness (m)	
HOLOCENE-PLEISTOCENE		Surficial deposits-alluvial, mass wasting				0-46	**Post-Iron Axis Volcanic and Sedimentary Rocks**
		Basaltic lava flows				0-61	
	1.6	Boulder deposits				0-61	
PLEISTOCENE-PLIOCENE		Eight Mile Dacite				0-152	
		Older piedmont-slope deposits				0-91	
PLIOCENE		Older basaltic lava flows				0-61	
MIOCENE		Alluvial deposits of Spring Creek				0-91	
	19	Racer Canyon Tuff				0-91	
	20.5	Pine Valley intrusion	Pine Valley Latite		Timber Mountain member	0-610	**Iron Axis Rocks**
					Rencher Peak member	0-457	
		Page Ranch Formation				0-122	
		The Dairy intrusion	Stoddard Mountain intrusion	Volcanic rocks of Comanche Canyon	Lava flow member	0-152	
					Ash-flow tuff member	0-61	
	21.5	Rencher Formation (Derived from Bull Valley Intrusion)			Upper ash-flow tuff member	0-46	
					Lower ash-flow tuff member	0-107	
	21.9	Pinto Peak intrusion	Rocks of Paradise		Sandstone member	0-7.6	
					Upper lava flow member	0-183	
					Lower lava flow member	0-122	
					Ash-flow tuff Member	0-91	
		Local volcanic rocks				0-24	**Pre-Iron Axis Volcanic Rocks**
	22.5	Quichapa Group	Harmony Hills Tuff			30-99	
			Little Creek andesite -allochthonous only in Pine Valley Mountains			0-183	
			Local volcanic mudflow breccia			4.5	
	23		Condor Canyon Formation		Bauers Tuff Member	24-61	
					Swett Tuff Member	0-9	
			Local sedimentary rocks			2	
	24		Leach Canyon Formation			46-152	
OLIGOCENE	27-26	Isom Formation			Hole-in-the-Wall Tuff Mbr	12	
					Baldhills Tuff Member Associatedlava flows and sedimentary rocks	15-30	
	30	Wah Wah Springs Fm., Needles Range Group				0-9	
OLIGOCENE-PALEOCENE		Claron Formation				137-213	**Pre-Iron Axis Sedimentary Rocks**
CRETACEOUS		Iron Springs Formation				1067-1280	
		Dakota Conglomerate				0-15	
JURASSIC		Carmel Formation			Upper Co-Op Creek Limestone Member	15-46	
					Lower Co-Op Creek Limestone Member	76-91	
		Temple Cap Formation				61	
		Navajo Sandstone				610	
		Kayenta Formation			Upper member	122	
					Middle member	122	

Figure 6. Composite table of stratigraphic (sedimentary and volcanic) and lithodemic (igneous intrusive) units mapped in the Pine Valley Mountains. Thicknesses relate to stratigraphic units only.



AGE	FORMATION	MEMBER	Thickness (m)	LITHOLOGY
OLIGOCENE-PALEOCENE	Claron Formation		137-213	
CRETACEOUS	Iron Springs Formation		1067-1280	
	Dakota Conglomerate		0-15	
JURASSIC	Carmel Formation	Upper Co-Op Creek Limestone Member	15-46	
		Lower Co-Op Creek Limestone Member	76-91	
	Temple Cap Formation		61	
	Navajo Sandstone		610	
	Kayenta Formation	Upper member	122	
		Middle member	122	

Figure 7. Generalized stratigraphic column of pre-Iron Axis sedimentary rocks.

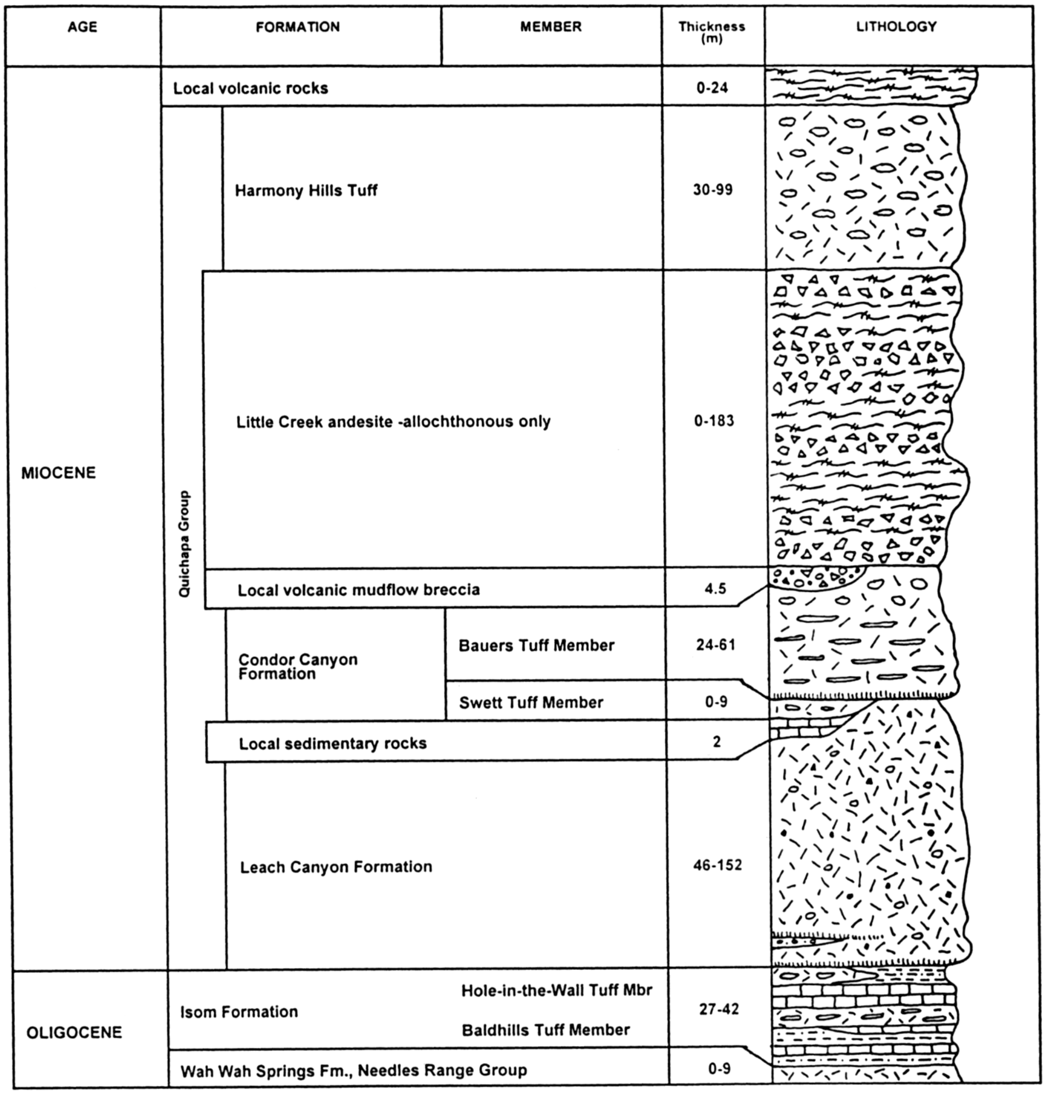
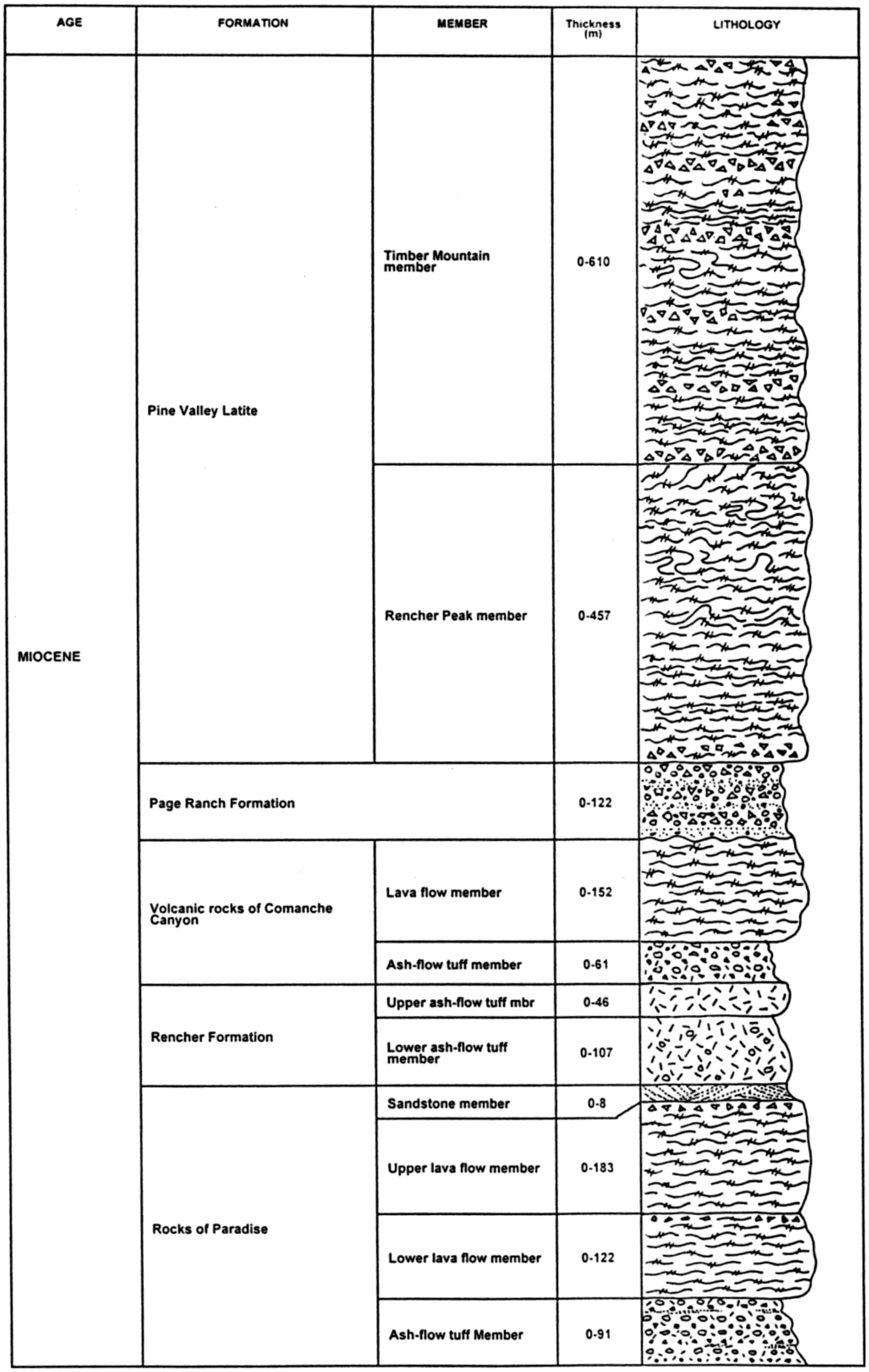


Figure 8. Generalized stratigraphic column of pre-Iron Axis volcanic rocks.

north to south within the Pine Valley Mountains that is reflective of their source areas being located to the north. The lower ash-flow tuffs (Wah Wah Springs, Isom, and Leach Canyon Formations) were deposited in a still-active Claron depositional basin where the tuffs became intercalated with lacustrine limestone and volcaniclastic sedimentary rocks. The sequence also contains some local lava flows that proved to be very valuable in determining the source areas of various gravity-slide masses derived from the Iron Axis intrusions. Some flows mapped within the gravity-slides proved to be totally allochthonous and could be correlated to autochthonous units located within the Bull Valley Mountains (e.g., Little Creek andesite unit). The pre-Iron Axis volcanic rocks crop out throughout the northern part of the study area within the eroded flanks of laccoliths or in parts of their preserved roof sequence, in erosional valleys, and in remnants of gravity-slides. The original southern extent of the pre-Iron Axis volcanic rocks is unknown due to extensive erosion, but are believed to have been present south of the Pine Valley intrusion based on outcrops found adjacent to the intrusion on its south side.

Iron Axis Rocks

The Iron Axis rocks (figure 9) exposed in the Pine Valley Mountains consist of early Miocene (22 to 20 Ma) hypabyssal, quartz monzonite, porphyry intrusive rocks and associated extrusive volcanics produced as part of the regional Iron Axis magmatic episode. The Pine Valley intrusion caps a large portion of the range. It crops



AGE	FORMATION	MEMBER	Thickness (m)	LITHOLOGY
MIOCENE	Pine Valley Latite	Timber Mountain member	0-610	
		Rencher Peak member	0-457	
	Page Ranch Formation		0-122	
	Volcanic rocks of Comanche Canyon	Lava flow member	0-152	
		Ash-flow tuff member	0-61	
	Rencher Formation	Upper ash-flow tuff mbr	0-46	
		Lower ash-flow tuff member	0-107	
	Rocks of Paradise	Sandstone member	0-8	
		Upper lava flow member	0-183	
		Lower lava flow member	0-122	
		Ash-flow tuff Member	0-91	

Figure 9. Generalized stratigraphic column of Iron-Axis rocks.

out over a 240 km^2 area and has a remaining thickness of over 900 m. The Iron Axis rocks crop out in the central core of anticlines, where erosion has breached the pre-Iron Axis roof rocks to expose intrusive rocks, or in the preserved paleo-lowlands between the laccoliths where the volcanic rocks were deposited. Volcanic activity during this magmatic episode was previously thought to be represented by ash-flow tuff and lava flow deposits of the Rencher Formation and lava flows of the younger Pine Valley Latite. Cook (1957) defined both formations from exposures in the Pine Valley Mountains, but did not document their source areas. Blank (1959) later documented that the volcanic rocks of the Rencher Formation erupted from the Bull Valley Intrusion. Cook (1957) believed the Pine Valley Latite covered its vent area and was later intruded by the Pine Valley intrusion (Cook, 1957).

Detailed mapping by Hacker (1998) and Rowley (unpublished data) has revealed that volcanic units of Rencher lithology have more than one source area. Field evidence shows that the Pinto Peak and Stoddard Mountain intrusions in the northern Pine Valley Mountains both erupted a sequence of ash-flow tuffs and lava flows distinct from the Rencher Formation. These are named the "rocks of Paradise" and the "volcanic rocks of Comanche Canyon" respectively. Field evidence also shows that the previously defined Paradise intrusion of Cook (1957) is in fact a large lava flow derived from the Pinto Peak intrusion, and was included as a member of the rocks of Paradise (Hacker, 1998). The name "Rencher Formation" is retained, but restricted to volcanic rocks derived only from the Bull Valley intrusion. Finally, mapping indicates that the Pine Valley Latite erupted from two source areas: (1) at Rencher Peak, possibly from an early episode of the Pine Valley intrusion, and (2) directly from the main Pine Valley intrusion. Figure 10 shows the inferred areal distribution of the volcanic units erupted from Iron Axis intrusions. Locally derived volcaniclastic rocks of the Page Ranch Formation are genetically related to erosion of the intrusive uplifts and provide a marker unit that aided in deciphering the structural history of the intrusions.

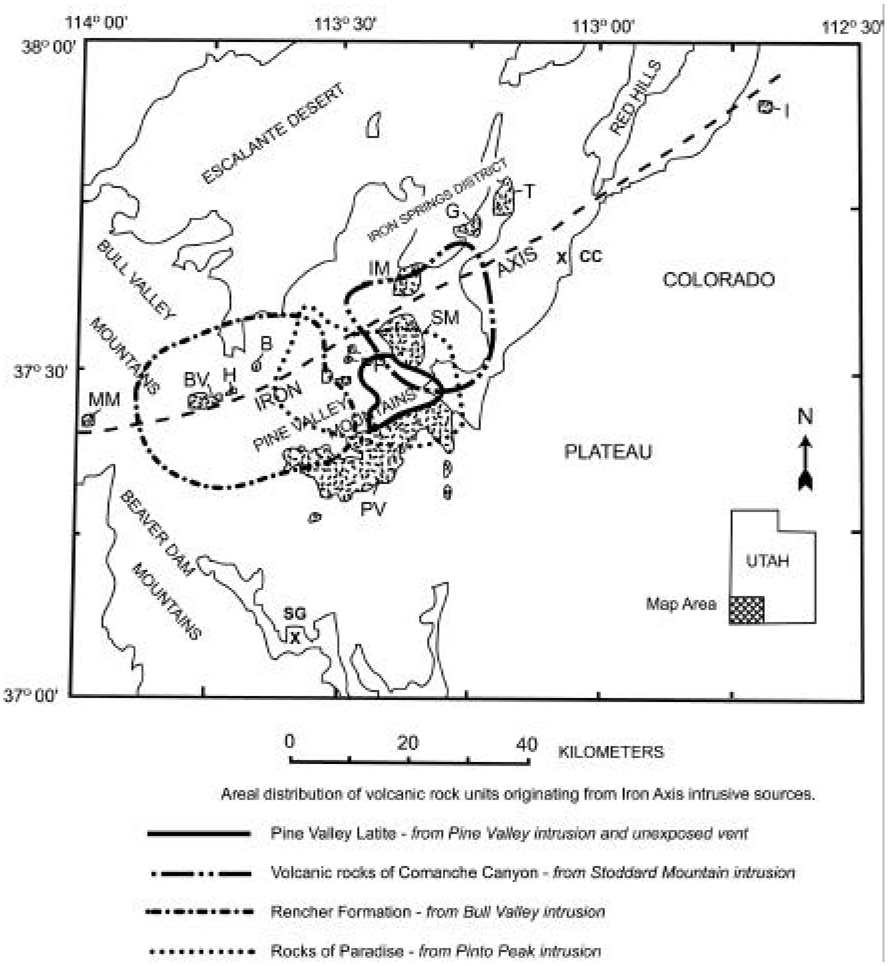
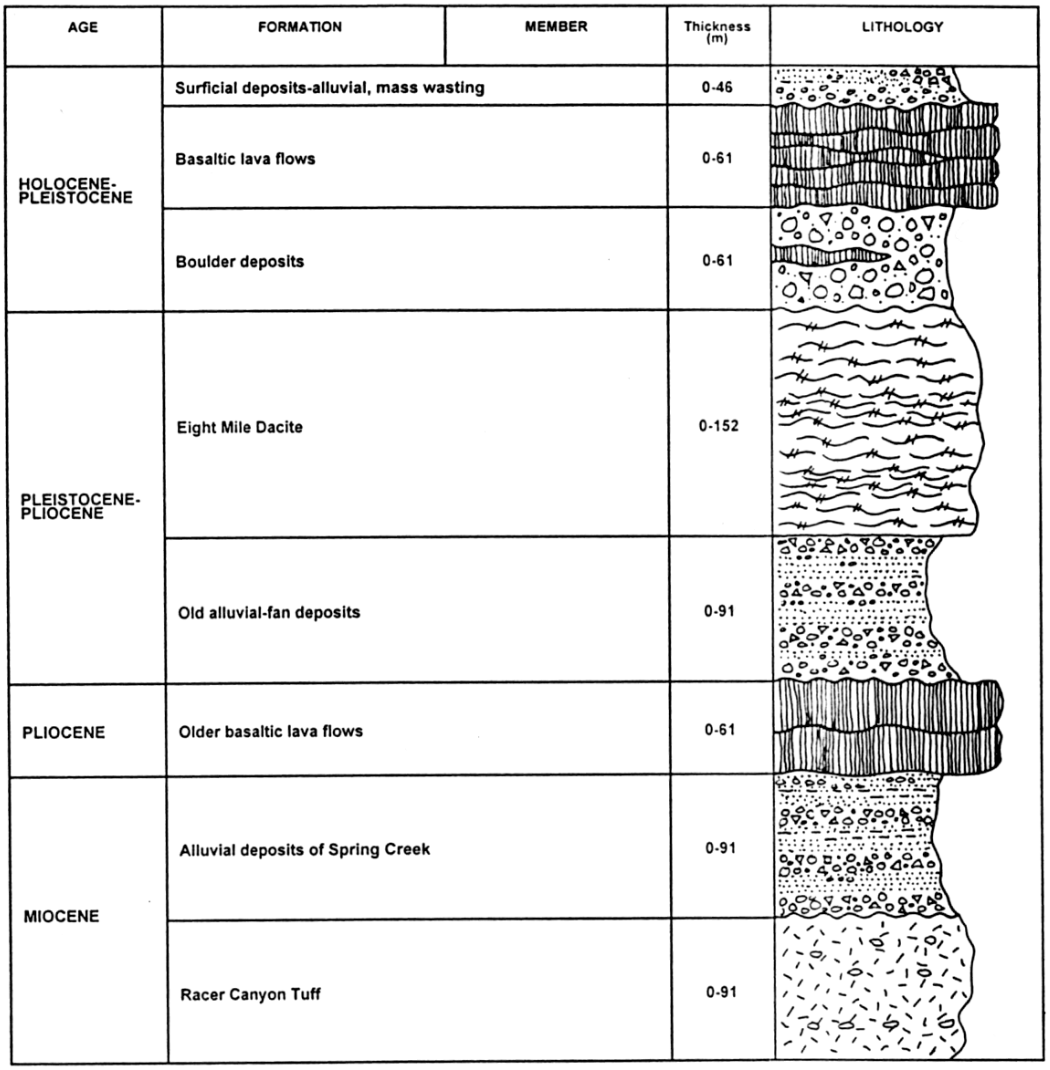


Figure 10. Generalized map of southwest Utah showing areal distribution of volcanic rocks erupted from laccoliths.

Post-Iron Axis Rocks

The post-Iron Axis rocks (figure 11) consist of upper Tertiary and Quaternary (19 Ma to recent) regional and local volcanic rocks deposited over a rugged topographic area developed by earlier Iron Axis intrusions and later by their erosion. The last regional ash-flow tuff to enter the area was the Racer Canyon Tuff derived from the Caliente caldera complex. Topographic barriers raised by the earlier intrusive structures to the south probably resulted in the thick Racer Canyon Tuff being restricted to the very northern part of the study area. Erosion of the remaining highlands in the late Miocene resulted in the deposition of alluvial deposits in the north that are now capped by late Tertiary (Miocene and Pliocene to Pleistocene) basaltic lava flows. Many stream valleys were dammed by lava flows and subsequently filled in with clastic sediment to form broad fertile plains (Pine Valley, Grass Valley, Grassy Flat, Diamond Valley, and the unnamed area south of the town of Central). Large aprons of Pliocene (?) and Pleistocene boulder alluvium, formed mostly by debris flows, are located on the flanks of the Pine Valley intrusion and represent the only other major sedimentary deposits.



AGE	FORMATION	MEMBER	Thickness (m)	LITHOLOGY
HOLOCENE-PLEISTOCENE	Surficial deposits-alluvial, mass wasting		0-46	
	Basaltic lava flows		0-61	
	Boulder deposits		0-61	
PLEISTOCENE-PLIOCENE	Eight Mile Dacite		0-152	
	Old alluvial-fan deposits		0-91	
PLIOCENE	Older basaltic lava flows		0-61	
MIOCENE	Alluvial deposits of Spring Creek		0-91	
	Racer Canyon Tuff		0-91	

Figure 11. Generalized stratigraphic column of post-Iron Axis rocks.

STRUCTURES - EARLY MIOCENE (IRON AXIS INTRUSIVE TYPE)

The major deformation in the Pine Valley Mountains occurred during early Miocene magmatism related to the forcible intrusion of quartz monzonite, as it formed laccolithic structures. Most successive intrusions deformed earlier structures formed by adjacent intrusions, thus complicating yet aiding in deciphering age relations. The intrusions made room for themselves by forcibly deforming the country rock by: (1) upflexing, producing domed or arched roofs containing extensional faults and attenuated bedding, and (2) shouldering aside the confining rocks, producing overturned monoclinal flexures, thrust faults, reverse faults, and fault-propagation folds. These structures are directly associated with the processes of intrusion and are familiar features of many laccoliths. Less familiar to the emplacement of laccoliths are structures formed indirectly by detachment faulting of large slabs of roof rock from the growing flanks of the intrusions. These low-angle, detached fault masses constitute a large allochthonous complex of gravity slides (figure 12). Although these are secondary structural features, they owe their existence to the doming of cover strata as the laccoliths intruded and made room for themselves at depth. The recognition of the detached masses as gravity slides formed from the flanks of rapidly growing laccoliths greatly aided in defining the structural history of the intrusions.

Gravity Slide Complex

Within the study area are numerous large allochthonous masses (sheets or blocks) of fractured, brecciated, sheared, and attenuated Tertiary volcanic and sedimentary rocks resting on low-angle faults. Faulting placed both younger rocks over older and older rocks over younger. Some of these anomalous structural relations within the Pine Valley Mountains were first hinted at by Cook (1954, 1957), who recognized a similarity between Tertiary rock units that he defined as the Grass Valley and Atchinson Formations and the older rock units underlying them. Cook suggested that the different formations could be the same rock units brought together tectonically by thrust faulting. Cook (1960) later abandoned the two formations and mapped some of the rocks on the north side of Grass Valley and in Grassy Flat Canyon as allochthonous faulted “slide masses,” and suggested that they were emplaced by gravity sliding along low-angle normal faults off the up-arched roof of an intrusive body exposed 5.6 km south-southeast of Pinto. This location corresponds to Cook’s (1957) Paradise intrusion that is now interpreted to be an extrusive lava flow (upper lava member of the rocks of Paradise), and therefore the lava flow could not have been an area of structural relief necessary for sliding.

Allochthonous rock masses were earlier recognized in the adjacent Bull Valley Mountains (Blank, 1959) and Iron Springs District (Mackin, 1960), and interpreted as slide masses related to gravitational unroofing of nearby intrusions (Blank, 1959, 1993; Mackin, 1960; Rowley and others, 1989, 2001). Recently, allochthonous rock masses have also been recently recognized in the northeastern end of the Iron Axis in both the Basin and Range sector (Red Hills area), and on the western Colorado Plateau sector (Markagunt Plateau area). These features have been interpreted to result from general gravity-sliding events initiated by different mechanisms at different times (Maldonado and others, 1997; Davis, 1999; Hatfield and others, 2000). The emplacement of the masses have been variously related to plateau uplift (Sable and Anderson, 1985), fault scarp collapse (Anderson, 1985), tilting above a batholith (Anderson, 1993), or tilting by crustal extension and associated block faulting (Maldonado, 1995; Maldonado and others, 1997). These allochthonous rock masses were emplaced during the Miocene (Maldonado and others, 1997; Hatfield and others, 2000).

Mapping for this study shows that the allochthonous rock masses within the Pine Valley Mountains are more extensive and more numerous than previously thought (e.g., Cook, 1960). A detailed study of the allochthonous stratigraphy, when compared with surrounding autochthonous units, also indicates that the allochthonous masses have more than one source area. Comparison of these allochthonous masses with others in the nearby eastern Bull Valley Mountains and Iron Springs district also strongly suggests that they have a common origin as detached masses derived from the growing uplifted flanks of nearby intrusive domes and were emplaced by gravitational sliding (Blank and others, 1992; Hacker and others, 1996). At least six major sliding episodes in this part of the Iron Axis can be attributed to the rapid emplacement of the Pinto Peak, Bull Valley-Big Mountain, Iron Mountain, Stoddard Mountain, and Pine Valley intrusions (Hacker, 1998). Individually, each sliding event was immediately followed by volcanic eruptions from their respective intrusions (except for the Iron Mountain intrusion) that broke through the uplifted cover rocks and erupted ash flows and/or lava flows that partly or totally covered the slide masses. Individual slide masses from different intrusions can be distinguished from one another by their position beneath or upon these newly discovered volcanic rock units, as well as the incorporation of these units in subsequent slide masses.

Distribution and Composition of Slide Masses: The general structural relations and distribution of displaced slide masses within the entire slide complex (i.e., the areas of the Pine Valley Mountains, eastern Bull

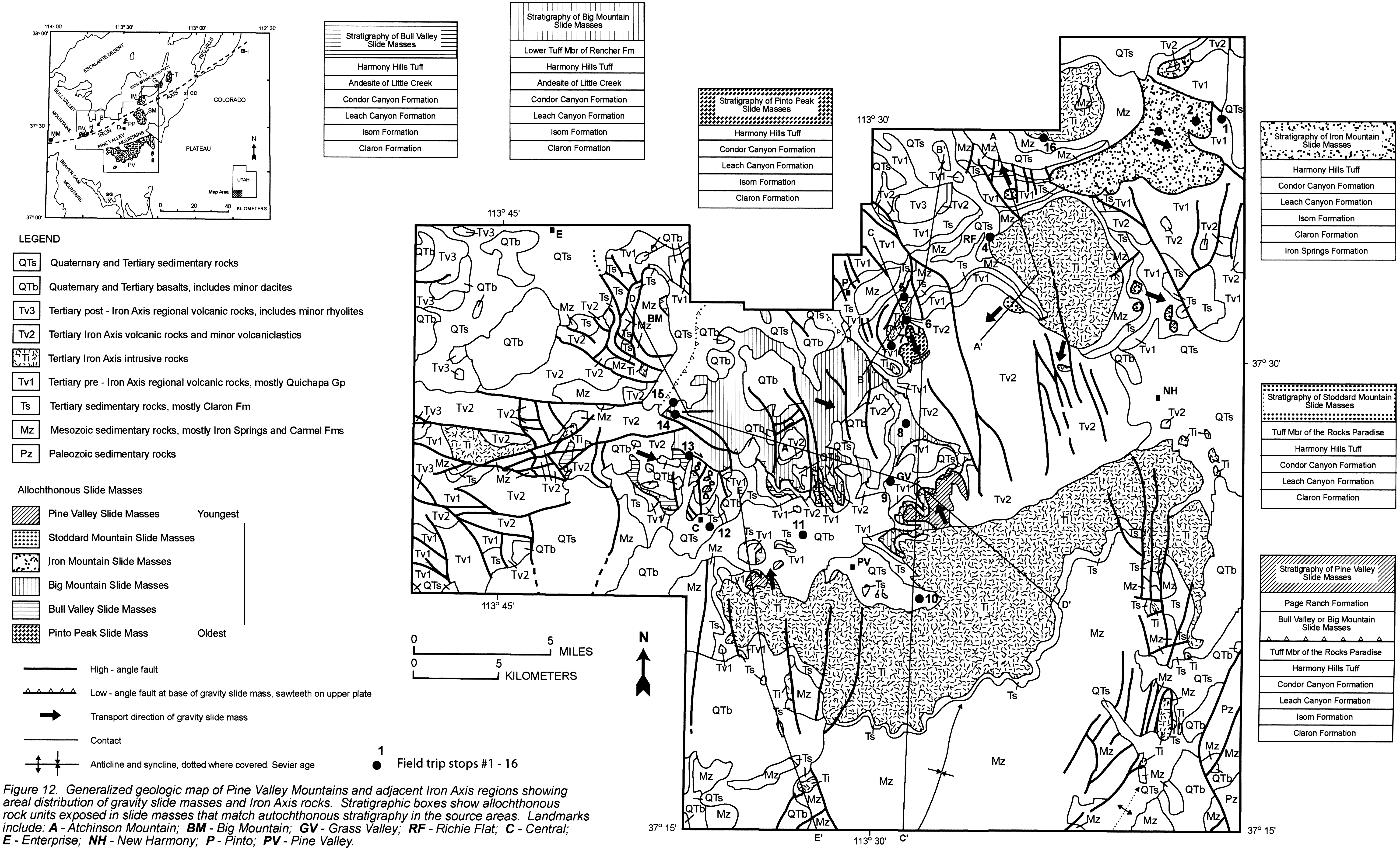


Figure 12. Generalized geologic map of Pine Valley Mountains and adjacent Iron Axis regions showing areal distribution of gravity slide masses and Iron Axis rocks. Stratigraphic boxes show allochthonous rock units exposed in slide masses that match autochthonous stratigraphy in the source areas. Landmarks include: ***A*** *- Atchinson Mountain;* ***BM*** *- Big Mountain;* ***GV*** *- Grass Valley;* ***RF*** *- Richie Flat;* ***C*** *- Central;* ***E*** *- Enterprise;* ***NH*** *- New Harmony;* ***P*** *- Pinto;* ***PV*** *- Pine Valley.*

Valley Mountains, and southwestern Iron Springs district) are displayed on a generalized geologic map (figure 12). Field relations strongly suggest that the present distribution of the scattered masses over the more than 600 km^2 area is not the result of subsequent erosion of a once continuous sheet. Instead, the distribution is the result of separate sliding events that displaced rock masses away from source areas occupied by intrusions. The slide structures are therefore grouped on figure 12 according to their inferred intrusive source. Arrows shown on the geologic map (figure 12) show the interpreted emplacement-direction of the individual slide masses. Opposing directions of transport and variable internal stratigraphy indicate the slide masses were not part of a single continuous slide sheet.

Displaced rocks that compose individual slide masses originated from nearby recognizable autochthonous formations and include volcanic and sedimentary rocks that range in age from Cretaceous (Iron Springs Formation) through early Miocene (Page Ranch Formation) (see figure 12). It should be noted that not every slide mass contains the same stratigraphy. This is due to: (1) the original lateral lithologic variations in the stratigraphic sequence at the different source areas, and (2) the timing of sliding events in relation to volcanic eruptions from different intrusions. Most slide masses can be distinguished from one another in the field by their lithology.

General Internal Structures of Slide Masses: The formations within the slide masses show attenuation and exhibit various internal structural complexities; however, normal internal stratigraphic succession is commonly maintained that allows individual rock units to be mapped. Typically, the formations exhibit pervasive internal fracturing and shattering but are well indurated. Some formations are brecciated and consist of pebble-to-boulder-sized, angular to subangular rock fragments with a crushed matrix of the same composition as the fragments. The brecciated zones are commonly matrix-poor, with the fragments commonly tightly packed in a jigsaw-puzzle mosaic separated by a cataclastically generated sand-to-granule-size matrix (figure 13). Most

Figure 13. Outcrop of allochthonous Harmony Hills Tuff showing rotated angular rock fragments in a comminuted matrix of the same composition as the fragments. In other areas where the matrix is sparse, the shattered rock resembles a three-dimensional jigsaw puzzle. The rocks shown in this photo are in Grass Valley and are part of the Big Mountain slide mass.

fragments have moved slightly relative to their neighbors, while others show some rotation. Rocks from adjacent formations are usually not mixed, but locally are chaotically juxtaposed along close-spaced shear domains. Omission or smearing out of stratigraphic units takes place along low-angle shear zones or bedding-plane faults. In fact, many of the formation contacts are actually fault contacts, but are mapped with a formation contact symbol due to the scale of mapping. Mechanically, the character of the internal deformation varies with rock lithology. The softer, moderately welded ash-flow tuffs (e.g., lower tuff member of the Rencher Formation and tuff member of the rock of Paradise) deformed along sheared zones as much as 3 cm thick containing pulverized rock flour material (cataclasite) with the consistency of clay, formed by the mechanical breakdown by crushing and grinding of the ash-flow tuff. In contrast, the more competent highly welded ash-flow tuffs (e.g., Leach Canyon Formation, Bauers Tuff Member of Condor Canyon Formation, and locally the Harmony Hills Tuff) deformed along brittle intersecting or anastomosing sets of shear fractures, which result in brecciation of the rocks.

In addition to internal deformation of the rock units, many structurally complex areas within the slide masses that contain extensively faulted and folded strata. Faulting occurs along: (1) local tear faults oriented parallel to transport direction, and (2) high-to low-angle normal and reverse faults oriented mostly perpendicular to transport direction.

Nature of Basal Fault Contacts: Overall, the slide masses closely resemble remnants of “erosional thrust sheets” in form. All deformation is confined to the slide masses themselves and abruptly terminates downward at low-angle, basal bounding detachment faults (referred to here as gravity-slide faults). The basal gravity-slide faults are a composite of four types of fault surfaces. They include: (1) a subhorizontal “bedding fault” (decollement) within sedimentary rocks of the Tertiary Claron or Upper Cretaceous Iron Springs Formation, in which younger rocks overlie older rocks, (2) subvertical “flanking faults,” which are lateral bounding tear faults with strike-slip movement, (3) a transgressive “ramp fault” that cuts upward across bedding, and (4) a subhorizontal “land surface fault,” which is a fault between the slide mass and the pre-existing land surface of that time, in which older rocks overlie younger rocks. All four fault components are not necessarily preserved beneath every slide mass due to differential erosion, especially at the proximal ends of the slide masses where the bedding and ramp structures were mostly located at higher elevations on the intrusive domes and therefore were more susceptible to erosion, or they were destroyed by subsequent extrusion of magma. The Big Mountain slide is the most complete structure that retains a set of all four bounding faults, as illustrated diagrammatically in figure 14. The contacts between individual slide masses and the underlying autochthonous rocks appear to be smooth and sharp. However, most contacts are covered and can only be located within the nearest meter or two.

Age of Sliding Events: The timing of all sliding is tightly constrained between 22 and 20.5 Ma. The Harmony Hills Tuff, a key pre-Iron Axis unit involved in all of the slide masses, yields a 22.03 ± 0.15 Ma ^{40}Ar/^{39}Ar plateau age (Cornell and others, 2001). The Pine Valley Latite, which overlies the youngest slide mass, yields a 20.54 ± 0.07 Ma ^{40}Ar/^{39}Ar plateau age (L.W. Snee, U.S. Geological Survey, written communication, 1995; Hacker and others, 1996). Gravity sliding was initiated abruptly after deposition of the Harmony Hills Tuff, as the oldest slide mass (the Pinto Peak slide) is overlain by the 21.93 ± 0.07 Ma rocks of Paradise (lower lava flow member; L.W. Snee, U.S. Geological Survey, written communication, 1995; Hacker and others, 1996). Importantly, this 1.5 m.y. time interval of gravity sliding corresponds to the age of Iron Axis intrusions within the Iron Axis region.

Speed of Emplacement: The velocity of the gravity slides is a matter of speculation, but has special implications in the origin of the intrusions. The speed is estimated to be rapid and most likely catastrophic. The absence of field evidence for continued thrusting and gouging and the internal structure (e.g., brecciation, extensional faulting) suggest movement by a body force (i.e. gravity). Very rapid to catastrophic movement of the gravity slides seems necessary to explain the following observations:

(1) There are no sedimentary deposits immediately beneath the slide masses indicative of erosion along an elevated area (i.e., laccolithic domes in this case) prior to sliding.

(2) No erosion material exists between the slide and the overlying volcanic rocks that erupted from the same source area following sliding. This indicates that volcanism was synchronous with or immediately followed sliding. Most of the volcanic material overlying the slides consists of ash-flow tuffs erupted from the laccoliths. The tuffs indicate violent eruptions and catastrophic emplacement.

(3) Extremely thin but stratigraphically preserved rock units that traveled at least 12 km over the former land surface.

(4) Internal brecciation and shattering of rock units within slide masses precludes a push from the rear and requires a “one-shot” emplacement mechanism, especially for such thin layers of rock.

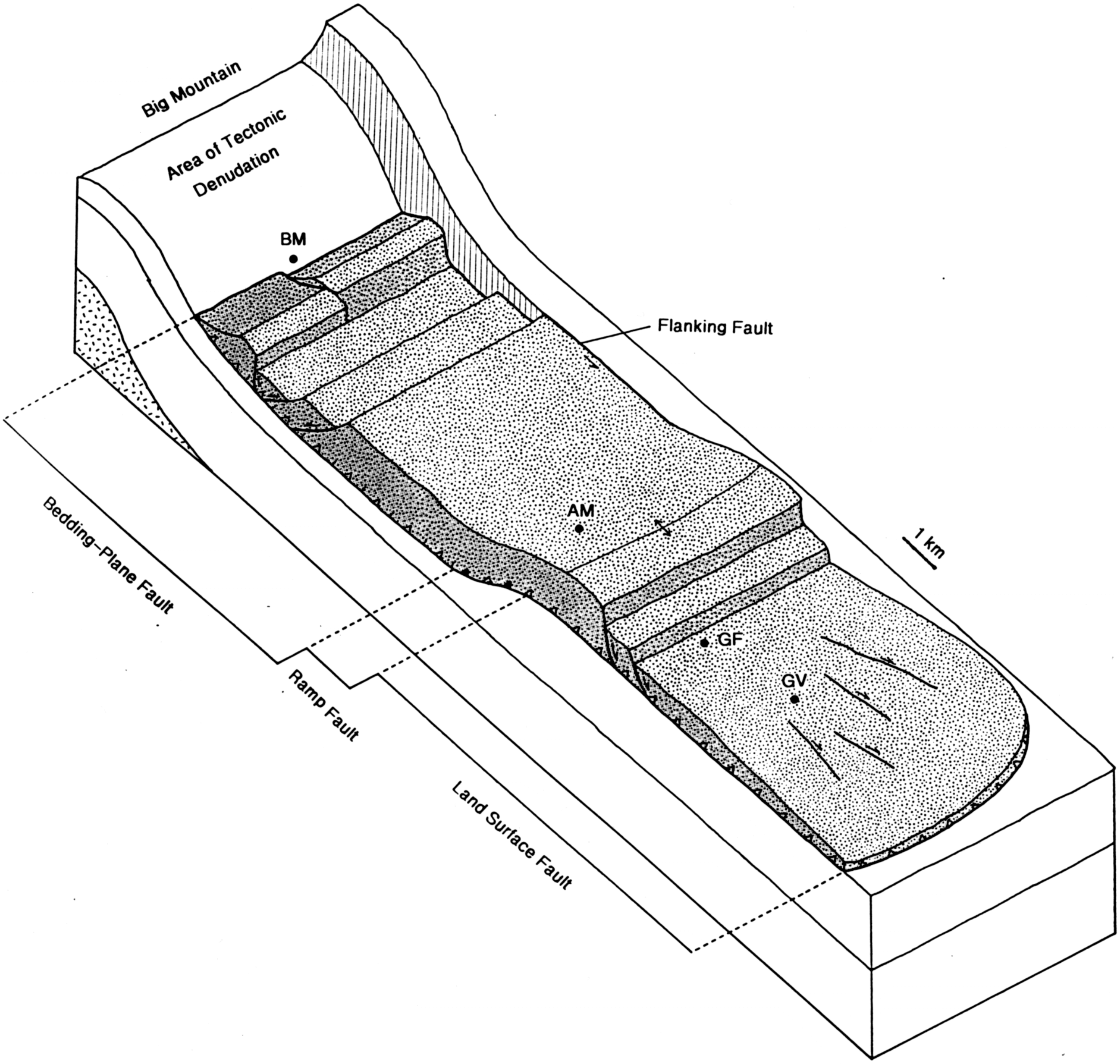


Figure 14. Sketch block diagram of Big Mountain slide showing four types of bounding faults.

Structural Interpretation: From the preceding descriptions and review of cross sections it is difficult to avoid the interpretation that the allochthonous masses originated by downslope sliding of large bodies of detached strata pulled by the force of gravity. A necessary prerequisite for downslope gravity sliding is the creation of tectonic elevations. The timing of formation of the gravity slide complex corresponds very well to the time of emplacement of intrusions that formed laccolithic structures with elevated domal roofs. Therefore, the interpretation that detachment and displacement of large rock masses were related to the dynamics of formation of the intrusions seems clear.

However, as pointed out by J. Hoover Mackin (1960), the concept of gravity sliding from igneous intrusive domes is not well known due in part to: (1) only a small percentage of intrusions form topographic features steep enough to cause sliding, (2) many intrusions of pre-Tertiary age are deeply eroded so that any slides that might have developed have been removed, and (3) slides from intrusive domes have been interpreted as thrusts due directly to the intrusive room-making process. However, timing of deformation was also contemporaneous with the extrusion of large volumes of volcanic material from the intrusions, which, when thick enough, can independently create abnormal stress on adjacent strata. A brief examination, therefore, of possible alternative mechanisms related to the formation of the intrusions and their associated volcanics seems warranted.

The emplacement of older rocks upon younger rocks along low-angle surfaces usually brings to mind deformation by compressional forces that produce thrust faults. Grant (1991) suggested that such a compressional force during growth of the Pine Valley intrusion was responsible for the shouldering aside of large volumes of volcanic rock to create the flat floor within sedimentary units of the Claron Formation. However, Grant's interpretation is not supported by field data that shows that the Pine Valley intrusion formed by uplifting its cover rocks (see section on laccoliths below). Some compressive forces were generated during growth of the Pine Valley intrusion, which produced minor thrusting of the peripheral country rocks with as much as 600 m of displacement (Hacker, 1995,1998). These features, however, post-date the slide structures.

The largest and best-preserved slide in the study area, the Big Mountain slide, has many structural characteristics comparable to the Heart Mountain fault in northwest Wyoming. The Heart Mountain fault covers 3,400 km^2 and has transported a thin (~750 m) allochthon tens of kilometers over a nearly horizontal detachment surface (Pierce, 1973, 1991; Hague, 1985, 1990). The Heart Mountain fault is similar to the Big Mountain fault in that it evolved progressively from a bedding fault, to a ramp fault, and finally to a land-surface fault. It is also similar in being partially buried by volcanic rocks, but is dissimilar in that it contains a breakaway fault. Moreover, the upper plate of the Heart Mountain fault broke into numerous blocks that became separated during detachment and transported individually over the fault surface. The Big Mountain slide apparently moved as a single mass, (although it became internally broken). It also probably contained a breakaway-type of fault that either has been eroded or is covered by volcanic rocks.

Two models have been proposed for the development of the Heart Mountain fault: (1) a tectonic-denudation model (Pierce, 1973), and (2) a continuous allochthon model (Hauge, 1985). The tectonic denudation model suggests that the discrete blocks of Paleozoic rocks moved at catastrophic rates along the detachment surface due to gravity and aided by earthquake oscillations. The slide area was blanketed immediately after faulting by catastrophic volcanism. In the continuous allochthon model, a continuous plate of thick volcanic and Paleozoic rocks moved over millions of years by the slow process of gravity spreading. In the Big Mountain slide, there are several lines of evidence that suggest the structure was not produced by gravity spreading or any compression from the rear: (1) the internal deformation of the displaced masses show extensive attenuation by internal extensional faulting and brecciation rather than compressional thickening, (2) the overlying volcanic rocks are very thin and show no signs of being deformed, (3) the preserved denudation surfaces at the proximal end of the slide are covered with volcanic rocks that are not deformed, and 4) volcanic rocks did not cover the entire structure and therefore spreading would be only from the proximal end and form compressional structures at the slide's toe. However, the toe is thin (<100 m) and could not have transmitted compressional stresses for any acceptable distance. Sliding in the Pine Valley Mountains is interpreted to be similar to the tectonic-denudation model of catastrophic emplacement of the allochthonous masses prior to burial by volcanic rocks. Catastrophic emplacement best fits the observation of a thin allochthonous sheet over the former land-surface fault segment that could not have sustained compression from the rear. Since the Big Mountain slide, as well as most slides of the study area, is overlain by explosive volcanic deposits, earthquake oscillations from the eruptions could also have aided in their emplacement as suggested by Pierce (1973) for the Heart Mountain fault.

Laccoliths

The intrusions within the Pine Valley Mountains area are variably eroded. Some of the intrusions are completely concealed beneath their cover so that the shape (anticlines or domes) reflects the general shape of the underlying intrusion (e.g., Pinto, Ritchie Flat, and Grass Valley areas). In other intrusions the cover has been partly removed so that the upper crests and flanks are exposed (e.g., The Dairy, Pinto Peak, Stoddard Mountain). Only the Pine Valley and Iron Peak (at the northeast end of the Iron Axis) intrusions have been eroded so that the floor is exposed. Between the intrusions are intervening synclines or basins formed of sedimentary and volcanic rocks and gravity slide masses. Local compression between the Stoddard Mountain and Iron Mountain intrusions formed an out-of-syncline thrust. The nature of the igneous intrusions in the Iron Axis province has been a controversial issue for some time. Workers in the Iron Springs district have suggested that the exposed intrusive bodies are parts of either laccoliths (Leith and Harder, 1908; Mackin, 1947, 1954, 1960; Blank and Mackin, 1967) or stocks (Butler, 1920; Ratte, 1963; Bullock, 1973). The present erosional level of the intrusions makes it difficult to distinguish stocks from laccoliths. Recent exploratory seismic reflection and exploratory drilling data clearly show that the Three Peaks intrusive body is a laccolith (Van Kooten, 1988). The laccolith floor was penetrated at a depth of 1,498 m in the ARCO Three Peaks No. 1 well while exploring the petroleum potential of the anticlinal structure located below the laccolith. The laccolith has a thickness of 788 m at the well location, and is both overlain and floored by units of the Jurassic Carmel Formation. Drilling and seismic data do not indicate the presence of additional stacked laccoliths below the Three Peaks laccolith, or the presence of a central stock that would have acted as a central feeder system for satellite laccoliths as envisioned by Hunt (1953) for the laccolithic centers in the Henry Mountains of the Colorado Plateau.

The laccolith interpretation best fits the observed field relations of all of the intrusions within the Pine Valley Mountains. Cross-sections show that most intrusions have a somewhat typical plano-convex shape (Gilbert, 1877), but are not circular when viewed in plan. The Pine Valley intrusion is more tabular in shape and resembles a large sill, although its cover has been entirely removed by erosion and therefore the curvature of its roof at the time of emplacement is unknown. Attempts to arbitrarily distinguish laccoliths from sills on the basis of diameter/thickness ratios (e.g., Billings, 1972) have not been followed in the literature due to the continuous transition between the two and differential erosion. A recent compilation of laccoliths by Corry (1988) led him to suggest that laccoliths be distinguished from a sill when the thickness is $\geq$ 30 m. Corry (1988) also referred to laccoliths in the generic sense as intrusions of any final form that domed the country rock above or below them through a forcible intrusive and thickening process. Later publications incorporate this usage (e.g., Henry and others, 1997) and is followed here.

Pine Valley Intrusion - Laccolith or Lava Flow?

The Pine Valley intrusion is the largest and most controversial intrusion within the Iron Axis region. The exposed body has a preserved maximum thickness of 900 m and a maximum horizontal dimension of over 32 km in a northeast-southwest direction. The original extent is unknown, but must have been much larger than its present areal extent of over 240 km^2 based on erosional remnants found on the south side where erosion was greatest. Its northern and western limits are still mostly preserved. Its eastern limit is also unknown due to burial beneath younger sediments (Grant, 1991). The igneous body exhibits an overall irregular shape in plan view as shown by separate lobes that extend from the main mass (e.g., north side of Pine Valley).

Early workers (Dobbin, 1939; Gardner, 1941) believed the igneous body was the product of lava flows, based on their observations on the south side of the mass where it has a layered appearance and on the aphanitic groundmass of the porphyry. Subsequently Cook (1954, 1957, 1960) made the first comprehensive study of the entire igneous body and interpreted the mass to be part of possibly the largest laccolith in the world. Cook based his intrusive interpretation on field evidence that included: (1) the intrusive nature of the contact on the north side of the mass, and (2) the lack of evidence that showed that the layering was due to separate volcanic lava flows. Cook's interpretation was that a single shot of magma intruded beneath at least 900 m of cover rocks. The petrographically identical Pine Valley Latite was interpreted by Cook to be pre-intrusive and part of the cover strata that was deformed by intrusive faulting. In a more detailed study of the nature of the layering, Mattison (1972) came to the same conclusion as Cook that the layering was not due to separate lava flows. However, Mattison believed the mass was extrusive in origin, probably formed as an "extrusion ridge," and not as a laccolith. The lack of any roof rocks remaining on top of the mass and the glassy groundmass of a great portion of the mass were cited as some of the reasons for rejecting Cook's intrusive concept. Mattison made no mention of the relationship between the extrusive Pine Valley Latite and intrusive rocks. Mattison's study was based only on transects through the igneous body and did not include field studies of the surrounding areas. Grant (1991) later adopted a similar interpretation as Mattison after a field study of the eastern side of the Pine Valley Mountains conducted during mapping of the New Harmony quadrangle. Grant interpreted the upper 305 m of the mass to be extrusive and made up of a lava flow dome that was subsequently intruded from below. Grant mapped the upper purple layer as the Pine Valley Latite and thus extended the Pine Valley Latite over the entire intrusion. Grant explained the missing pre-Iron Axis volcanic rocks above the Claron Formation by shouldering aside the rocks during the initial lava flow dome formation, allowing the Pine Valley Latite to rest temporarily on the Claron Formation, before magma later intruded between the Claron and the latite. Grant suggested that the folded rocks in Comanche Canyon were part of the bulldozed rocks and that the rest were buried below parts of the dome.

A variation of Cook's laccolith interpretation best fits the overall field relations, with the addition of widespread eruptive phases as well. In short, field data show that magma was intruded first beneath a thin cover (<250 m), and uplifted its roof to form a laccolithic structure that broke through to the surface and extruded lava over the country rock. This lava forms part of the Pine Valley Latite (i.e. the Timber Mountain flow unit of the Pine Valley Latite). The upper layer of igneous rock, interpreted as extrusive by Grant, is envisioned here to be the upper glassy zone of the intrusion just below the cover rocks that have been removed. Thus, field relations from this study support Cook's interpretation that the igneous mass was an intrusive laccolith, although there is also evidence of extrusive activity resulting from the laccolith breaking through its roof and erupting latite lava.

Evidence of intrusion: The evidence of forcible intrusive activity is demonstrated in: 1) the lack of extrusive layering, (2) the intrusive character of the basal contact, (3) the intrusive character of the peripheral contact, and (4) the presence of gravity-slide structures.

(1) Field evidence from this study supports the conclusions of Cook (1957) and Mattison (1972) that the colored layers or zones that are so prominent in the south side of the intrusion are not the products of separate lava flows or even due to separate intrusions. The contact between the zones is gradational and irregular and shows no visible changes in phenocryst size from one layer to the next. The contact between the upper purple and middle white zones was examined closely for evidence of an intrusive contact, as envisioned by Grant (1991). This contact should be sharp or show signs of a basal vitrophyre in the purple zone, which is prominent everywhere at the base of the Timber Mountain flow member of the Pine Valley Latite. However, no physical difference was found at this contact to support two separate events.

(2) Along the south side of the intrusion the igneous mass rests upon units of the Claron Formation. The only exception is at Cedar Knoll, where the intrusion appears to rest on the Iron Springs Formation displaced along a high-angle fault that strikes northeast. North of the intrusion the Claron Formation is overlain by a thick section of pre-Iron Axis volcanic rocks. There is no evidence that the volcanics were removed by erosion to expose the Claron prior to emplacement of the intrusion. There are no sedimentary units found beneath the intrusion that are indicative of erosion of the volcanics. The contact between the Claron and intrusive rocks is mostly sharp to irregular and the layer below the contact is fractured and sheared and slightly baked or bleached. The basal glass clearly visible in outcrops becomes highly sheared near the peripheral areas of the intrusion and suggests continued movement of the mass.

(3) Compelling evidence of intrusive contacts are found on the preserved northern and western flanks of the intrusion where upturned to overturned strata are highly attenuated adjacent to intrusive rock. The overturned strata include rocks of the Claron Formation and units of the Quichapa Group, as well as parts of the Bull Valley and Big Mountain slide masses. North of Pine Valley, the intrusion compressionally deformed the flanking country rocks forming a small thrust fault and a fault-propagation-fold. The thrust fault dips 20 degrees toward and under the intrusion and has a maximum displacement of over 300 m. Country rocks adjacent to the intrusion are overturned indicating that thrusting occurred after the intrusion grew vertically by upturning its roof rocks.

Cook (1957) interpreted the intrusion's northern extent, from Water Canyon eastward to New Harmony, to be in intrusive fault contact with members of the volcanic sequence and the Pine Valley Latite. Detailed examination of the exposed contact did not reveal a high-angle fault contact surface. Instead, upturned and overturned units of the Claron and overlying volcanic units (but not the Pine Valley Latite) were observed. These host units are highly attenuated and could very easily be missed in a reconnaissance study such as Cook's (1954, 1957). At Water Canyon the intrusive and extrusive rocks are in contact with each other and, due to lack of fault indicators, are separated in mapping with a gradational intrusive-extrusive contact instead of the intrusive fault envisioned by Cook (1957). The Timber Mountain flow overlies the eastern Pine Valley gravity slide mass (which was contemporaneous with the intrusive rocks), so the extrusive rock appears to postdate the initial growth of the intrusion and was not intrusively faulted at this contact. Evidence can be found in a stream cut on the south side of the Bench (in Water Canyon), where intrusive rock appears to be in direct fault contact with the eastern Pine Valley slide mass, but upon close examination it is separated by a few meters of brecciated Claron and smeared-out Quichapa units. These rocks represent part of the upturned country rock sequence that is attenuated (by brecciation and shearing) vertically as the intrusion grew. The intrusive rock in contact with this "crush breccia" is a black glass that is also locally highly sheared. The crush breccia ends abruptly upward where the intrusive and extrusive rocks (i.e., Timber Mountain flow) come in contact with each other. No Claron or Quichapa units are between the two rocks that would be expected if the Timber Mountain flow pre-dated the intrusion. The extrusive rocks show more horizontal flow layering than the intrusive rocks, but are not upturned or faulted as observed by Cook (1957). Glass flow breccias appear to be more common closer to the intrusion, and in one location (at First Water), a red and black flow breccia overlies steeply dipping foliated intrusive rock. These relationships suggest that the intrusion pre-dated the Timber Mountain flow and was the source of some of the extrusive rocks.

(4) As just discussed in the "Gravity Slide Complex" section, the presence of allochthonous rock masses interpreted as gravity slides (i.e., the Pine Valley slides) demonstrates indirectly, yet compellingly, the need for an elevated area. The structural and stratigraphic indicators of slide masses on the north side of the intrusion, in the Grass Valley and Mahogany Creek area (Pine Valley slide system), show their source area to be to the south in the area now occupied by the Pine Valley igneous mass. The presence of gravity slides related to the Pine Valley intrusion is evidence against the igneous mass being extrusive in nature since extrusive lava does not have the strength to lift large areas of the landscape.

Evidence of extrusions: The Timber Mountain flow member of the Pine Valley Latite partially covers the gravity slide masses that originated from the elevated roof of the Pine Valley intrusion. This critical relationship indicates that the intrusion occurred before the latite flow formed, contrary to the ideas of Cook (1957) and Grant (1991), who both suggested the latite formed first and was part of the cover rocks that deformed during subsequent intrusive activity. However, the latite is not deformed where it is in contact with the intrusive rocks. The contact between the two rock units is hard to distinguish in the field because of the identical textures of the two units. An intrusive-extrusive contact is even hard to distinguish on air photos. If the intrusive rock formed first and was exposed by erosion at the time of latite emplacement, then a prominent contact should have formed as a result of the flowing lava butting up against the elevated intrusion as is found with the Stoddard Mountain intrusion. Instead, the contact in most places appears to be gradational as flow layered latite grades into massive layered latite and then into structureless intrusive rock. The latite flow layers strike parallel to the inferred contact but have various dips into or away from the intrusion. The varying dips reflect flow folding in the latite and are not deformational folds from later intrusive activity.

It is proposed here that the Timber Mountain flow originated from the intrusion and flowed northward onto country rocks and parts of the Rencher Peak flow that originated at a vent in the Rencher Peak area. It cannot be ruled out however, that the Timber Mountain flow also originated from a separate vent in the Timber Mountain area and flowed south and commingled with contemporaneously forming latite flows erupting from the intrusion.

Depth of emplacement: Field evidence shows that the Pine Valley intrusion intuded laterally within the sedimentary units of the Claron Formation. The Claron beneath the intrusion ranges in thickness from 120 to 150 m. Stratigraphic study of the Claron north of the intrusion indicates that the Claron is 180 m thick and that the intrusion was therefore emplaced into the upper part of the formation.

The maximum total thickness of the cover is estimated to be only 150 to 200 m at the time of intrusion. This includes ~40 m of upper Claron units, ~120 m of various volcanic units (see units listed in stratigraphic boxes of the Pine Valley slide masses in figure 12, which represents the detached parts of the roof), and ~20 m of previous slide masses (i.e. Big Mountain and Bull Valley slides), which did not cover the entire intrusion area. This maximum thickness is from the measured thicknesses of the rock sequence found just north of the intrusion. The thickness of the cover was most likely thinner on the south side due to southward depositional thinning of the volcanics and the absence of the Bull Valley or Big Mountain slide masses. The actual thickness to the south is unknown, but could have been as little as 100 m based on thinning of the Quichapa volcanics observed from north to south.

The Pine Valley intrusion is therefore estimated to have been emplaced at a maximum depth of <200 m based on field measurements of the thickness of the proposed cover rock stratigraphy, and it is inferred to have had a structural relief of at least 900 m based on the maximum remaining thickness of the intrusion. The estimated cover rock thickness (from approximately 150 to 200 m) is considerably less than the 1,219-1,523 m estimated by Cook (1957). The discrepancy is based mostly on the involvement of the Pine Valley Latite which Cook included as part of the cover but which field relations presented here show was formed shortly after or during the intrusion and therefore should not be included as part of the cover.

DISCUSSION

Relationship of Gravity Sliding to Volcanic Eruptions

The areas of structural and topographic relief that were necessary for the formation of gravity slide structures were formed by the forceful intrusion of quartz monzonite magma into sedimentary rocks and structurally uplifting their host rocks and forming laccoliths. Field evidence shows that volcanic activity closely followed gravity sliding events. This close relationship suggests that eruptions from the laccoliths could not take place until the initiation of gravity sliding. It is proposed that catastrophic gravity sliding initiated volcanism from growing laccoliths by sloughing off part of the roof and thus reducing the overburden pressure on the magma which led to violent magma frothing (vesiculation) and catastrophic eruptions of some of the intrusions. Most initial volcanic activity was of the violent ash-flow tuff type, giving support to gravity slides being the "initiators" of volcanism.

The initiation of volcanism by gravity sliding is not new to volcanic terrains. One of the best-known slides in recent times was the spectacular one initiating the devastating Mount St. Helens eruption of May 1980 (Lipman and Mullineaux, 1981). The growth of magma within the core of the stratovolcano caused gradual oversteepening of its flank, which collapsed and caused a huge landslide. The landslide unroofed the magma chamber, releasing the confining pressure, and thus initiating a simultaneous pyroclastic eruption.

Rate of Emplacement and Growth of the Laccoliths

It has been shown through experimental studies and observations that sill emplacement is rapid, on the order of 7 years or less in some areas (Corry, 1988), during which magma can migrate over several kilometers without crystallization taking place. The rate of emplacement of sills and the growth of laccoliths have been modeled experimentally and suggest rapid rates of emplacement ranging from 1 to 20 m/sec (e.g., Spera and others, 1985). It seems reasonable that the large Pine Valley laccolith also grew within a relatively short period of time, notwithstanding its large size. McDuffie and Marsh (1991) calculated the crystallization time for the Pine Valley intrusion to be approximately 2,500 years based on crystal setting rates. The emplacement and vertical growth of the Pine Valley laccolith had to be less than 2,500 years and could conceivably have been less than the general 100-year laccolith growth time frame suggested by Corry (1988). Certain field observations support the rapid rate of emplacement and growth time of the Pine Valley laccolith. These include:

(1) No evidence of erosion before gravity-slide emplacement. No sedimentary deposits were found below the slide that could be attributed to erosion of the dome prior to gravity sliding. Sedimentary rocks of the Page Ranch Formation are the only erosional products below and within some of the gravity slides, but they were formed by partial erosion of the Big Mountain slide, which formed an elevated area to the northwest. Fanglomerates of the Page Ranch Formation thicken to the northwest and overlap the edges of the Big Mountain slide mass and thin to the southeast toward the Pine Valley Mountains. Therefore the Page Ranch Formation was not derived from the initial Pine Valley dome.

(2) No evidence of erosion after gravity sliding and prior to eruption. No sedimentary deposits were found between the Pine Valley gravity slide masses and the overlying Pine Valley Latite that would indicate a pause in laccolith growth prior to its eruption.

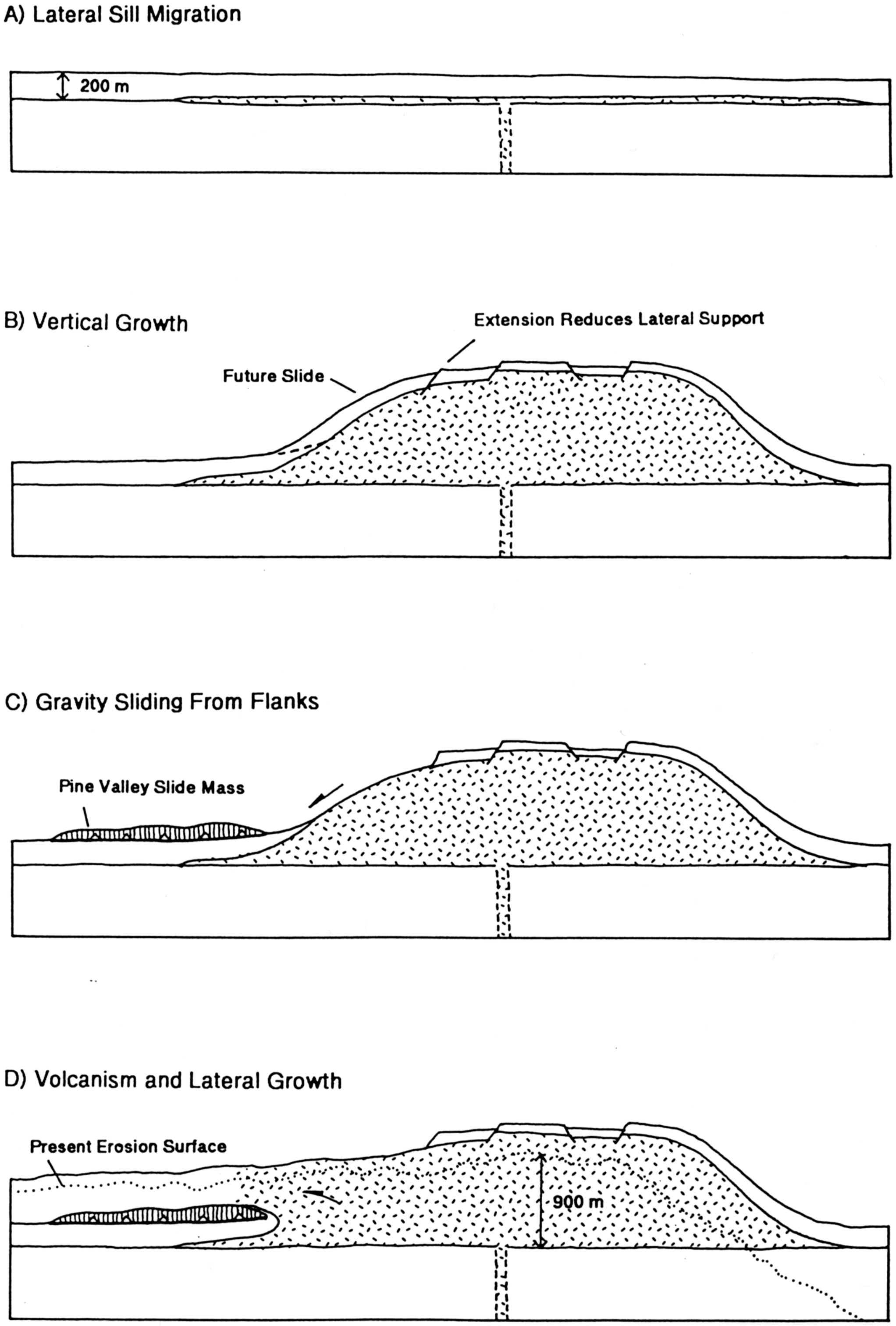
(3) The porphyry. The glassy nature of the quartz monzonite porphyry and its overall thickness indicates rapid intrusion of magma. Otherwise the mass would cool too quickly and show brittle deformation during any subsequent internal pulses.

(4) The thin cover. The extremely thin cover rock (<250 m thick) would allow more rapid cooling of the porphyry by not supplying a sufficient "blanket" against rapid heat loss. Therefore, the magma intruded rapidly to avoid rapid congealing.

Genesis of the Iron Axis Laccoliths

The field observations presented here have major implications on the intrusive evolution of the laccoliths of the Iron Axis. Based on the new field evidence pertaining to the structure of the large Pine Valley laccolith, a new model of emplacement and growth of the intrusion is presented below (figure 15). This model can be used to view the growth of the other Iron Axis laccoliths. The model is broken down into sequential stages of growth, culminating in the eruption to the surface of the Pine Valley latite lava. The progression of stages is viewed here as a continuum as suggested by Corry (1988) for the genesis of most laccoliths in general.

Stage 1 -- Initial Lateral Sill Emplacement: The magma that formed the laccolith was a half-crystallized mush at the time of emplacement (as evidenced by the large percentage of phenocysts in the porphyritic rock) and spread laterally as a thin sill within the sedimentary strata of the Claron Formation at a depth of



A) Lateral Sill Migration

B) Vertical Growth

C) Gravity Sliding From Flanks

D) Volcanism and Lateral Growth

Figure 15. Proposed model of Pine Valley intrusive growth based on field evidence. Growth is envisioned to be a continuum from one stage to the next. (A) Initial lateral migration of the sill to its fullest extent, followed by (B) vertical growth of laccolith, (C) gravity sliding of flanks, and (D) lateral growth by overturning the remaining peripheral flanks and extruding onto surface.

approximately 200 m (figure 15A). The actual thickness of the sill is unknown but Corry (1988) suggests that the sills (or protolaccoliths) in most laccoliths were less than 30 m thick regardless of the overall diameter of the laccolith. The magma is assumed to have spread to its full length as a sill prior to vertical growth of the laccolith (based on studies by Corry, 1988), and was fed from below by a feeder dike system. Although feeder dikes are not exposed beneath the Pine Valley laccolith (even though its floor is exposed over its entire southern flank), a dike system with a northeast-southwest trend would explain the anomalous length of the laccolith along this line. Also, the natural cross-section of its western side shows that the lowest elevation of the intrusive floor occurs in the middle of the laccolith where the Claron is thinnest. This is the assumed location of the dike system as the intrusion steps up through the Claron toward the intrusive periphery. Feeder dikes are present below the Iron Peak intrusion (figure 3), which is the only other laccolith in the Iron Axis group with an exposed floor. The Iron Peak intrusion also intruded into the Claron Formation and erosion has exposed a tightly clustered swarm of feeder dikes that average approximately 2 m thick and cut up through the nearly horizontal Claron beds at an angle of 40 to 90 degrees (Spurney, 1984). Evidence of lateral spreading as a sill is lacking due to the absence of shattered rock structures that should form in the country rock if the magma bulldozed its way through them at its full vertical thickness. Spreading as a sill explains the pattern of intrusion for the Dairy and Stoddard Mountain laccoliths. Both intrusions apparently emanated from a common conduit north of Pinto Peak (most likely the same conduit that fed the Pinto Peak intrusion). Finding the area to the south blocked by the already congealed Pinto Peak intrusion, the magma migrated to the east (for Stoddard Mountain intrusion) and west (for the Dairy intrusion). The thick lava flows from the Pinto Peak intrusion provided greater lithostatic pressure and also acted as a barrier to sill migration. Therefore, the sills migrated east and west (Richie Flat and Pinto anticlines) before migrating to the south to their fullest extent. By analogy, the Pine Valley laccolith also began as a sill and a remnant sill-like structure is interpreted to connect with the intrusion beneath Grass Valley (see cross section in figure 21).

Stage 2 -- Vertical Growth and Roof Deformation: Following emplacement of the sill to its full lateral extent, bending of the entire 200 m of overburden began as magma was continually added to the now vertically thickening laccolith (figure 15B). As the intrusion inflated, the overlying host rock was gently rotated and arched into doubly hinged flexures around the periphery, but was probably flat-roofed over most of its center due to its large lateral extent. Due to the shallow emplacement (thin overburden), extension over the upflexed area was accommodated by brittle fracturing and high-angle normal faulting of the roof. Although the peripheral flexures of the Pine Valley intrusion are only partially preserved and the crestal extensional faults have presumably been eroded, the described features are consistent with those found preserved in the crest of Dairy and Stoddard Mountain intrusion east of Pinto. The shape is consistent with experimental results of the bending of a circular plate driven upward by a thickening laccolith (Jackson and Pollard, 1988).

This stage of growth is in contrast to the concept of Cook (1957), who envisioned the periphery of the laccolith to be marked by high-angle intrusion faults similar to a bysmalith. These peripheral faults would occur if initial failure of the roof was by shearing at the periphery as suggested by Corry (1988) for what he termed punched laccoliths (bysmalith type). However, the structural geometry of this interpretation does not fit the requirements needed for gravity sliding and extrusive activity described in Stage 3.

Stage 3 -- Gravity Sliding from Peripheral Flexures: As the intrusion continued its vertical growth, the limb of the overlying peripheral flexure steepened as the hinges tightened. The extensional faulting at the hinge crest reduced the lateral support of the limb on an otherwise already steepened and unstable slope. The resulting slabs of roof rock detached within the sedimentary beds of the Claron Formation and slid intact onto the former land surface below (figure 15C). Detachment of large slabs of the roof rock greatly reduced the effective thickness of the peripheral overburden to only about 20 m of Claron strata on the remaining part of the flexure.

Stage 4 -- Vertical and Lateral Thickening and Magma Extrusion: The sudden loss of peripheral overburden greatly reduced the lithostatic pressure that was essentially holding the roof “down” in advance of the upward loading forces applied by the thickening magma. The now thin Claron units of the peripheral flexure left behind by the gravity slides were not strong enough to hold back the magma, which continued to undergo vertical thickening in conjunction with lateral growing. The magma was able to overturn the remaining peripheral roof rocks and extrude onto the surface, burying the previous gravity slide masses with lava flows (figure 15D). Subsequent removal of the roof by erosion left the present erosional profile shown in figure 15D. In areas where gravity sliding did not take place, the intrusive porphyry exerted enough lateral compression to form small displacement, low-angle thrust faults, or completely overturned the adjacent strata and broke out onto the former land surface. At this stage, the loss of lateral support resulted in total breakdown of the flanking strata by thrust faulting, flexure overturning, and compartmental high-angle faulting due to lateral growth. In the other intrusions that vented more violent ash-flows prior to lava eruptions (e.g., Pinto Peak, Stoddard Mountain, and Bull Valley intrusions) the sudden release of overburden by gravity sliding most likely resulted in immediate frothing of the magma due to massive pressure release, as with the 1980 Mount St. Helens eruption (Lipman and Mullineaux, 1981).

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FIELD TRIP ROAD LOG

Mileage		Description
Increment	Cumulative	
0.0	0.0	Mileage for this road log begins at the traffic light at the intersection of 200 North and Main Street in Cedar City, Utah. All stops during the trip are shown on figures 1 and 12. Proceed west from this intersection on SR-56 (200 North).
0.9	0.9	Overpass of I-15. Continue west on SR-56.
4.5	5.4	The “Y” intersection where SR-56 bends to the left and a paved road that goes to Iron Springs and Desert Mound intersects on the right. Continue west on SR-56. The Three Peaks intrusion with visible mine dumps on its flanks can be seen to the north (at right) at 3:00 to 4:00. The Three Peaks, Granite Mountain, and Iron Mountain intrusions make up the Iron Springs mining district, which contains the largest iron deposits in the western United States (Mackin, 1947, 1960; Mackin and Ingerson, 1960; Blank and Mackin, 1967; Mackin and others, 1976; Mackin and Rowley, 1976; Rowley and Barker, 1978; Blank and others, 1992; Barker, 1995; Rowley and others, 2001). These three laccoliths consist of porphyritic quartz monzonite emplaced concordantly at the base of, or within the middle of, the Middle Jurassic Temple Cap Formation that was displaced along one or more Late Cretaceous to early Tertiary Sevier thrust faults. The iron deposits consist of huge hematite and magnetite replacement ore bodies in the Homestake Limestone Member of the Jurassic Carmel Formation, overlying the thin Temple Cap Formation (Mackin 1947, 1968; Barker 1995). The ore forming fluids resulted from deuteric breakdown of ferromagnesian minerals in the outer parts of the crystallizing porphyritic mush of the intrusions (Mackin and Ingerson, 1960; Barker, 1995). Floors of the intrusions are not exposed, but an exploratory petroleum well was drilled on the southwest side of the Three Peaks body and intersected the bottom contact of the pluton at 1,500 m depth, proving that the intrusion is a laccolith (Van Kooten, 1988). Additional seismic reflection profiles show that the buried part of this intrusion connects with Granite Mountain and together they have a nearly planar floor and an irregular but concordant roof. The near hills on the right (2:00) are the Eightmile Hills made up of Tertiary ash-flow tuffs dipping southeast off the flank of Granite Mountain intrusion, which is partially visible above the hills. Large abandoned open-pit iron mines surround Granite Mountain.
2.9	8.3	Low area on left is Quichapa Lake, normally a dry playa on the valley floor in Holocene and Pleistocene valley bottom alluvial deposits (Mackin and others, 1976). Harmony Mountains (10:00 to 12:00) consist mostly of faulted Tertiary ash-flow tuffs. Swett Hills

		are visible at 1:00 to 2:00 and consist of east-dipping Tertiary ash-flow tuffs as found in Eightmile Hills to the right.
3.2	11.5	Crest of low hill formed in older, poorly consolidated Miocene to Pleistocene fanglomerates.
0.7	12.2	Enter Leach Canyon with outcrop of east-dipping Harmony Hills Tuff on right.
0.1	12.3	Bureau of Land Management sign on right.
0.1	12.4	Bauers Tuff Member of Condor Canyon Formation on right.
0.1	12.5	**STOP 1.** LEACH CANYON – EXAMINATION OF AUTOCHTHONOUS ASH-FLOW TUFFS. Pull off to the right to view three regional Tertiary ash-flow tuffs belonging to the Quichapa Group, on the north side of road. The first unit is the pink, crystal-poor Leach Canyon Formation (24 Ma) located where we park. Walk up the hill to the northeast to the overlying crystal-poor Bauers Tuff Member (22.8 Ma) of the Condor Canyon Formation. The contact is marked by a black basal glass zone of the Bauers Tuff that is overlain by a stony red devitrified zone, then an upper gray vapor-phase zone. Above the Bauers is the brown and tan, crystal-rich Harmony Hills Tuff (22.5 Ma). These unfractured ash-flow tuffs represent autochthonous rocks tilted eastward by the Iron Mountain Intrusion to the west. Return to vehicles and continue west on SR-56.
0.1	12.6	Highly fractured Leach Canyon Formation on the right. We just crossed over the frontal fault of the Iron Mountain slide. From here to mile 18.9 we will be traveling over allochthonous rocks of the Iron Mountain slide.
0.4	13.0	**STOP 2.** LEACH CANYON – EXAMINATION OF ALLOCHTHONOUS ASH–FLOW TUFFS. Pull off to the right and park to view highly fractured, steeply east-dipping, ash-flow tuffs above a low-angle detachment of the Iron Mountain slide. The roadcut on the north side of the road contains the light-pink Leach Canyon Formation on the right (east), underlain by purple and brown ash-flow tuffs of the Isom Formation (27-26 Ma) to the left (west). Sedimentary rocks of the Eocene to Oligocene Claron Formation farther to the west underlie the Isom Formation. These allochthonous rocks were highly fractured and steeply tilted during sliding eastward from Iron Mountain, but the stratigraphic order of the units is preserved. Return to vehicles and continue west on SR-56.
0.5	13.5	Roadcut on right of highly fractured red and yellow allochthonous Claron Formation within the Iron Mountain slide.
0.6	14.1	Old Woolsey Ranch and Duncan Creek on left (south) and Mount Claron on right (north). Mount Claron is the type locality of the Claron Formation and is made up of faulted and fractured red and white limestone, sandstone, and mudstone above red and tan sandstone and mudstone of the Upper Cretaceous Iron Springs Formation exposed at the base of the hill. The basal low-angle gravity slide plane formed in the incompetent rocks of the Iron Springs Formation. To the left (south), the large hills on the near horizon (Flat Top Mountain) consist of allochthonous Quichapa rocks and represent the farthest reaching (about 6 km) of the slides from the Iron Mountain intrusion (Rowley and others, 2001).
0.8	14.9	**STOP 3.** ALLOCHTHONOUS CLARON ROCKS. Pull off to the left to view highly fractured and pulverized outcrop of red Claron Formation on north side of road. These steeply dipping allochthonous rocks, along with others viewed from this vantage point, are formed of resistant narrow ribs of Claron beds that form an arcuate pattern around the Iron Mountain intrusion. Beyond Mount Claron (now behind us at 5:00 to 6:00) is a view of the Swett Hills bounded by a south-facing cliff. At the base of the high cliff is the

		Woolsey Ranch fault, a northwest-striking, left-lateral tear fault (bounding slide fault) that offsets to the east the allochthonous rocks that we have been driving on relative to the autochthonous rocks of the Swett Hills. Overall, the Iron Mountain slide covered an area of at least 30 km^2 (10 mi^2) (Rowley and others, 2001). Return to vehicles and continue west on SR-56.
1.2	16.1	Diamond Z Ranch driveway on left. Ribs of resistant Claron beds to the left and right of the road. Duncan Mountain to the south (at 10:00) consists of allochthonous Quichapa rocks. Continue on SR-56.
1.0	17.1	Comstock Road (gravel) on the right leads to abandoned Comstock Mine on the northeast flank of Iron Mountain. Mount Stoddard of the Stoddard Mountain intrusion at 10:00. Continue on SR-56.
0.3	17.4	Junction of SR-56 and road to Pinto (Dixie National Forest road 009) on left. **Turn left onto gravel surfaced Pinto road and proceed southwest.**
0.4	17.8	Highly fractured Claron in hills to right and left.
1.1	18.9	Cross over approximate western edge of Iron Mountain slide complex. On the left are Claron and Iron Springs rocks that are overturned to the north by the Stoddard Mountain intrusion, which forms the high gray cliffs directly to the south (left). The laccolith intruded into the Iron Springs Formation at a shallower level than the other intrusions of the Iron Springs mining district and did not form any strata-bound iron deposits.
0.8	19.7	Near-vertical units of Claron on right (north) and Iron Springs on left (south). Quartz monzonite of the Stoddard Mountain intrusion exposed in cliff to the south. Hill to the north consists of Quichapa rocks that have been thrust to the north over the Harmony Hills Tuff by the shouldering action of the Stoddard Mountain intrusion.
1.0	20.7	Hill to the right is capped by a small slide mass from the Stoddard Mountain intrusion (see cross section in figure16) composed of Claron, Leach Canyon, Bauers, and Harmony Hills units transported northward. The mass rests on the former land surface consisting of Harmony Hills Tuff and a local andesitic lava flow (local volcanic rocks of Hacker, 1998) derived from fissures and deposited above the Harmony Hills before Iron Axis magmatic activity. Iron Springs country rocks on the left, adjacent to Stoddard Mountain intrusion.
1.2	21.9	**STOP 4.** STODDARD MOUNTAIN INTRUSION. Pull off to right and view roadcut in Stoddard Mountain intrusion on south side of road. The intrusion consists of two recognizable zones of quartz monzonite porphyry. The outermost zone is a 20 to 60 m thick chilled margin known as the "peripheral shell phase" in the intrusions of the Iron Springs district (Mackin and Rowley, 1976; Mackin and others, 1976). This zone, exposed here in the roadcut, consists of resistant, light-gray, unaltered porphyritic rock with medium-to coarse-grained phenocrysts set in an aphanitic groundmass. Note the fresh appearance of the plagioclase, biotite, and ferromagnesian phenocrysts of this peripheral phase. The inner zone, which makes up the majority of the intrusion, consists of an "interior phase" (Mackin and Rowley, 1976) of crumbly, light-grayish-green and pink, deuterically altered porphyritic rock. McKee and others (1997) determined a K-Ar date on biotite of 21.5 ± 0.9 Ma from peripheral shell samples taken from this roadcut. Return to vehicles and continue southwest on Pinto road, crossing Little Pinto Creek.
0.1	22.0	Historic Page Ranch on right. Junction with gravel road on left (Dixie National Forest road 029) that leads southeast to New Harmony. Continue southwest on Pinto road.

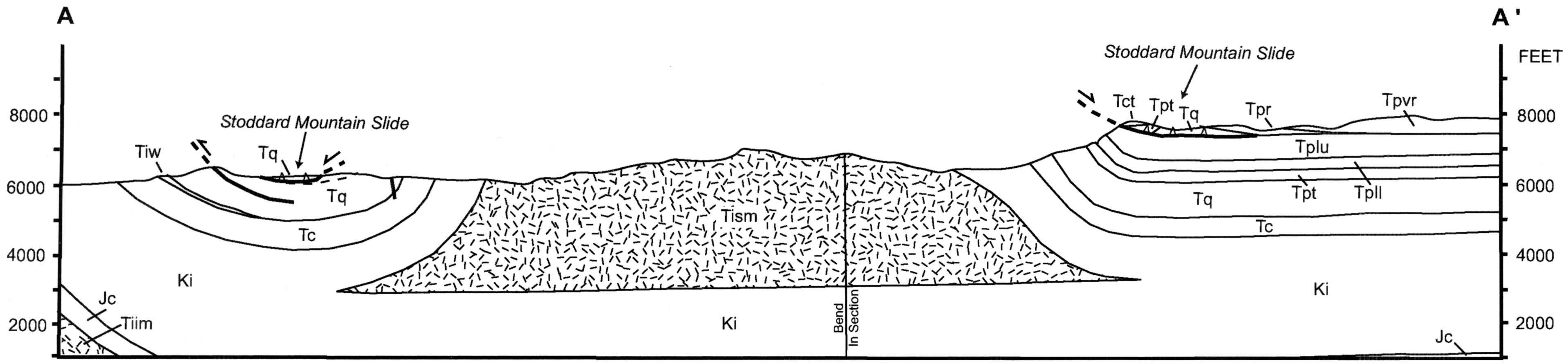


Figure 16. Geologic cross section A-A' (north-south with A to the north) through the Stoddard Mountain intrusion (see figure 12 for location of section). No vertical exaggeration.

Key (For figures 16, 17, & 18)

Tertiary	Trc	Racer Canyon Tuff
	Tpvr	Rencher Peak flow member of Pine Valley Latite
	Tpr	Page Ranch Formation
	Tct	Ash-flow tuff member of volvanic rocks of Comanche Canyon
	Tism	Stoddard Mountain intrusion
	Tid	The Dairy intrusion
	Tiim	Iron Mountain intrusion
	Tplu	Upper lava flow member of rocks of Paradise
	Tpll	Lower lava flow member of rocks of Paradise
	Tpt	Ash-flow tuff member of the rocks of Paradise
	Tpv	Vent Unit of the rocks of Paradise
	Tipp	Pinto Peak intrusion
	Tv	Local volcanic rocks
	Tq	Quichapa Group - includes: Harmony Hills Tuff, Little Creek andesite, Condor Canyon Fm, Leach Canyon Fm, and local sedimentary rocks
	Tiw	Isom and Wah Wah Springs Formations
	Tc	Claron Formation - includes thin units of Isom Fm and Wah Wah Springs Fm
Cretaceous	Ki	Iron Springs Formation
Jurassic	Jc	Carmel Formation

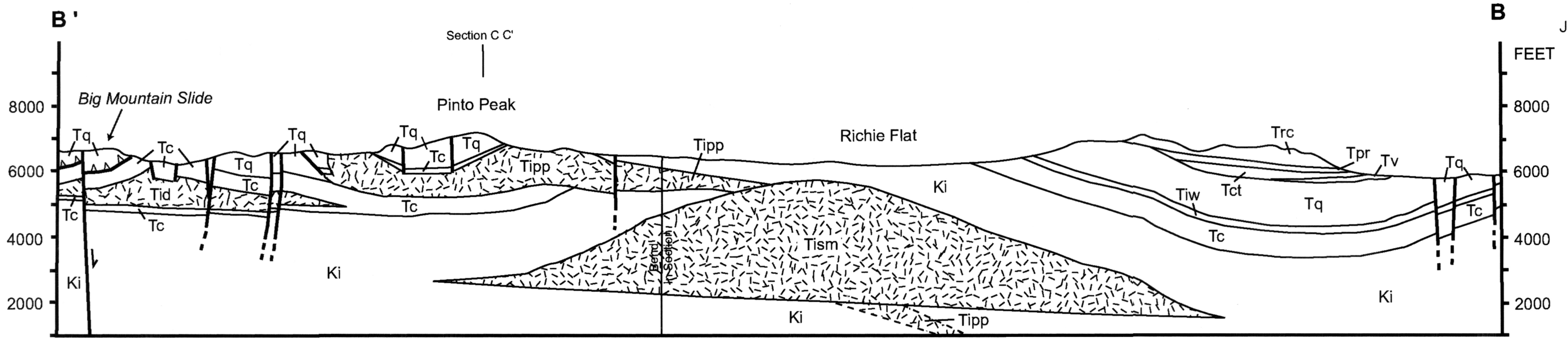


Figure 17. Geologic cross section B-B' (north-south with B' to the north) through Richie Flat and Pinto Peak (see figure 12 for location of section). No vertical exaggeration.

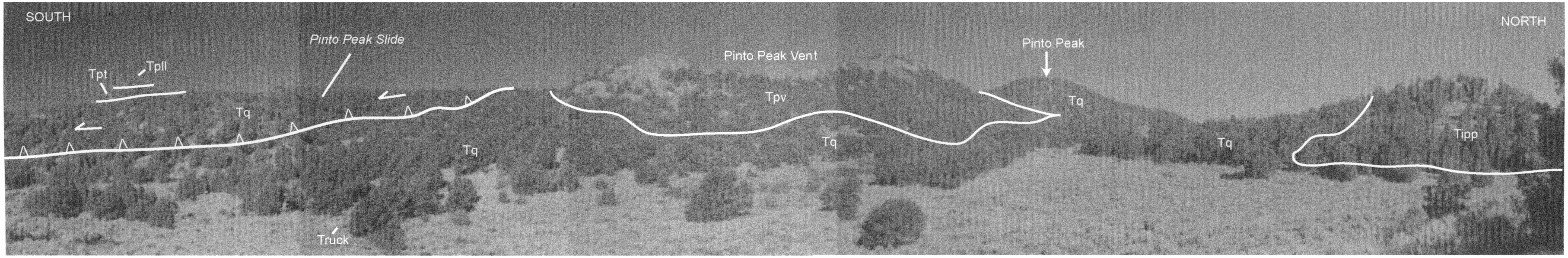


Figure 18. View looking west at vent area of the Pinto Peak intrusion at stop 6. Truck in foreground for scale. See figure 20 for closeup of vent sequence rocks.

0.1	22.1	Crest of low rise, entering Richie Flat straight ahead. Richie Flat is floored by the Iron Springs Formation that has been arched upward by the unexposed part of the Stoddard Mountain intrusion (figure 17) to form a northeast-southwest trending anticline, which the Pinto road follows. This anticlinal valley is due to the brittle nature of the overlying roof rocks that were extensionally broken along crestal faults and fractures during uplift that made them more susceptible to erosion.
0.2	22.3	Paradise ridge on left contains tilted (by Stoddard Mountain intrusion) units of the Claron Formation, Quichapa Group, and volcanic “rocks of Paradise.” The rocks of Paradise (Hacker, 1998) consist of an ash-flow tuff unit (white layer toward the top) overlain by two lava flow units (capping the ridge) that vented from the Pinto Peak intrusion. The lava flow units were originally mapped as the Paradise intrusion by Cook (1957), who believed it connected with the Stoddard Mountain intrusion. As we will see at stop 6, the lavas can be followed into the vent of the Pinto intrusion.
0.7	23.0	Kane Point ridge on right contains northward tilted (also by Stoddard Mountain intrusion) units of the Claron Formation and Quichapa Group, capped by nearly flat lying Racer Canyon Tuff (not visible from here). Hill on near right (2:00) is a Quaternary (?) slide mass made up of pink and white Claron rocks and darker colored Quichapa units that slumped southward from the cliff. Detachment occurred in the Iron Springs Formation.
0.5	23.5	Iron Springs Formation on right involved in Quaternary landslide.
1.4	24.9	Junction with road on left (Dixie National Forest road 014) leading south to Pinto Spring. **Turn left and proceed south on dirt road after crossing cattle guard.** To the left is a good view of the intrusive anticline in Richie Flat looking east along the axis of the anticline, with the Stoddard Mountain intrusion in the distance. Kane Point ridge to the north, with visible Racer Canyon Tuff cliff on top, and Paradise ridge on the south help define the scale of the anticline.
1.5	26.4	**STOP 5.** PINTO PEAK INTRUSION. Pull off into small dirt road trail on right to examine outcrop of Pinto Peak intrusion on the left (east) side of the road. The quartz monzonite is typically a resistant to crumbly, light gray to light-purplish-gray porphyry with an aphanitic to glassy groundmass. The rock is partly to highly deuterically altered with the mafic phenocrysts having a visible red halo of hematite. The more altered rock is less resistant to weathering and forms lowland areas covered in grass. Locally the intrusion has a chilled margin as thick as 2 m, consisting of a black vitrophyre with abundant white plagioclase phenocrysts. The porphyry intruded from the north into the Iron Springs Formation, then stepped up section into the Claron Formation to the south (see map of intrusion in figure 19). Hike directly west about 50 yards to a small tree-covered hill where the intrusion is in contact with the Claron Formation. The Claron here consists of a resistant basal quartzite conglomerate, which could be correlative with the Grand Castle Formation (Goldstrand and Mullet, 1997) on the Colorado Plateau to the east. The ash-flow tuff unit of the rocks of Paradise contains abundant broken quartzite clasts from this conglomerate Which originated when the intrusion vesiculated and exploded within this unit. Return to vehicles and continue south on dirt road. Pinto Peak is at 2:00 and is made of Harmony Hill Tuff.
0.8	27.2	**STOP 6.** PINTO PEAK VENT. Pull off into the sagebrush to view the vent area for the extrusive rocks mapped as the “rocks of Paradise.” Hike up hill to the east a short distance and turn around to face west to view the west part of the vent area as shown in figure 18 (see map of vent in figure 19). The dark-colored rocks at the top of the hills consist of an intra-vent sequence of ash-flow tuff intruded by dikes of dark-gray, glassy porphyritic dikes (see figure 20) with various textures ranging from dense glass to highly vesicular. The dike rocks show faint vertical flow foliation and grade down into intrusive

quartz monzonite. Harmony Hills Tuff (at Pinto Peak in the background) forms the vent wall and is in fault contact with sheared Claron rocks. To the left in the trees is the Pinto Peak slide mass consisting of Quichapa and Claron units that once covered the vent area overlying the Harmony Hills Tuff. The small knoll capping the slide mass on top consists of outflow ash-flow tuff and lava from the vent area. Hike over to and examine the vent rocks, then return to vehicles and continue south on dirt road.

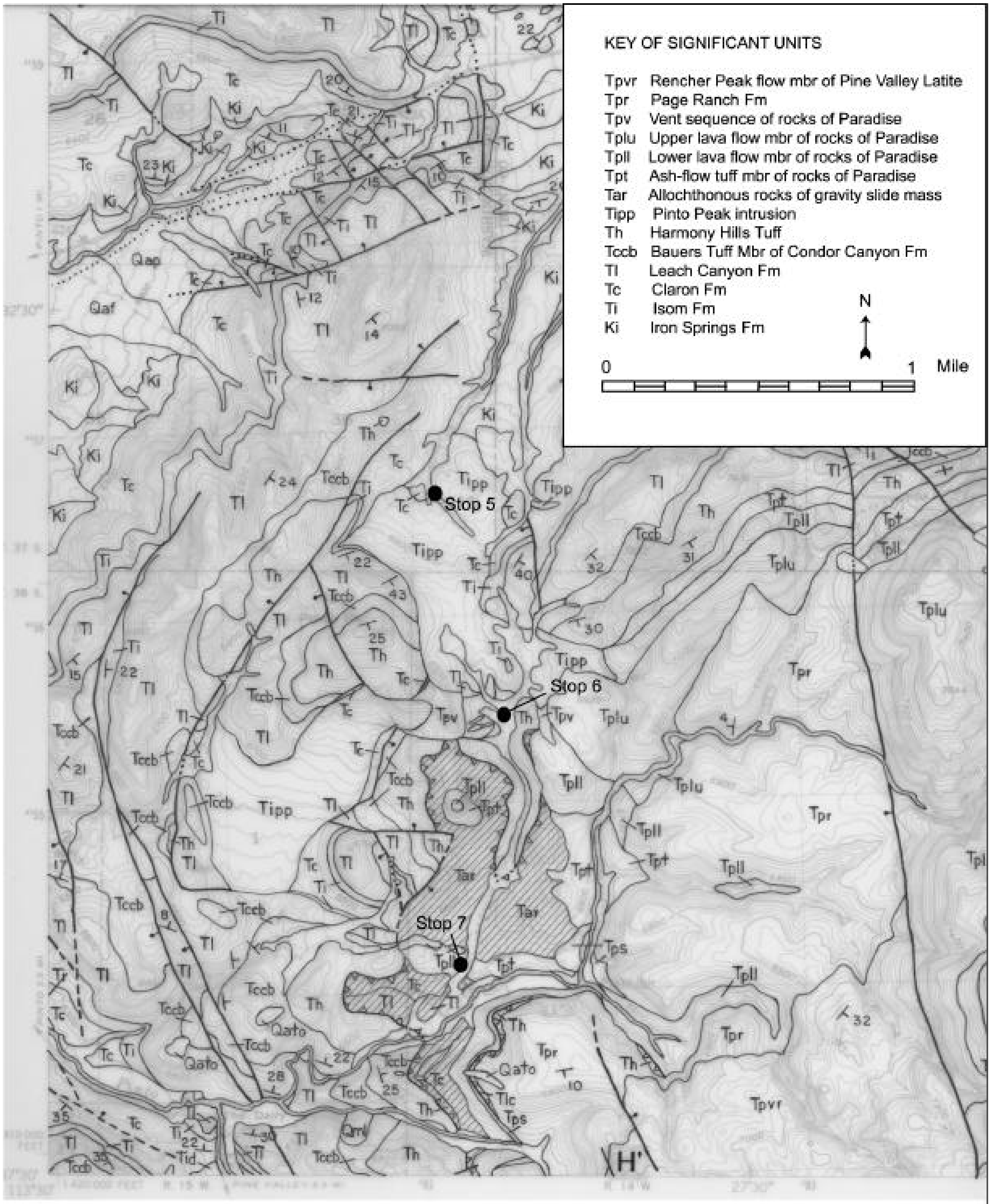


Figure 19. Geologic map of the Pinto Peak intrusion area. Based on unpublished mapping of 1:24,000 Page Ranch Quadrangle.

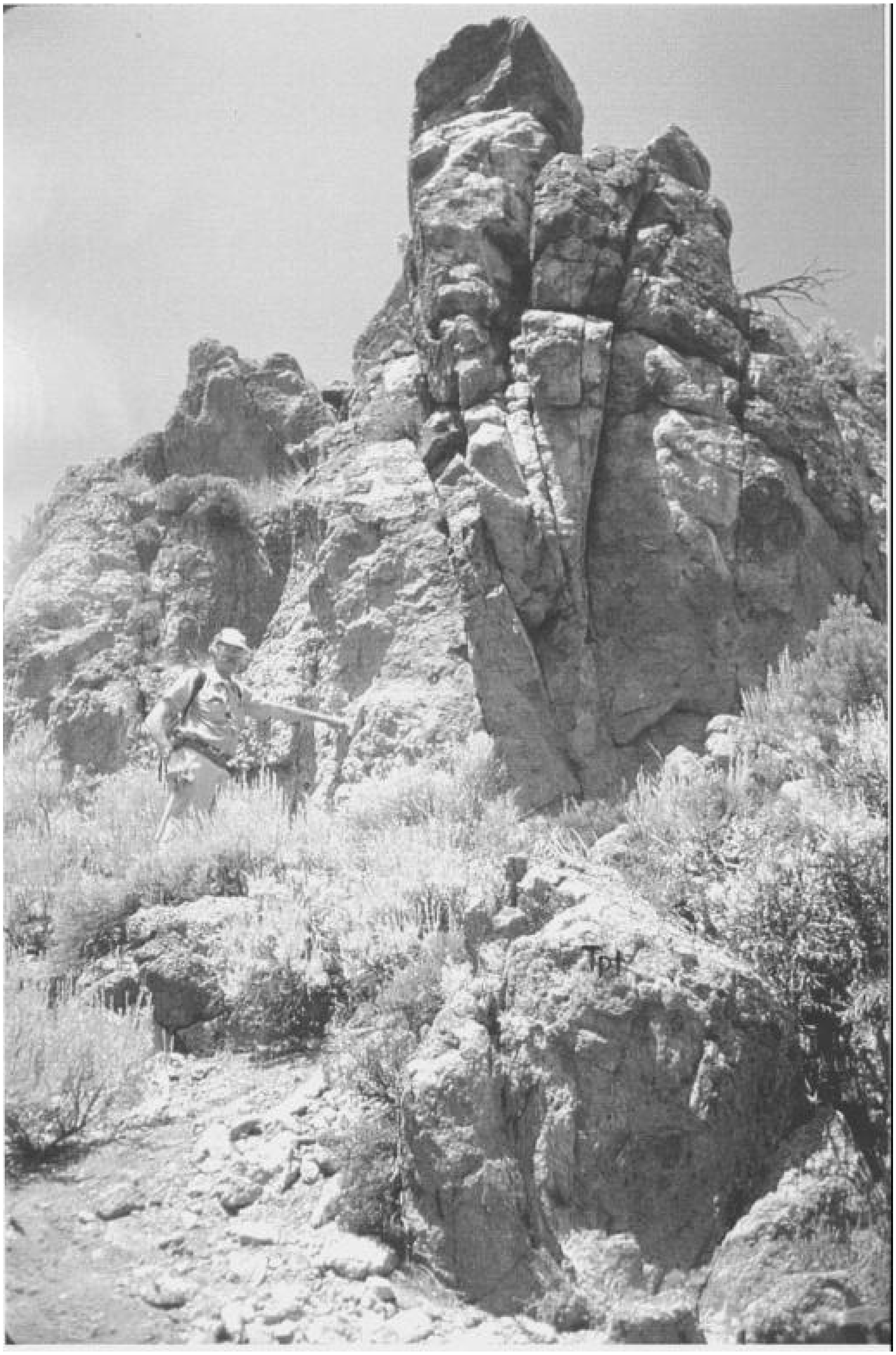
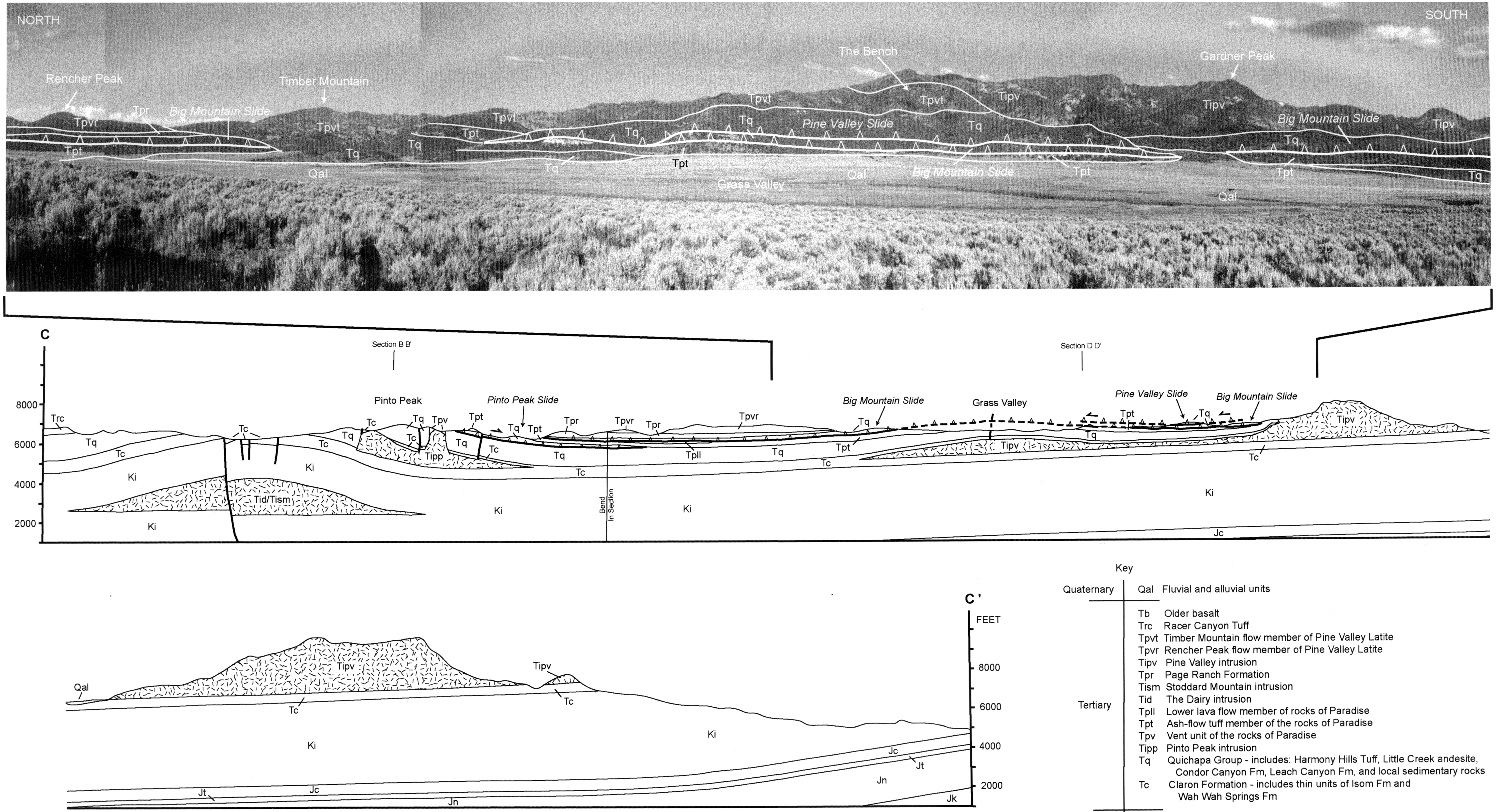


Figure 20. Photo of vent sequence with ash-flow tuff (next to Dick Blank's hand) cut by vertical dike systems of glassy lava.

0.9	28.1	Crossroads. Pinto Springs road bends to the left. Continue straight on dirt road (Dixie National Forest road 346) that then bends to the right.
0.2	28.3	**STOP 7.** ASH-FLOW TUFF MEMBER OF THE ROCKS OF PARADISE. Pull off to right on small trail road. South of the road are broken units of the Claron and Leach Canyon in the Pinto Peak slide mass. On the north side of the road are good outcrops of ash-flow tuff member that vented from the Pinto Peak intrusion. This member overlies rock units of the Pinto Peak slide mass. Note the large pumice holes and the quartzite clasts derived from the Claron Formation, which we examined previously at Stop 5. Cook (1957) initially mapped this unit as the “pebble tuff” member of the Rencher Formation because of the abundant quartzite clasts. In the field, the presence of these quartzite clasts distinguishes this tuff from similar tuff of the Rencher Formation (derived from the Bull Valley intrusion emplced into the Jurassic Carmel Formation) and volcanic rocks of Comanche Canyon (derived from the Stoddard Mountain intrusion emplaced into the Iron Springs Formation). Return to vehicles and turn around and return to Pinto road in Richie Flat.
3.3	31.6	Junction with Pinto road (Dixie National Forest road 009). **Turn left and proceed west on Pinto road.**
0.4	32.0	Crest of hill in intrusively faulted Claron and Quichipa units.
2.2	34.2	Pinto settlement. Junction of Pinto road and gravel road on right (Dixie National Forest road 011) leading north to the town of Newcastle. Continue west on Pinto road.
0.2	34.4	Junction of Pinto road and gravel road on left (Dixie National Forest road 011) leading south to Grass Valley and Pine Valley. **Turn left and proceed south on Grass Valley road.**
0.4	34.8	Houses in valley on left are located on Iron Springs Formation. Red and white Claron and brown Quichapa rocks cap the surrounding hills. Pinto is in a domal valley created by an unexposed intrusion that connects with The Dairy intrusion to the south and probably with the Stoddard Mountain intrusion to the west.
2.9	37.7	The Dairy intrusion. Quartz monzonite intruded into upper white Claron rocks that can be seen dipping north and south off the laccolith. The Dairy did not vent any volcanic units, but it deformed the rocks of Paradise and the Pinto Peak slide mass (both derived from the Pinto Peak intrusion), as well as Big Mountain slide mass (from the Bull Valley-Big Mountain arch).
0.2	37.9	Claron and the Wah Wah Springs Formation of the Needles Range Group on the right. In the northern Pine Valley Mountains and eastern Bull Valley Mountains (Blank, 1993) the ash-flow tuffs of the Wah Wah Springs Formation and members of the Isom Formation are intercalated with lacustine carbonates in the top of the Claron indicating persistence of the Claron lake in this area of Utah before succumbing to the ash-flow sheets of the Quichapa Group.
0.1	38.0	Tilted Leach Canyon Formation on right above Claron Formation.
0.1	38.1	Harmony Hills Tuff on right.
0.1	38.2	Ash-flow tuff member of the rocks of Paradise on right, overlying Harmony Hills Tuff. Lower lava flow member of the rocks of Paradise (Hacker, 1998) on left, across South Fork of Pinto Creek. This volcanic flow breccia was initially mapped by Cook (1957) as the “blue breccia” member of the Rencher Formation. This lava flow originated from the

		Pinto Peak intrusion from the vent at Stop 6.
0.2	38.4	Bridge over South Fork of Pinto Creek. Lower lava flow member of the rocks of Paradise on right.
0.2	38.6	Less brecciated zone of lower lava flow member of the rocks of Paradise on left. Samples from this exposure yielded a preliminary ^{40}Ar/^{39}Ar date on biotite of 21.93 ± 0.07 Ma (L.W. Snee, U.S. Geological Survey, written communication, 1995; Hacker and others, 1999).
0.5	39.1	Rencher Peak flow member of the Pine Valley Latite forms high cliffs at 11:00. This part of the Pine Valley Latite is slightly older than the comagmatic Pine Valley intrusion and extruded as a large lava dome and coulée type flow from the Rencher Peak area north of Grass Valley.
0.6	39.7	Outcrop on left of allochthonous Little Creek andesite, within the Big Mountain slide.
0.5	40.2	**STOP 8.** RENCHER FORMATION AND BIG MOUNTAIN SLIDE. Pull off on right by large Ponderosa pine tree. Roadcuts on left (east) side of road is of the lower ash-flow tuff member of Rencher Formation (Blank, 1993), overlain by fractured and brecciated allochthonous rocks within the Big Mountain slide. The lower ash-flow tuff member of Rencher Formation is white to light-gray, poorly to moderately indurated, crystal-rich ash-flow tuff with subrounded cognate inclusions that are darker or lighter than the host matrix and resemble pumice clasts but without vesiculation. Angular lithic fragments consist of mostly Quichapa rocks and porpyritic quartz monzonite. The lithics and cognate inclusions decrease in size and abundance eastward from the vent area in the Bull Valley Mountains and are almost nonexistent here in the Grass Valley area. Walking a short distance southward along the road, you will see exposures of Isom, Leach Canyon, and Little Creek andesite within the Big Mountain slide mass. This portion of the slide is on the former land surface composed of the lower ash-flow tuff member of Rencher Formation. Return to vehicles and continue south on Grass Valley road.
0.5	40.7	Roadcut on left of allochthonous Leach Canyon Formation.
0.1	40.8	Roadcut on left of allochthonous white and pink Claron, and a thin unit of Wah Wah Springs Formation.
0.2	41.0	Outcrops on left and right of allochthonous Little Creek andesite. The Little Creek andesite is an autochthonous unit present between the Bauers and Harmony Hills units in the Bull Valley Mountains, but in the Pine Valley Mountains appears in allochthonous rocks within slide masses only (Bull Valley slide and Big Mountain slide). Crest of hill (at cattle-guard in road) is the drainage divide of South Fork of Pinto Creek (draining north into the Escalante basin) and Grass Valley Creek (part of the Colorado River drainage basin).
0.3	41.3	Road on left to Mill Canyon trailhead and Broken Arrow Ranch (formerly Rencher Ranch, which is the type locality of the Rencher Formation). Continue south on Grass Valley road.
1.2	42.5	**STOP 9.** GRASS VALLEY. Pull off at crest of hill to view geologic features on east side of Grass Valley (see figure 21). Grass Valley is another eroded anticline produced by an unexposed intrusion that is interpreted to be an extension of the Pine Valley intrusion exposed in the hills to the right. The white cliffs just above the valley floor are exposures of the ash-flow tuff member of the rocks of Paradise overlain by the Big Mountain slide mass. The Rencher Formation was not deposited this far east. The Big Mountain slide is overlain by fanglomerates of the Page Ranch Formation and the Pine

		Valley slide mass. Overlying the Pine Valley slide is the Timber Mountain flow member of the Pine Valley Latite that extruded northward from the Pine Valley laccolith following the collapse of its flank by gravity sliding. Rencher Peak (source area for the slightly older Rencher Peak flow member of the Pine Valley Latite) is the high peak visible on the north side of Grass Valley. Return to vehicles and continue south on Grass Valley road.
1.0	43.5	Cinder cone of quartz-bearing basalt on right. Lava from this and other vents dammed Grass Valley, which then filled in with fluvial sediments and minor lacustrine deposits to form the relatively broad valley floor. Many fertile valleys in this area formed in this manner, including Pine Valley, Grassy Flat, and Diamond Valley.
1.7	45.2	Road turns to the southeast, with a view of Pine Valley and rugged topography of the Pine Valley laccolith. The Pine Valley laccolith intruded into the middle to upper part of the Claron Formation.
0.4	45.6	Junction with Central-Pine Valley road (paved). **Turn left (east) at stop sign and proceed to town of Pine Valley.**
0.5	46.1	Descend hill of basalt into Pine Valley, which is drained by the Santa Clara River. Hills to left are made of autochthonous units of the Claron Formation (tree covered red and white rocks at base of hill) and Quichipa Group rocks. Gardner Peak is on the left at 11:00 and is part of a westward extending lobe of the Pine Valley laccolith. Mapping of the peripheral structures of the laccolith indicates that this lobe did not extend across Pine Valley and connect with the main part of the intrusion to the south. The Santa Clara River and Pine Valley therefore occupy a natural, easily erodable V-shaped discontinuity in the laccolith.
0.2	46.3	Cross Santa Clara River.
0.5	46.8	Junction with Main Street in town of Pine Valley. **Turn left (east) at stop sign and proceed through town.**
0.1	46.9	Dixie National Forest Pine Valley visitor center and workstation on right.
1.4	48.3	Pine Valley Recreation Area sign on right.
0.7	49.0	Red soil on left belongs to the Claron Formation, which is capped by quartz monzonite porphyry of the Pine Valley intrusion.
0.2	49.2	Pine Valley Reservoir on right.
0.5	49.7	**STOP 10.** PONDEROSA PICNIC AREA. Turn right into picnic parking area for lunch stop. We are now within the eroded part of the laccolith and about level with its base. Alluvial boulders of quartz monzonite porphyry litter the picnic area. Samples taken from the campground area yielded a K-Ar date on biotite of 20.9 ± 0.6 Ma (Mckee and others, 1997). After lunch return to vehicles and turn left out of parking area and retrace route back to the town of Pine Valley.
2.9	52.6	Junction with road to Central. **Turn right (north).**
1.2	53.8	Junction with Grass Valley road to Pinto on right. Continue westward on paved Central-Pine Valley road.
0.5	54.3	Junction with road to Grassy Flat on right. Continue westward on paved main road that is on basalt that erupted from a buried northeast striking basin-range fault or fissure that is lined with cinder cones (to the left and right).

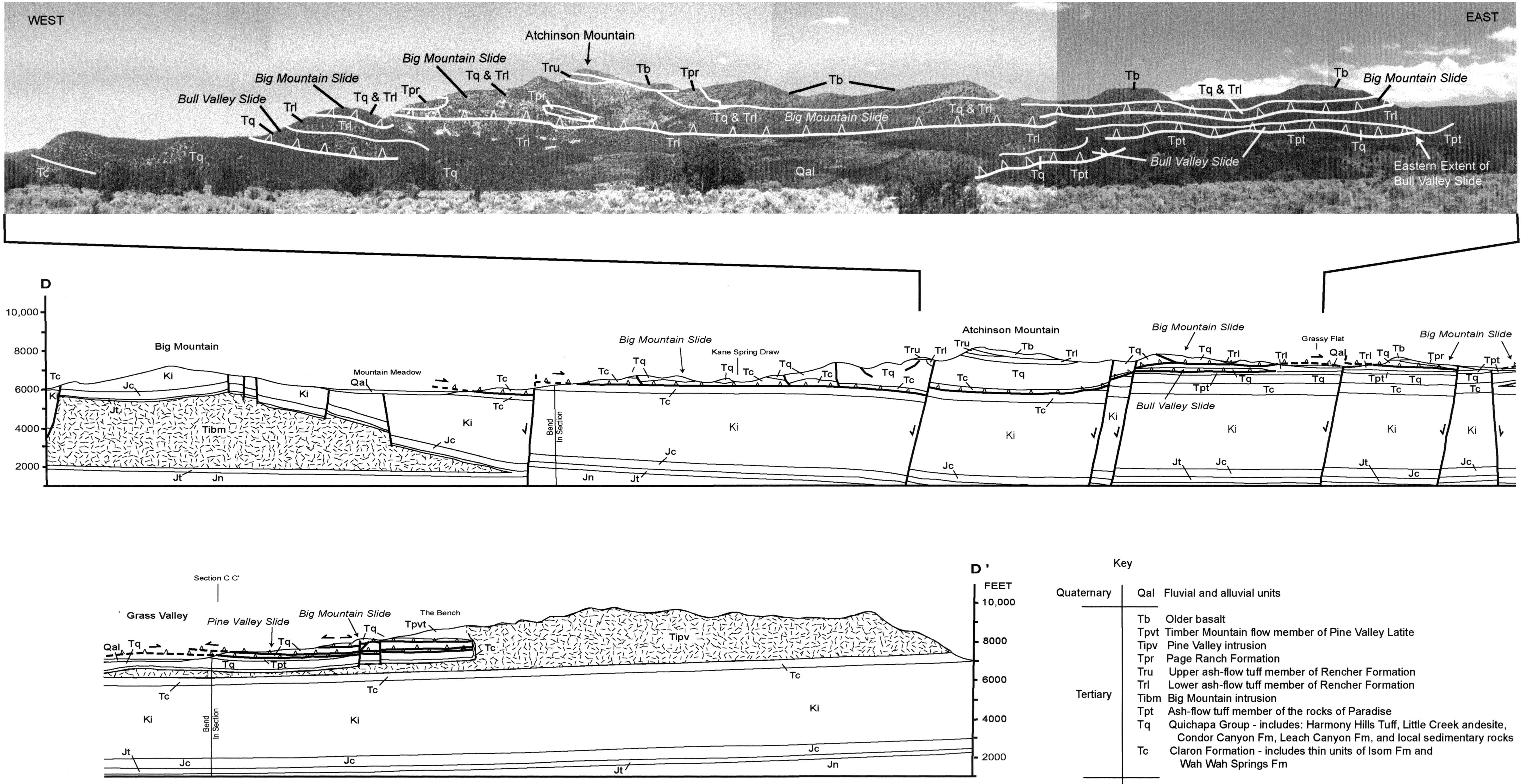


	Key	
Quaternary	Qal	Fluvial and alluvial units
Tertiary	Tb	Older basalt
	Trc	Racer Canyon Tuff
	Tpvt	Timber Mountain flow member of Pine Valley Latite
	Tpvr	Rencher Peak flow member of Pine Valley Latite
	Tipv	Pine Valley intrusion
	Tpr	Page Ranch Formation
	Tism	Stoddard Mountain intrusion
	Tid	The Dairy intrusion
	Tpll	Lower lava flow member of rocks of Paradise
	Tpt	Ash-flow tuff member of the rocks of Paradise
	Tpv	Vent unit of the rocks of Paradise
	Tipp	Pinto Peak intrusion
	Tq	Quichapa Group - includes: Harmony Hills Tuff, Little Creek andesite, Condor Canyon Fm, Leach Canyon Fm, and local sedimentary rocks
	Tc	Claron Formation - includes thin units of Isom Fm and Wah Wah Springs Fm
Cretaceous	Ki	Iron Springs Formation
Jurassic	Jc	Carmel Formation
	Jt	Temple Cap Formation
	Jn	Navajo Sandstone
	Jk	Kayenta Formation

Figure 21. Geologic cross section C-C' (north-south with C to the north) through the Pine Valley Mountains (see figure 12 for location of section) and photo view of Grass Valley area (looking east) at stop 9. No vertical exaggeration.

0.8	55.1	View of Square Top Mountain in the far distance at 10:00 and the Bull Valley Mountains to its right (north) at 11:00 to 12:00. The white rocks visible in the Bull Valley Mountains consist of ash-flow tuffs of the Rencher Formation located very near their vent from the Bull Valley intrusion. Square Top Mountain consists of subhorizontal Pennsylvanian and Permian rocks thrusted eastward over rocks as young as the Upper Cretaceous Iron Springs Formation during the Sevier orogeny (Hintze, 1986). Atchinson Mountain at 2:00.
1.1	56.2	**STOP 11.** ATCHINSON MOUNTAIN. Pull off to right into driveway of Forest Service dump station to view geologic features of Atchinson Mountain (see figure 22). Both the Bull Valley and Big Mountain slides are present here. Return to vehicles and continue westward on Central-Pine Valley road.
1.0	57.2	Red and white autochthonous Claron to right (north of road) overlain by autochthonous Quichapa rocks that lack the Little Creek andesite between the Bauers and Harmony Hills.
0.9	58.1	Junction with jeep trail (Dixie National Forest road 823) to right that leads to Eight Mile Spring. The Big Mountain slide forming the rugged topography west of Atchinson Mountain contains allochthonous Little Creek andesite rocks within the Qichapa Group and rests upon a low-angle bedding detachment fault within the Claron Formation. The near hill at 2:00 is made of Pliocene (?) to Pleistocene Eight Mile Dacite of Cook (1957) that forms a series of northeast trending volcanic domes erupted along a basin-range fault similar to the basalt flows at mile 54.3. Continue westward on Central-Pine Valley road.
0.2	58.3	Older alluvial-fan deposits of Pliocene to Pleistocene age overlain by Pleistocene lava flows. Apparent dike of basalt cuts the alluvium on right.
0.2	58.5	View of Eight Mile Dacite on right showing conspicuous flow layering and folding within the volcanic domes. Basalt flows on south side of road (left).
0.6	59.1	Dixie National Forest boundary. Hill at 2:00 above the town of Central is composed of autochthonous Claron and Quichapa (without Little Creek andesite). At the top are remnants of the Bull Valley slide forming klippes of allochthonous Claron and Quichapa resting on the former land surface of Harmony Hill Tuff and ash-flow tuff member of the rocks of Paradise. Behind the hill to the north is the high-angle southern bounding fault of the Big Mountain slide. The far hill is composed of a thick section of allochthonous Little Creek andesite faulted against autochthonous rocks exposed in the hill just described above the town of Central.
0.8	59.9	**STOP 12.** GRAVITY-SLIDE MASS OF PINE VALLEY LACCOLITH. Pull off on left opposite the Saucer Ranch (on right) to view geologic features of the Pine Valley laccolith to the south (see figure 23). This marks the western extent of the Pine Valley laccolith. The floor of the laccolith can be observed where the dark quartz monzonite of the intrusion overlies red Claron rocks. The small hill to the right contains a full Claron section and a partial Quichapa section (see figure 23). Return to vehicles and continue westward on Central-Pine Valley road.
0.2	60.1	Basin-range faulted autochthonous Claron and Isom (across from 300 East Street) rocks in roadcuts on right.
0.5	60.6	Intersection with SR-18. **Turn right at stop sign and proceed north toward the town of Enterprise.**

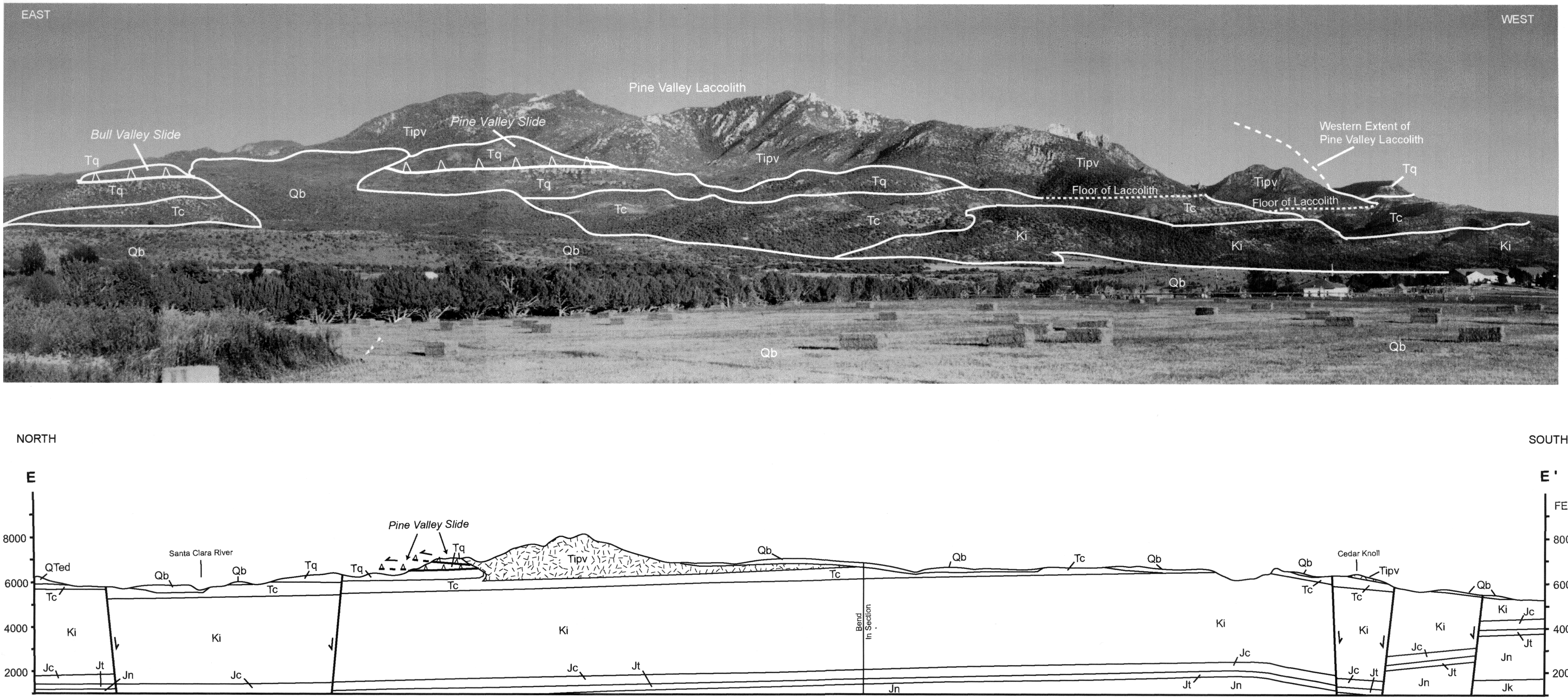
0.6	61.2	Hills from 1:00 to 3:00 are capped with klippes of the Bull Valley slide composed of Quichapa and Claron rocks overlying the former land surface fault. Hills on the left are composed of the autochthonous lower ash-flow tuff member of the Rencher Formation, capped by younger basalts. The Rencher here overlies the Bull Valley slide.
1.7	62.9	**STOP 13.** SOUTHERN FLANKING FAULT OF THE BIG MOUNTAIN SLIDE. Pull off road to right (0.1 mile before Kane Spring). We just crossed over the southern strike-slip bounding fault of the Big Mountain slide. This fault can be seen in the notch between the two hills on the left (east) side of road. Looking east, the hill on the left (north) is capped with a thick section of Little Creek andesite underlain by Bauers and Leach Canyon rocks. These allochthonous units are part of the Big Mountain slide and have traveled eastward away from the viewer. The hill on the right consists of a thinner section of autochthonous Bauers and Harmony Hills Tuff (with no intercalated Little Creek andesite) capped by klippes of the Bull Valley slide. Return to vehicles and continue north on SR-18.
0.2	63.1	SR-18 bends to the left (west). Hills to the right are composed of mostly Bauers, Little Creek andesite, and Harmony Hills Tuff of the Big Mountain slide above a bedding detachment fault.
0.7	63.8	Hill at 2:00 to 3:00 consists of white rocks of the lower ash-flow tuff member of the Rencher Formation, capped by darker red rocks of the upper ash-flow tuff member of the Rencher Formation. The lower Rencher here is allochthonous and part of the Big Mountain slide and capped by the autochthonous upper Rencher , which immediately followed the Big Mountain sliding episode.
0.5	64.3	Roadcut of allochthonous white, lower ash-flow tuff member of the Rencher Formation on left and right. The lower Rencher here contains larger lithic clasts and cognate inclusions than found in Grass Valley, indicating a closer proximity to the vent (Bull Valley intrusion). Road bends to the right (north).
0.5	64.8	**STOP 14.** BIG MOUNTAIN SLIDE IN ROADCUT. Pull off road to the right and park near the “Mountain Meadow ½ Mile” sign. Roadcut on right consists of highly fractured and brecciated rocks of Little Creek andesite and Harmony Hills Tuff in the Big Mountain slide mass. These and lower units override the bedding detachment fault segment of the slide. Detachment is within the lower Claron and/or upper Iron Springs. Return to vehicles and continue north on SR-18.
0.5	65.3	Entrance road to Mountain Meadow monument of left. **Turn left on the monument road** and proceed to the overlook parking area on Dan Sill Hill. You will pass a gravel road on the left that leads to the gravesite in Mountain Meadow. This is the site of the infamous 1857 massacre of about 120 emigrants while they were traveling the Old Spanish Trail that traverses Mountain Meadow.
0.1	65.4	**STOP 15.** MOUNTAIN MEADOW OVERLOOK. Park in monument parking lot and take short paved trail to monument overlook on Dan Sill Hill. The hill is made of the upper ash-flow tuff member of the Rencher Formation overlying allochthonous Claron rocks of the Big Mountain slide. Northwest of Mountain Meadow is Big Mountain (with radio towers on top) at the northern end of the Bull Valley-Big Mountain arch. The Big Mountain dome consists of Iron Springs rocks and locally some Carmel limestone and intrusive quartz monzonite. In this denuded area of the arch, the Iron Springs and/or Claron are overlain by the upper ash-flow tuff of the Rencher Formation. The hills on the east side of Mountain Meadow (east of SR-18 behind you) consist of the Big Mountain slide that originated on Big Mountain. These hills contain a thick section of the conspicuous allochthonous white lower Rencher, which has slid eastward from the



Key

Quaternary	Qal	Fluvial and alluvial units
Tertiary	Tb	Older basalt
	Tpvt	Timber Mountain flow member of Pine Valley Latite
	Tipv	Pine Valley intrusion
	Tpr	Page Ranch Formation
	Tru	Upper ash-flow tuff member of Rencher Formation
	Trl	Lower ash-flow tuff member of Rencher Formation
	Tibm	Big Mountain intrusion
	Tpt	Ash-flow tuff member of the rocks of Paradise
	Tq	Quichapa Group - includes: Harmony Hills Tuff, Little Creek andesite, Condor Canyon Fm, Leach Canyon Fm, and local sedimentary rocks
	Tc	Claron Formation - includes thin units of Isom Fm and Wah Wah Springs Fm
Cretaceous	Ki	Iron Springs Formation
Jurassic	Jc	Carmel Formation
	Jt	Temple Cap Formation
	Jn	Navajo Sandstone

Figure 22. Geologic cross section D-D' (east-west with D to the west) through the Pine Valley Mountains (see figure 12 for location of section) and photo view of Atchinson Mountain area (looking north) at stop 11. No Vertical exaggeration.



Key

Quaternary	Qb	Basalt
	QTed	Eight Mile Dacite
Tertiary	Tipv	Pine Valley intrusion
	Tq	Quichapa Group - includes: Harmony Hills Tuff, Little Creek andesite, Condor Canyon Fm, Leach Canyon Fm, and local sedimentary rocks
	Tc	Claron Formation - includes thin units of Isom Fm and Wah Wah Springs Fm
Cretaceous	Ki	Iron Springs Formation
Jurassic	Jc	Carmel Formation
	Jt	Temple Cap Formation
	Jn	Navajo Sandstone
	Jk	Kayenta Formation

Figure 23. Geologic cross section E-E' (north-south with E to the north) through west end of Pine Valley laccolith (see figure 12 for location of section) and photo view of Pine Valley laccolith (looking south) at stop 12 near the town of Central. No vertical exaggeration.

		crest or flank of the Bull Valley – Big Mountain arch prior to the eruption of the upper Rencher. Return to vehicles and return to SR-18 via the overlook access road.
0.1	65.5	Intersection with SR-18. **Turn left at stop sign and continue north on SR-18.**
2.2	67.7	Junction with gravel road on right leading to Pinto. Continue northward on SR-18. Iron Springs Formation to the left, on flank of Big Mountain.
1.2	68.9	Roadcuts of Iron Springs Formation on north flank of Big Mountain dome.
1.5	70.4	Road crosses Big Mountain normal fault on the west flank of Big Mountain that strikes north and is down-to-the-west. Big Mountain formed by arching of the laccolith (Blank, 1993). Here Claron (exposed in the roadcuts) is juxtaposed against Iron Springs on the west, and farther south on the arch, Claron is juxtaposed against Carmel.
1.0	71.4	Roadcut through red ash flow-tuff (Bald Hills Member) of the Isom Formation.
1.4	72.8	Dixie National Forest boundary sign. Road is cutting through Piocene/Pleistocene valley-fill deposits of Blank (1993).
1.3	74.1	T-Junction (at stop sign) with road to left leading to the town of Enterprise and the continuation of SR-18 to the right. **Turn right (northeast) at stop sign and continue on SR-18.**
0.8	74.9	Hill on right at 3:00 is Gum Hill, consisting of basin-fill deposits capped by basalt of Gum Hill of Blank (1993), which has recently been dated with a minimum age of 5.7 Ma (Cornell and others, 2001). The basalt dips to the east into the southern extension of the Antelope Range fault (range-front fault) of Siders and others (1990).
1.8	76.7	Junction with paved road on right leading to the town of Newcastle with SR-18 bending left. **Turn right on Newcastle road.** View of southern extension of Antelope Range that is currently being mapped by Tom Butler and Don Cornell of Kent State University as part of their Masters theses (see Butler and others, 2001; Cornell and others, 2001).
3.4	80.1	Old Spanish Trail sign on right. Red and white rocks in the footwall block of the Antelope Range fault at 12:00 consist of basin-range faulted Iron Springs and Claron units capped by darker volcanics of the Isom Formation and Quichapa Group.
3.0	83.1	Greenhouses on left at 10:00 are heated by a blind geothermal resource associated with the range-front fault. The resource was delineated by geophysical surveys and drilling (Mabey and Budding, 1987; Blackett and others, 1990). Thickness of the basin west of the range-front fault is estimated to be ~ 3 km based on geophysical surveys (Pe and Cook, 1980).
2.3	85.4	Junction of Main Street of Newcastle with SR-56. **Turn right at stop sign and proceed east toward Cedar City.**
1.3	86.7	Hills on north side of the road at 10:00 consist of rhyolite lava domes (rhyolite of Silver Peak) dated at 8.4 ± 0.4 Ma (Shubat and Siders, 1988; Siders and others, 1990).
3.7	90.4	Gravel road to Desert Mound on left. Continue eastward on SR-56. View of iron mines on flank of Iron Mountain straight ahead. Stoddard Mountain is the tallest peak to the right of Iron Mountain just above the road in the distance.
1.6	92.0	The red hill north of the road on right (in the distance) consists of north-dipping, orange

		clastics of the Iron Springs Formation with a cap of red Claron. The Claron wraps around the Iron Mountain intrusion. Iron Mountain is the tallest peak, with radio towers and a mine scar on top. Outcrops to right along the road also consist of Iron Springs and Claron dipping to the south. We are now traveling on the western extension of the up arched dome of the Iron Mountain intrusion.
2.7	94.7	Gravel road to old Irontown ruins on left. Continue eastward on SR-56. Irontown was an early iron ore-smelting site constructed and operated by pioneers in the late 1800s.
1.4	96.1	Gravel turnoff on left to Blowout pit. **Turn left on gravel drive for optional stop at iron workings at Blowout pit.** Keep to the right on gravel drive until small dirt trail on left.
0.2	96.3	**STOP 16.** OPTIONAL STOP AT BLOWOUT PIT. Stop and park by dirt trail on left. Walk a short distance north up trail to the rim of pit, now partially filled with water. Here hematite replacement ore deposits formed in the vertical to overturned Homestake Limestone Member (gray rocks in pit wall) of the Carmel Formation in the upper plate of the Sevier Iron Springs Gap thrust. Quartz monzonite porphyry is exposed in the opposite pit wall. Return to vehicles and return to SR-56.
0.3	96.6	Junction with SR-56. **Turn left and continue eastward to Cedar City.**
1.4	98.0	Junction with Pinto road on right. Continue eastward to Cedar City.
17.4	115.4	End of field trip at intersection of SR-56 (200 North) and Main Street.

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