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Cenozoic volcanism and plate-tectonic evolution of the Western United States. I. Early and Middle Cenozoic†

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CONTENTS

Variations in Cenozoic volcanism in the Western United States correlate rather closely with changes in tectonic setting: intermediate-composition rocks and their associated differentiates were erupted through orogenic or fairly stable crust, whereas basaltic or bimodal basalt-rhyolite suites were erupted later—concurrently with crustal extension and normal faulting.

Lower and middle Cenozoic continental lavas, erupted onto postorogenic terranes, are predominantly intermediate types (andesite to rhyodacite), commonly with closely associated more silicic ash-flow sheets. Compositional zonations in individual ash-flow sheets, from rhyolite upward into quartz latite, record magmatic differentiation in underlying batholithic source chambers. The intermediate lavas probably represent the greater part of these batholiths and the ash-flow tuffs their differentiated tops. Continental volcanic activity of this type was most voluminous in the northwestern United States in Eocene time, but shifted southward in the Oligocene; contemporaneous sea-floor basalts occur in the Oregon–Washington coast ranges.

Largely intermediate-composition calc-alkalic igneous suites, that become more alkalic toward the continental interior, are characteristic of most of the North and South American cordilleran belt. Similar volcanic associations are forming now around most of the Pacific margin where continental plates override oceanic crust along active subduction systems, marked by Benioff seismic zones and oceanic trenches. A similar subduction mechanism probably operated in the Western United States until late Cenozoic time. Analogy with chemical variations across active island arcs suggest that early and middle Cenozoic subduction occurred along two subparallel imbricate zones that dipped about 20° eastward. The western zone emerged at the continental margin, but the eastern zone was entirely beneath the continental plate, partly coupled to the western zone below the low-velocity layer.

Predominantly intermediate-composition volcanism persisted throughout the Western United States until the initial intersection of North America with the East Pacific rise started the progressive destruction of the subduction system.

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Introduction

The Pacific margins of North and South America are marked by a belt of Mesozoic and Cenozoic volcanism, plutonism, and tectonism about 500 km wide that grades eastward into terranes of gentle structure and minor igneous activity. In middle North America, however, the belt bulges eastward to form a zone as wide as 1500 km (figure 1) that is characterized by complexly overlapping structures and igneous activity. Within this anomalously wide area, we

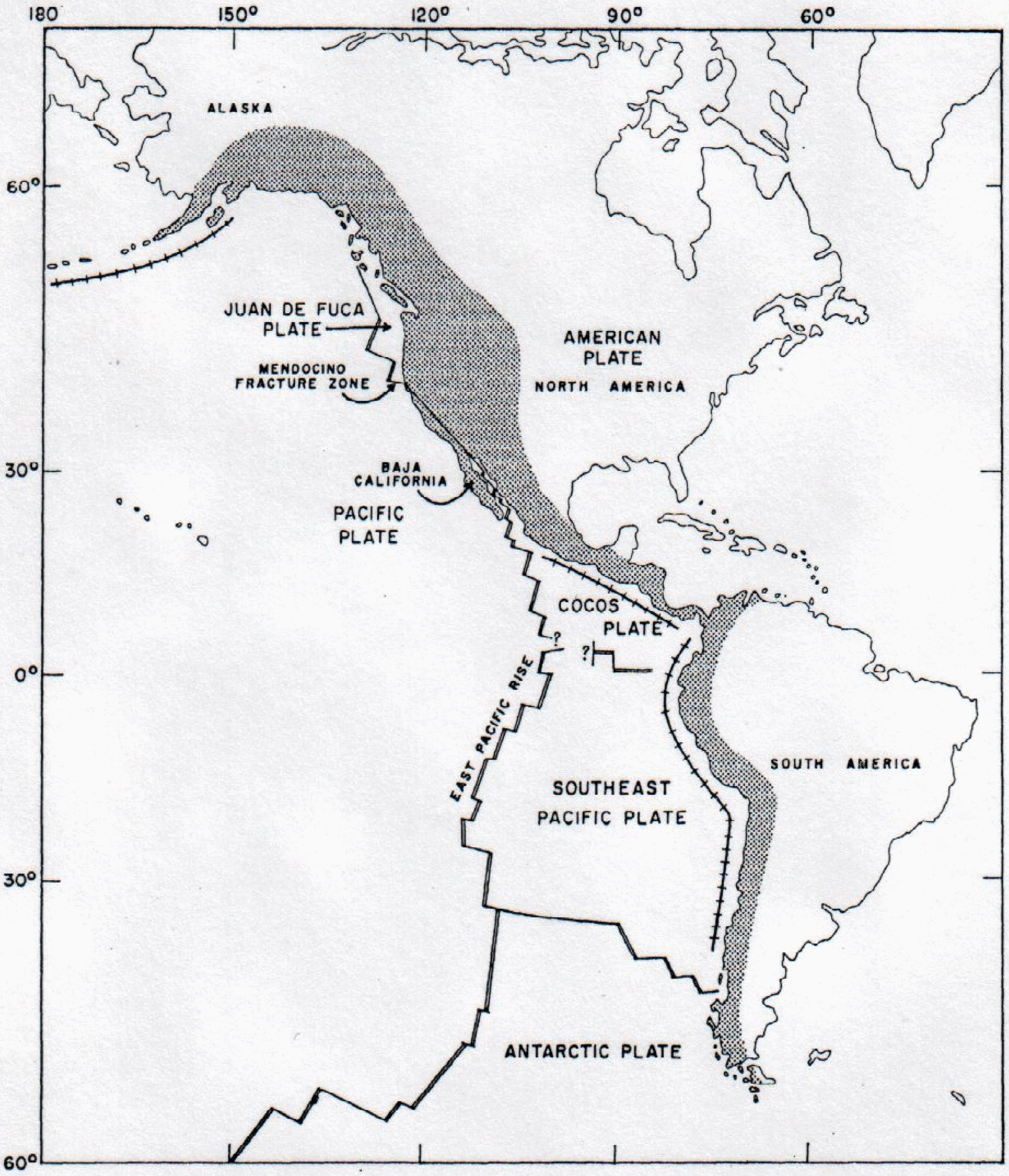


Figure 1. Present boundaries of major tectonic plates of the eastern Pacific region (modified from Bullard (1969) and Morgan (1968)), and approximate extent of late Mesozoic and Cenozoic calc-alkalic igneous belt (stippled) in western North and South America. Belt is about 500 km wide, except in the Western United States, where it is as much as 1500 km wide (width of belt in Alaskan area is exaggerated by Mercator projection). Double lines indicate spreading ridges; single lines, transform faults; 'railway tracks', oceanic trenches.

believe that variations in Cenozoic volcanism can be correlated rather closely with tectonic setting. In this and the second part of this paper (Christiansen & Lipman, this volume, p. 249) we summarize the characteristics of Cenozoic volcanism in the Western United States,†

† This and the following part of this paper are concerned with events in the coterminous Western United States, i.e. the United States exclusive of Alaska and Hawaii. For convenience we use the term 'Western United States' to refer to this region.

describe a shifting pattern of volcanism in relation to regional tectonic features, and relate these changing volcano-tectonic systems to concepts of the plate-tectonic evolution of the north-eastern part of the Pacific Ocean and western North America.

The plate-tectonic concept is becoming too widely known and generally accepted to require extensive discussion here. In brief, this concept holds that most active tectonic features of the Earth's surface are related to motions between a small number of semi-rigid plates of lithosphere. New volcanic crust is formed at divergent plate boundaries where plates are moving apart; crust is consumed at convergent boundaries where one plate slides beneath the other along an inclined subduction zone; crust is neither created nor destroyed where plates slide horizontally past each other along transform faults. Most of the world's active volcanism is associated with plate convergence or divergence or with extensional rifting within a plate. Plate convergence along subduction zones, which are defined by intermediate to deep seismic activity, localizes andesitic volcanism in island and continental-margin arc systems; folding, thrusting, and metamorphism are associated tectonic features. Some recent papers that develop these concepts are Morgan (1968), Isacks, Oliver & Sykes (1968), LePichon (1968), McKenzie & Morgan (1969), Hamilton (1969*a*), Dewey & Bird (1970), and Dickinson (1970).

We interpret the complex Cenozoic volcanic and tectonic features of the Western United States as products of an essentially two-stage history involving a régime of continental-margin plate convergence and related andesitic volcanism that lasted until middle Cenozoic time (Lipman 1970; Lipman, Steven & Mehnert 1970), followed by fundamentally basaltic volcanism associated with extension within the continental plate that has continued to the present time (Christiansen & Lipman 1970).

The observation that Cenozoic volcanic activity over large areas of the Western United States began with eruption of voluminous andesites and related rocks, followed by rhyolites and basalts, was first made many years ago (Lindgren, Graton & Gordon 1910, pp. 42–46; also Callaghan 1951). Only recently, however, has the availability of numerous radiometric age determinations permitted regional synthesis of Cenozoic igneous activity.

Available data indicate one peak of igneous activity of late Cretaceous to very early Tertiary age, approximately contemporaneous with 'Laramide' tectonism (Damon & Mauger 1966), that will be considered only briefly here. Laramide igneous activity appears to be generally similar in distribution and in petrology to the subsequent middle Cenozoic activity (Eocene–Oligocene). Post-Laramide Cenozoic volcanic rocks, erupted before widespread commencement of basin-range extensional faulting in Miocene time, are mainly intermediate types (andesite and rhyodacite), and associated more silicic differentiates (quartz latite and low-silica rhyolite). These rocks, which we relate to convergent plate motions, constitute a petrologically coherent group that was erupted under conditions of relative crustal stability.

Late Cenozoic igneous activity, associated in time and space with basin-range extensional faulting, is the subject of part II of this paper (Christiansen & Lipman, this volume, p. 249); this activity, although petrologically diverse, is described by us as fundamentally basaltic, because it produced mainly basalt fields, differentiated alkalic basaltic suites, and bimodal associations of basalt and alkalic high-silica rhyolite. The basin-range faulting that we consider to be the characteristic tectonic association of fundamentally basaltic volcanism is not confined to just the basin-range physiographic province but is a late Cenozoic tectonic feature of much of the Western United States.

Our correlations between tectonic environment and volcanic association, if valid, have important implications for hypotheses of origin of the contrasting igneous suites; however, such petrogenetic problems are not critical to our volcano-tectonic interpretations and are, accordingly, not considered in detail.

Pre-basin-range igneous activity

The lower and middle Cenozoic volcanic fields and associated intrusives that we relate to plate convergence constitute a broad petrologic association of predominantly intermediate compositions. Lavas are typically andesite, dacite, rhyodacite, and quartz latite; the common hypabyssal intrusives are granodiorite, monzonite, and quartz monzonite. Somewhat more silicic rocks, especially ash-flow tuffs of quartz latite and low-silica rhyolite that are abundant in some areas, can plausibly be interpreted as high-level differentiates of intermediate-composition magmas. Very mafic and very silicic rocks are much less abundant; in many fields they are sparse or absent. Deviations from this broad outline are most conspicuous at the Pacific and the continental-interior margins of the Cenozoic volcanic province where mafic rocks are more abundant. The major areas of lower and middle Cenozoic igneous rocks for which age and petrologic data are available are listed in table 1. The nature of this lithologic association is best indicated by description of a few carefully studied areas, as presented below, but first the distribution of early and middle Cenozoic igneous activity is summarized.

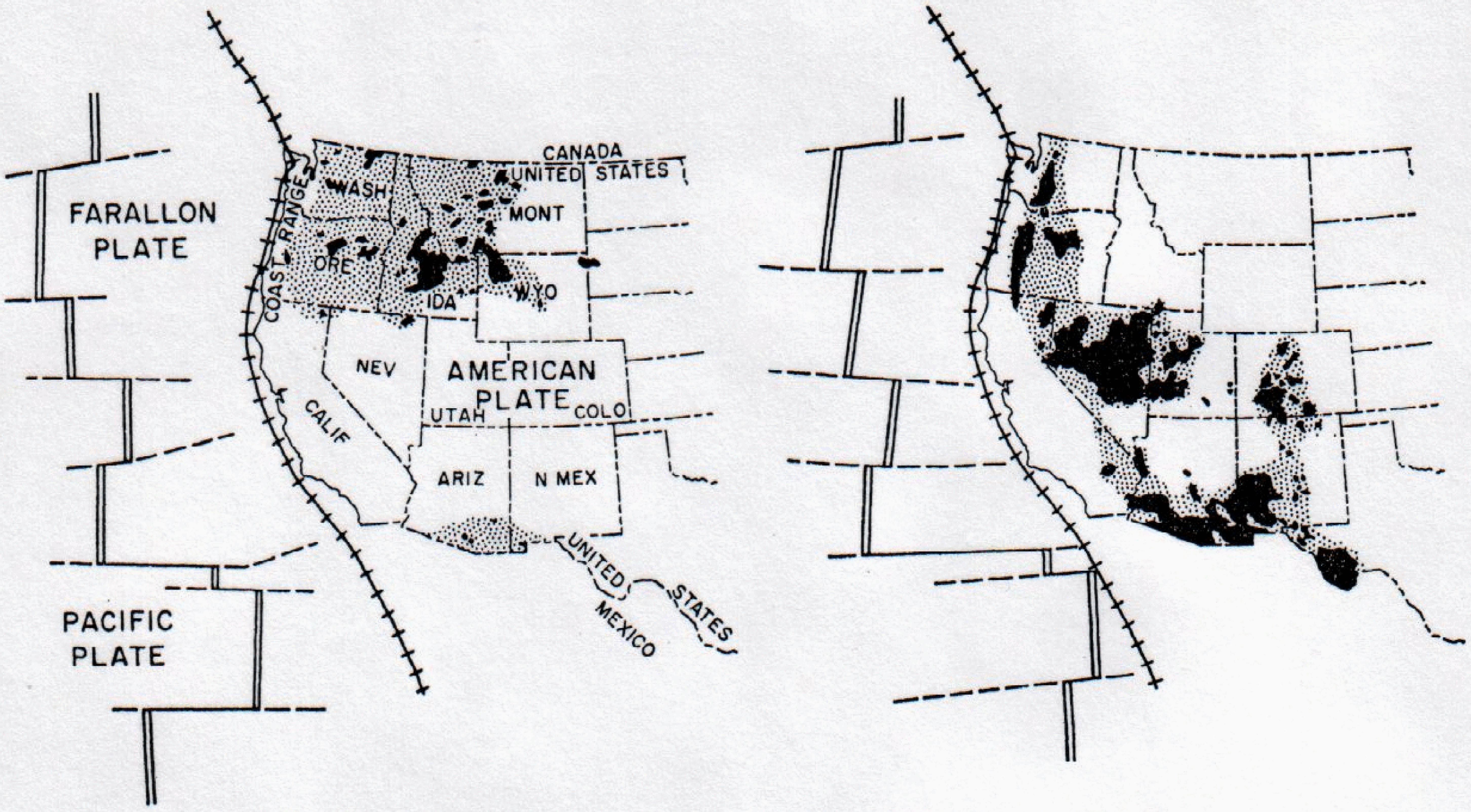
Distribution

Intermediate-composition igneous rocks of Late Cretaceous and Palaeocene age (Laramide) are widely distributed in the Rocky Mountain region, but in many areas data are inadequate to permit confident separation of those rocks from those produced by younger Cenozoic activity. Laramide volcanic rocks, although probably once widespread, are preserved only locally; well-documented areas include western Montana (Robinson, Klepper & Obradovich 1968), northern Washington (Daly 1912), southwestern Colorado (Dickinson, Leopold & Marvin 1968), southern New Mexico (Jones, Hernon & Moore 1967), and southern Arizona (Simons 1964; Drewes 1969; Bikerman & Damon 1966). Laramide intrusive rocks occur in these areas, as well as in others where any formerly associated volcanics have been removed by erosion: for example, in the central Colorado mineral belt (Tweto & Sims 1963), in the Rico, La Plata, and Ute Mountain laccolithic centres of the Colorado Plateau (Armstrong 1969), and in north-central Nevada (McKee & Silberman 1970*b*; Hotz & Willden 1964).

Continental igneous rocks of Eocene age are widespread in the northwestern states (figure 2*a*), and probably once formed a nearly continuous region of coalesced volcanic fields in eastern Washington, Oregon, Idaho, northwestern Wyoming, and much of western Montana. Eocene volcanic rocks also probably extended nearly continuously north through British Columbia and the Yukon Territory of Canada (Souther 1970, figure 3). Scattered Eocene igneous rocks, mainly intrusive, are present in southernmost Arizona and southwestern New Mexico, and probably also in northern Mexico (Ohmoto, Hart & Holland 1968; Damon, Mauger & Bikerman 1964). Eocene igneous activity has not been recognized in the Western United States approximately between latitudes 33° and 41° north. Basaltic marine volcanics of Eocene age are abundantly present in the Coast Ranges of Oregon and Washington (Snavely, MacLeod & Wagner 1968).

The distribution of Oligocene igneous activity is somewhat different (figure 2*b*). In the northwest, Oligocene volcanism is restricted to central Washington and Oregon. Oligocene volcanic rocks are voluminous in Nevada, western Utah, central Colorado, New Mexico, Arizona, and southwest Texas, where little or no Eocene activity has been recorded. Oligocene volcanism also apparently extended along much of eastern California (dates cited by Dalrymple 1963, 1964; Drewes 1963; Olmsted 1968), but age and petrologic data for this region are sparse.

Basin-range faulting began in late Oligocene or early Miocene time over much of the southwest, as discussed in part II of this paper, but pre-basin-range andesitic volcanic suites of Miocene age are present in northwestern and west-central Nevada, in northeastern and east-central California, and in central and eastern Oregon and Washington.



(*a*) Eocene (*b*) Oligocene

FIGURE 2. Approximate distribution of lower and middle Cenozoic igneous rocks in the Western United States and inferred geometry of major crustal plates. The distribution of continental igneous rocks (present distribution, dark pattern; inferred original extent, stippled pattern) is interpreted from the references cited in table 1 and from other sources. Sea-floor basaltic rocks in the Oregon–Washington coast ranges are not shown. The plate geometry is inferred for the times indicated from the magnetic anomaly map of the northeast Pacific (Atwater & Menard 1970), from interpretations of plate motions by Atwater (1970), and from our inferences concerning motions of these plates based on the distribution of Cenozoic igneous rocks. Double lines indicate spreading ridges; single lines, transform faults; 'railway tracks', trench-subduction boundary. (*a*) Distribution of Eocene igneous rocks (40 to 55 Ma) and the plate geometry at about 50 Ma ago (magnetic anomaly 20); (*b*) distribution of Oligocene igneous rocks (25 to 40 Ma) and the plate geometry at about 30 Ma (magnetic anomaly 9).

Absaroka volcanic field

The 24000 km^2 Absaroka field of northwestern Wyoming and southwestern Montana (Hague, Weed & Iddings 1896; Iddings & Weed 1894; Rouse 1937; H. W. Smedes & H. J. Prostka, unpubl. data) consists principally of calc-alkalic andesite, some dacite lava flows and breccias (figure 3*A*), and relatively minor potassic mafic lavas of the shoshonite suite (Iddings 1895; Joplin 1968). Rhyodacitic ash-flow tuffs that are felsic differentiates of the potassic suite constitute only a small percentage of the volume of the field. The roots of composite stratovolcanoes which were the sources of the volcanic rocks are marked by at least thirteen major vent complexes. These vent complexes contain central stocks and plugs of granodiorite, quartz

TABLE 1. AREAS OF EARLY AND MIDDLE CENOZOIC IGENOUS ACTIVITY IN THE WESTERN UNITED STATES (FOR WHICH AGE AND CHEMICAL DATA HAVE BEEN UTILIZED IN THIS PAPER). LOCATIONS OF AREAS ARE SHOWN ON FIGURE 9

locality	rock types	K_2O at 60% SiO_2	age	references
Texas				
1. Davis Mountains	lavas and tuffs of trachybasalt, trachyandesite, and alkali rhyolite. Intrusives of gabbro and syenite	5.0	Oligocene (30–40 Ma)	Goldich & Elms 1949; Erickson 1953; Wilson *et al.* 1968; Dasch *et al.* 1969
New Mexico				
Datil volcanic field:				
2. West (Blue Range)		3.0		Kottlowski *et al.* 1969; Elston *et al.* 1968; Ratté *et al.* 1969; J. C. Ratté written commun. 1970; Jicha 1954; Elston 1957; Tonking 1957; Kuellmer 1954; Jones *et al.* 1967
3. Central (Gila Wilderness)	andesite and rhyodacite flows and breccias; quartz latite to rhyolite ash-flow tuffs	2.9	Oligocene	
4. Southeast		3.2		
5. South (Santa Rita area)		2.7		
6. Organ Mountains	quartz monzonite stock	4.3	27 Ma	Dunham 1935; Kottlowski *et al.* 1969
7. Southern Sangre de Cristo Mountains	andesite and rhyodacite flows, breccias, and intrusives	3.5	Oligocene; one K–Ar date is 35 Ma	C. L. Pilmore written commun. 1970
8. Little Hatchet Mountains	diorite, monzonite, and quartz monzonite intrusives	3.1	Eocene (?)	Lasky 1947; Kottlowski *et al.* 1969
9. Carrizozo area	flows, breccias, and tuffs of andesite; latite stocks; laccoliths of monzonite; syenite dikes; sills of diabase, andesite, latite, trachyte, phonolite, and rhyolite	—	Oligocene	Thompson 1964; Budding 1964; Perhac 1964; Weber 1964; Elston & Snider 1964
10. Cerillos area	flows, tuffs and tuff-breccias of andesite and latite; intrusives of monzonite and latite	—	Oligocene	Disbrow & Stoll 1957; Stearns 1953
Arizona				
11. Ajo area	andesite flows and breccias; quartz diorite and quartz monzonite stocks	—	early to middle Tertiary	Gilluly 1946; Rose & Cook 1966; McDowell 1966
12. Cochise County	andesite to rhyolite; granodiorite and granite intrusives	3.2 (poor)	Oligocene (?)	Gilluly 1956; Erickson 1968
13. Galiuro Mountains	andesite lavas, interlayered with quartz latite to rhyolite lavas and tuffs (may be part of younger association)	2.9	early Miocene (22–26 Ma)	Simons 1964; Krieger 1968
14. Santa Rita Mountains	quartz latite to rhyolite lavas, tuffs, and related intrusives	2.6	Oligocene (26–28 Ma)	Harald Drewes written commun. 1970
15. Sierrita Mountains	andesite lavas	2.8	Oligocene	Harald Drewes & J. R. Cooper written commun. 1970

locality	rock types	K_2O at 60% SiO_2	age	references
Arizona (cont.)				
16. Oatman district	older flows and breccias of andesite, latite and trachyte (probably correlative with Patsy Mine volcanics; may be part of younger volcanic association)	3.4	Miocene	Wells 1937; Ransome 1923; R. E. Anderson 1963; Longwell 1963
17. Yuma area	andesite–basalt lavas; andesite–rhyolite tuffs	—	Oligocene (26–29 Ma)	Olmsted 1968
Colorado				
San Juan volcanic field:				
18. Southeast 19. Northeast 20. Central 21. West	andesite, rhyodacite, and quartz latite lavas with interlayered and overlying quartz latite and rhyolite ash-flow tuffs	3.0 2.9 3.1 3.1	lavas: mostly 35–30 Ma; more silicic types as young as 26.5 Ma. Ash-flow tuffs: 30–36.5 Ma	Larsen & Cross 1956; Steven & Ratté 1960; Steven *et al.* 1967; Burbank 1932; Luedke & Burbank 1968; Varnes 1963; Lipman *et al.* 1970; Lipman 1968; Armstrong 1969; McDowell 1966; Ratté & Steven 1967; P. W. Lipman unpubl. data; Olson *et al.* 1968; Dings 1941
22. Spanish Peaks	gabbro, syenogabbro, diorite, syenodiorite, syenite, granodiorite, granite stocks and dikes	3.6	39.5 Ma (East stock)	Knopf 1936; Johnson 1964, 1968; Armstrong 1969
23. Wet Mountain Valley (Rosita Hills; Silver Cliff)	trachyancesite, latite, rhyodacite, trachyte, rhyolite lavas; diorite and latite intrusives	4.0	38–33 Ma K–Ar dates bracket much of activity	Cross 1896; Siems 1968
24. Thirtynine mile volcanic field	andesite, latite, and rhyolite flows; overlain by and interlayered with rhyolite ash-flow tuffs	3.3	40–34 Ma	Chapin & Epis 1964; Epis & Chapin 1968; R. C. Epis written commun. 1969; Van Alstine 1969
25. Cripple Creek	flows, breccias, and intrusives of latite, phonolite, syenite, and alkalic basalt	5.5	33 Ma	Lindgren & Ransome 1906; McDowell 1966; Koschmann 1960; Loughlin & Koschmann 1935
26. Mt Princeton area	intermediate-composition volcanic breccia and quartz latite welded tuff, intruded by quartz monzonite. Probably a deeply eroded caldera	3.5	35 Ma	Dings & Robinson 1957; Lipman *et al.* 1969; P. J. Toulmin oral commun. 1970
27. Elk and West Elk Mountains	granodiorite stocks, laccoliths, sills, and dikes	2.8	35.5–29 Ma	Godwin & Gaskill 1964; Obradovich *et al.* 1969; Lipman *et al.* 1969
28. San Miguel Mountains	stocks and laccoliths mainly of monzonite, granodiorite, and quartz monzonite	2.5	25.5; 27 Ma (two stocks)	Bromfield 1967; McDowell 1966; Lipman *et al.* 1970
29. Never Summer Mountains	lavas and tuffs of andesite, quartz latite, and rhyolite	4.3	27–28 Ma	Corbett 1965, 1968; Wahlstrom 1944
30. Middle Park	lava, breccia, and tuff; mainly rhyodacite, quartz latite and rhyolite	3.2	33 Ma	Izett 1966, 1968
31. North Park	lava, breccia, tuff, sills, and dikes; mainly andesite, rhyodacite, quartz latite, and rhyolites; capping basalt probably younger	3.0	Oligocene (?)	Hail 1965, 1968; Grout *et al.* 1913

TABLE 1 (*cont.*)

locality	rock types	K_2O at 60% SiO_2	age	references
Utah				
32. Thomas Range	rhyodacite and quartz latite lavas, rhyolite ashflow tuff	2.6	older than 16 Ma	Staatz & Carr 1964; Whelan 1969
33. Iron Springs district	flows and intrusions of andesite, dacite, trachyte, and rhyolite	4.3	Oligocene and Miocene	Leith & Harder 1908; Armstrong 1970; Roberts & Peterson 1961
34. Marysvale	pyroxene andesite–latite porphyry flows; tuffs; monzonite and related intrusives	4.8	25–30 Ma	Callaghan 1939; Bassett *et al.* 1963; Kerr *et al.* 1957
35. Park City–Alta area	andesite flows; dioritic–quartz monzonite intrusives	3.2	31–35 Ma	Boutwell & Woolsey 1912; Crittenden & Kistler 1968; Armstrong 1970
36. Bingham area	monzonite–quartz monzonite stocks; quartz latite porphyry dikes; flows and breccias of andesite, latite, quartz latite, and rhyolite	4.4	32–34 Ma	Moore *et al.* 1968; Stringham 1953; Gilluly 1932; Boutwell 1905
37. Tintic area	andesite, quartz latite, and rhyolite lavas; intruded by monzonite–quartz monzonite stocks	4.2	31–34 Ma	Gilluly 1932; Lindgren & Loughlin 1919; Tower & Smith 1899; Laughlin *et al.* 1919
38. Gold Hill	lavas of andesite, latite, trachyte, and rhyolite; quartz monzonite intrusives	3.8 (poor)	Eocene and Oligocene (?) (42 Ma)	Nolan 1935; Armstrong 1970
39. Henry Mountains	stocks and laccoliths of diorite porphyry and monzonite porphyry	2.1	44 Ma	Hunt *et al.* 1953; Armstrong 1969
40. Abajo Mountains	stocks, laccoliths, dikes, and sills of quartz diorite porphyry	2.2	28 Ma	Witkind 1964; Armstrong 1969
41. La Sal Mountains	laccoliths and stocks of diorite–monzonite porphyry	2.6	23–26 Ma	Hunt 1958; Armstrong 1969; Stern *et al.* 1965
Nevada				
42. Virginia City area	andesite–quartz latite lavas, rhyolite tuff, granodiorite intrusive rocks	2.2	Miocene	Thompson & White 1964
43. Shoshone Range	basalt–andesite lavas; quartz latite welded tuffs, granodiorite intrusives	2.8	30–38 Ma	Gilluly & Masursky 1965; Armstrong 1970; Silberman *et al.* 1969; Gilluly & Gates 1965; McKee & Silberman 1970*a*
44. Osgood Mountains	mostly andesite flows; local rhyolite welded tuffs	3.0	middle Tertiary	Hotz & Willden 1964
45. Cactus Range	calc–alkalic andesite–quartz latite lavas and intrusions	3.0	18–22 Ma	Anderson & Ekren 1968; Kistler 1968
46. Goldfield	flows and tuffs of andesite, dacite, latite, and rhyolite	2.9	Miocene (21 Ma)	Ransome 1909; Albers & Cornwall 1968; Kistler 1968
47. Tonopah	mainly flows and breccias of andesite, quartz latite, and rhyolite; quartz latite and rhyolite welded tuffs	2.9	Miocene (17–22 Ma)	Spurr 1905; Albers & Kleinhampl 1970

locality	rock types	K_2O at 60% SiO_2	age	references
ada (cont.)				
. Eureka	lavas and tuffs of andesite, dacite, quartz latite, and rhyolite	3.3	Oligocene and Miocene (?)	Hague 1892; McKee & Silberman 1970*a*; Blake *et al.* 1969
. Ely	rhyolite lavas; monzonite porphyry intrusives	4.1	Oligocene	Spencer 1917; Blake *et al.* 1969
. Hot Creek—Pancake Ranges	andesite–dacite lavas and breccias	2.3	Oligocene (30–34 Ma)	W. D. Quinlivan, written commun. 1971
. Delamar district	andesite, dacite, and rhyolite lavas and related tuffs; granodiorite and quartz monzonite porphyry intrusives	3.3	Oligocene (28 Ma)	Callaghan 1937; Clarke 1915; Armstrong 1970; Westgate & Knopf 1932
. Roberts Mountains	lavas of andesite and quartz latite; alaskite and rhyolite intrusives	2.6	Oligocene (?)	Merriam & Anderson 1942; McKee & Silberman 1970*a*
fornia				
Sierra Nevada				
. North	basalt and andesite flows and breccias; rhyolite tuffs	1.8	Miocene and Pliocene	Turner 1894, 1896; Dalrymple 1963, 1964
. Central	andesite–dacite flows	2.1	Miocene and Pliocene	Wilshire 1957; Dalrymple 1963, 1964; Slemmons 1966
oming				
Absaroka volcanic field:				
. Southwest	andesite–rhyodacite flows and breccias; diorite plugs and dikes	2.0	Eocene	Hague *et al.* 1899; Shultz 1962; H. W. Smedes and H. J. Prostka, unpubl. data
. Northeast	andesite, latite, shoshonite, absarokite flows and breccias; syenogabbro-syenite plugs and dikes	3.3	Eocene	Chadwick 1970; W. H. Nelson unpubl. data
. Black Hills	rhyolite, quartz monzonite, monzonite, soda–rhyolite, and phonolite intrusives	4.6	Eocene	Noble 1948; Irving 1899; McDowell 1966
. Rattlesnake Hills	phonolite, latite, vogesite, dacite intrusives and minor flows	4.1	Eocene	Carey 1959; Love 1970
ntana				
. Garnet Range	andesite and basalt lava flows and breccias; minor intrusives of andesite, basalt, and dacite	2.5	Eocene	Gwinn 1961; Konizeski 1959; H. J. Prostka unpubl. data
. Lowland Creek area	quartz latite, rhyodacite, and rhyolite flows and pyroclastic rocks	2.2	50–48 Ma	Smedes 1962, 1966; Smedes & Thomas 1965; Becraft *et al.* 1963; H. W. Smedes unpubl. data
. Castle Mountains and Little Belt Mountains	granite, diorite, syenite, lamprophyres; minor basalt and rhyolite flows	3.8	Eocene	Weed & Pirsson 1896*b*
. Crazy Mountains	diorite, granodiorite, gabbro, lamprophyres	3.1	Eocene	Wolff 1938; Larsen 1940

TABLE 1 (*cont.*)

locality	rock types	K_2O at 60% SiO_2	age	references
Montana (cont.)				
63. Highwood Mountains	quartz latite—mafic phonolite lavas; syenite, shonkinite, monzonite, and granite intrusives	4.2	Eocene	Pirsson 1905; Hurlbut & Griggs 1939; Larsen 1940
64. Bearpaw Mountains	intrusive shonkinite, syenite, latite and monzonite; lavas and pyroclastics of mafic phonolite and latite	4.7	Eocene	Schmidt *et al.* 1964; Hearn *et al.* 1964; Weed & Pirsson 1896*a*
Washington				
65. Republic Graben area	andesite, latite, and rhyodacite lavas and tuffs; rhyodacite and quartz monzonite intrusives	2.8	Eocene (54–46 Ma)	Muessig 1967; Parker & Calkins 1964; Staatz 1964; Yates & Engels 1968; J. D. Obradovich oral commun. 1970
66. Mt Rainier area (Ohanapecosh Fm)	andesite volcaniclastic rocks and lava flows; minor basalt, dacite, and rhyolite lavas and tuffs	1.1	late Eocene	Fiske *et al.* 1963; R. S. Fiske written commun. 1970
67. Northern Cascades	andesite, trachyandesite, and dacite flows, breccias and tuffs; quartz diorite, granodiorite, and quartz monzonite stocks	1.2	Eocene (48–13 Ma)	Yeats & Engels 1968, 1970; Tabor *et al.* 1968; Misch 1966; F. W. Cater written commun. 1971
68. Coryell batholith	monzonite, syenite, shonkinite, granite	5.3	50–58 Ma	Daly 1912; Little 1960; Yates & Engels 1968; Baadsgaard *et al.* 1961
Oregon				
69. Western Cascades	basaltic andesite, andesite, and dacite lavas and breccias	1.5	Eocene–Oligocene	Peck *et al.* 1964; Callaghan 1933; Buddington & Callaghan 1936
70. John Day Basin (Clarno Fm)	andesite lavas, breccias, and tuffaceous sediments, minor intrusives	1.3	41.0–34.2 Ma	Evernden & James 1964; Evernden *et al.* 1964; Swanson & Robinson 1968; Robinson 1969; R. E. Wilcox written commun. 1970
Idaho				
71. Challis volcanic field	latite, quartz latite, andesite, rhyolite lavas and pyroclastic rocks; contemporaneous quartz monzonite and granodiorite intrusions	3.9	Eocene	Axelrod 1968; Ross 1962; McDowell & Kulp 1969

monzonite, diorite, syenogabbro, and syenite (Parsons 1939; Iddings 1899; Emmons 1908; Lovering 1929). Peripheral to the deeply dissected volcanoes are well-bedded deposits of epiclastically reworked volcanic debris, comprising well-sorted volcanic breccia, conglomerate, sandstone, and tuff, which compose about half the volume of the present Absaroka field. The preserved area of the Absaroka volcanic field is probably less than half of the original extent of the field.

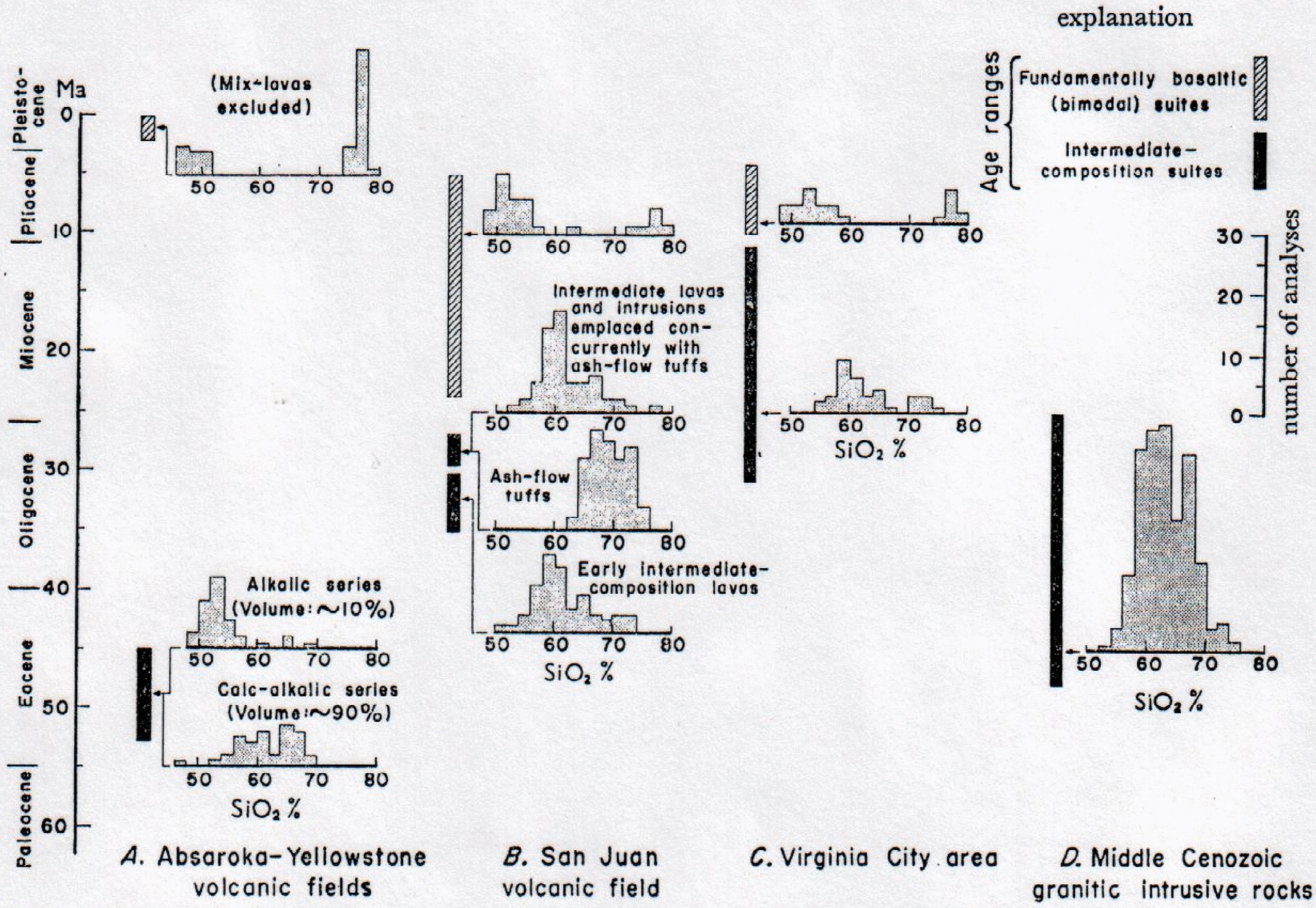


FIGURE 3. Histograms of SiO_2 contents and age ranges of major rock groups of (*A*) the Absaroka–Yellowstone volcanic fields, (*B*) the San Juan volcanic field, (*C*) the Virginia City area, and (*D*) middle Cenozoic granitic intrusive rocks. Interval for SiO_2 is 2%. For sources of analyses see table 1. The frequency distribution of analysed rocks is only an approximate estimate of volumetric proportions of lithologic types, mainly because many geologic studies oversample extreme rock types. For the histograms of this figure a statistically more valid sampling would generally enhance peak amplitudes but would not shift their positions appreciably with respect to SiO_2.

Petrologically, about 90% of the volcanic rocks of the Absaroka field are calc-alkalic andesites and dacites (figure 3*A*), which are very similar to many Pacific-margin andesites, especially the more alkalic ones of Kamchatka (Gorshkov 1958) and Indonesia (figure 5).

Andesites of the Absaroka field become progressively more alkalic to the northeast, across the strike of the field as defined by alinement of vents (Chadwick 1970).

Potassic rocks of the shoshonite suite make up the remaining 10% of the field; these lavas were erupted late but from the same vent complexes that earlier produced the calc-alkalic andesites. In this respect, the Absaroka field is similar to Pacific arc regions—Indonesia, New Guinea, Fiji, and Kamchatka—where shoshonites are also known to occur (Neumann Van Padang 1951; Morgan 1966; Jakes & White 1970; Gill 1970; Gorshkov 1958).

Chemical, petrographic, and isotopic data for the Absaroka rocks indicate that the alkalic lavas could not have been derived by simple differentiation or contamination of a parental calc-alkalic andesite magma (Iddings 1895; Nicholls & Carmichael 1969; Peterman, Doe & Prostka 1970). Rather the andesitic and shoshonitic lavas are thought to have been derived independently by partial melting of the upper mantle at progressively greater depth with time.

San Juan volcanic field

In contrast to the Absaroka volcanic field, silicic volcanics are abundantly associated with andesites in the 25000 km² San Juan field in southwestern Colorado (Larsen & Cross 1956; Lipman, Steven & Mehnert 1970; Steven & Lipman, unpubl. data), which constitutes the

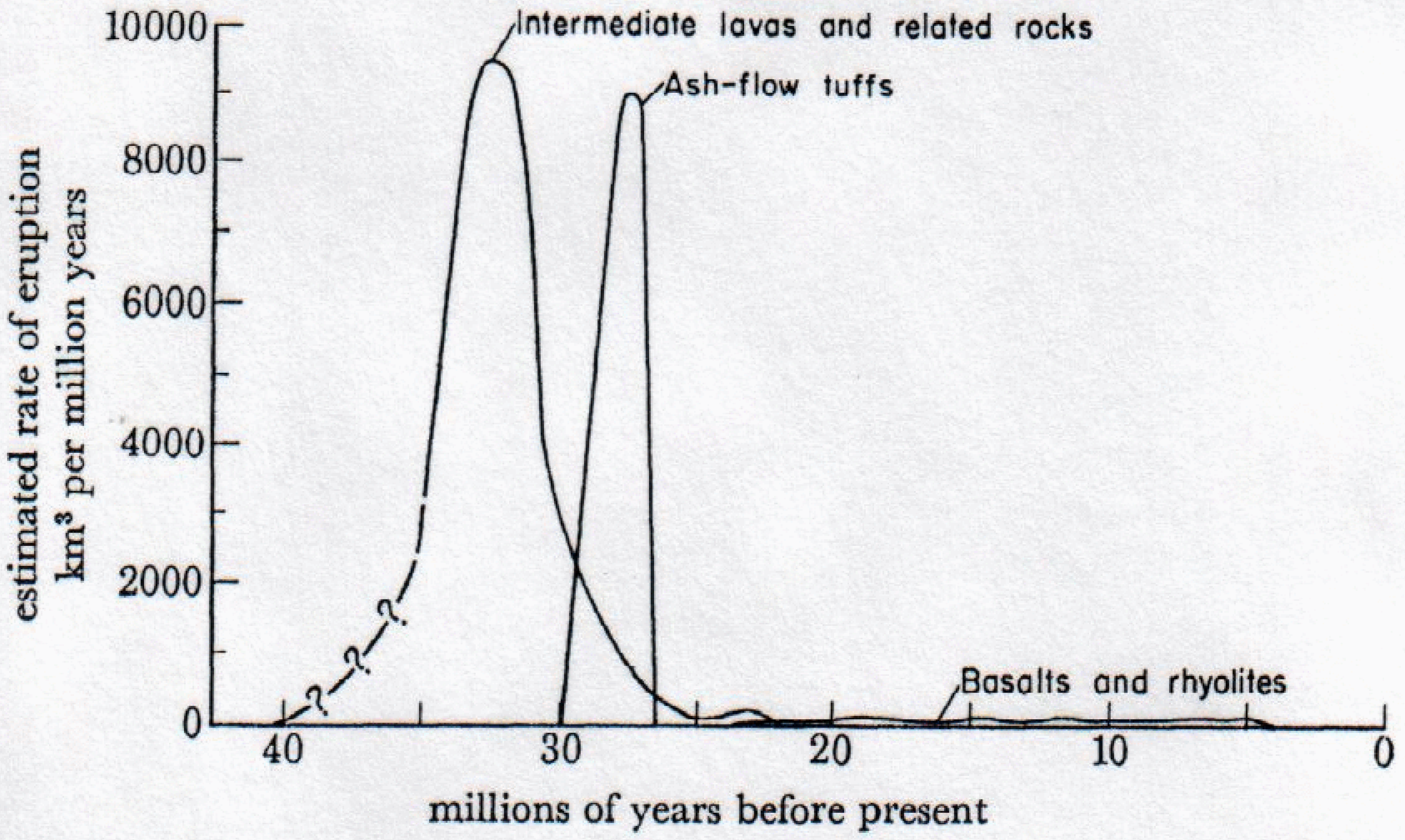


FIGURE 4. Relation between petrologic type, volume, and time, San Juan volcanic field.

largest erosional remnant of a once nearly continuous volcanic field that extended over much of the southern Rocky Mountains in Oligocene time (Steven & Epis 1968). Initial eruptions everywhere in the San Juan field were lavas and breccias, mainly of calc-alkalic andesite, rhyodacite, and mafic quartz latite (figure 3*B*, early lavas). Despite extensive petrologic study, basaltic lavas have not been found within the early sequence, and rhyolitic lavas (maximum SiO_2 content: 73%) are very scarce. The predominantly intermediate composition lavas are widely overlain by ash-flow sheets of quartz latite and low-silica rhyolite (SiO_2 < 74%); eruption of intermediate composition lavas and breccias continued in diminished volume during the ash-flow activity. The lavas and breccias constitute about two-thirds the volume of the field; the ash-flow tuffs about one-third (figure 4).

Considerable evidence suggests that the intermediate composition lavas and the more silicic ash-flow tuffs are comagmatic. Caldera structures resulting from the ash-flow eruptions are within the more extensive region of vents for the early intermediate rocks (Lipman, Steven & Mehnert 1970, figure 1), and the area of calderas coincides with a large negative gravity anomaly, the dimensions of which are compatible with the subsurface presence of a large shallow batholith (Donald Plouff, oral communication, 1970). Non-systematic alternation of rhyolitic and quartz latitic ash-flow sheets indicates that both magma types were available during the period of ash-flow eruptions. Compositional zonations in some individual sheets, from rhyolite upward into quartz latite, represent, in inverse order vertical differentiation of the source magma chambers (Ratté & Steven 1964, 1967), and bridge the compositional range between the lavas and the more silicic tuffs (figure 3*B*). Although most of the intermediate lavas predate the greater part of the silicic rocks, the eruption of some andesitic lavas and the intrusion of intermediate composition stocks during the ash-flow activity indicate that more mafic magmas were available along with their differentiates. For example, one ash-flow caldera was filled immediately after collapse by more than 1200 m of andesitic lava (Lipman & Steven 1970). The caldera complexes thus appear to have formed by collapse resulting from eruption of silicic cupolas over larger, more mafic batholiths. The andesitic and related lavas probably represent the greater part of these batholiths (mainly quartz diorite, granodiorite, and monzonite), and the ash-flow sheets represent their differentiated granitic tops.

Miocene and Pliocene volcanic rocks of the San Juan field are notably different in composition (figure 3*B*). Whereas the Oligocene volcanic rocks are predominantly intermediate lavas and related silicic differentiates, the younger rocks are largely a bimodal assemblage of basalt and high-silica alkali rhyolite that were erupted contemporaneously with the beginning of extensional basin-range faulting in the region. These rocks constitute representatives of the younger volcano-tectonic association discussed in part II of this paper (Christiansen & Lipman, this volume, p. 249).

Available data suggest similar interpretations are valid for other middle Tertiary volcanic fields in which silicic rocks are relatively voluminous. Silicic volcanic rocks, mostly ash-flow tuffs, preceded by or interlayered with intermediate composition lavas, in association with compositionally similar intrusives, occur in the Challis volcanic field of central Idaho (table 1, no. 71), in the Lowland Creek area of western Montana (table 1, no. 60), in west-central Utah (table 1, nos. 32, 34), in central Arizona (table 1, nos. 13, 14), in the Datil field of south-central New Mexico (table 1, nos. 2 to 5), and probably in the Sierra Madre Occidental in Mexico (Mackin & Waitt 1969).

Virginia City area, Nevada

Andesitic activity continued through the Miocene in western Nevada and northeastern California (table 1, nos. 42, 45 to 47, 53, 54), and Pliocene to Quaternary andesites are present in the Cascade Range of Oregon and Washington. In the Virginia City–Steamboat Springs area of northwestern Nevada (Thompson & White 1964); for example, the lithologic sequence—voluminous intermediate composition volcanics followed by a bimodal association of mafic and silicic lavas (figure 3*C*)—is generally similar to that observed for the San Juan field (figure 3*B*), but the age of the petrologic change is distinctly younger. Miocene volcanic rocks in the Virginia City area consist of voluminous andesite and rhyodacite lava flows and breccias (Alta and Kate Peak formations), in association with more silicic ash-flow tuffs (Hartford Hill rhyolite tuff). In contrast, the Pliocene and Quaternary volcanic rocks are a bimodal assemblage of basalt or basaltic andesite (Lousetown formation) and silicic alkalic rhyolite (Steamboat Hills rhyolite).

Intrusive complexes

In many areas middle Cenozoic intrusive rocks are common, but any contemporaneous volcanic rocks that may have been present have been largely or completely removed by erosion; representative examples are the intrusive rocks of the Elk Mountains region of central Colorado (table 1, no. 27), the Colorado Plateau laccoliths (table 1, nos. 28, 39 to 41), and the intrusive rocks of west-central Montana (table 1, nos. 61 to 64). In such areas most of the epizonal intrusive rocks have intermediate compositions, rather similar to those of contemporaneous volcanic rocks in adjacent regions. The mean SiO_2 content of analysed middle Cenozoic laccolithic rocks of the Colorado Plateau is about 62 to 64% (data from sources cited in table 1, nos. 28, 39 to 41), as is the mean SiO_2 content for all analysed middle Cenozoic granitic intrusions of the Western United States (figure 3*D*). The age, sequence, and compositional range of intrusive rocks of the Elk Mountains region, Colorado, is strikingly similar to the volcanic history of the San Juan field to the south (Lipman, Mutschler, Bryant & Steven 1969). The Eocene stocks and laccoliths of west-central Montana consist of both calc-alkalic and alkalic rocks (table 1), which are compositionally similar to associated remnants of extrusive rocks and to the voluminous volcanic rocks of the Absaroka field. These similarities led Larsen (1940) to group both the Absaroka extrusive and the Montana intrusive rocks as the 'petrographic province of central Montana'.

Comparison with circum-Pacific andesites

Calc-alkalic andesites, with associated basalts and dacites, are the characteristic volcanic rocks of active island and continental-margin volcanic arcs around the margins of the Pacific basin; these rocks are associated with plate convergence and subduction where chains of andesitic volcanoes are alined above dipping Benioff seismic zones (Dickinson & Hatherton 1967; Dickinson 1968; Hamilton 1969*b*; Isacks *et al.* 1968).

Systematic chemical variations transverse to individual arcs, especially in alkali contents, are characteristic of these arcs. Rittmann (1953) noted an increase in alkali contents of Indonesian volcanoes across the arc from the trench northward toward the Malay Peninsula. A similar transverse increase of alkalis, especially in basalts, has been shown to occur across the Japanese arc by Kuno (1959), Katsui (1961), and Sugimura (1960), who correlated this increase with depth to the Benioff seismic zone. More recently, the ratio of K_2O to SiO_2 in island-arc andesite suites has been shown to increase systematically with increasing depth to the Benioff zone (Dickinson & Hatherton 1967; Hatherton & Dickinson 1969).

The relatively alkalic andesitic suites of the western interior of the United States, some of which include voluminous associated silicic differentiates, and the calcic or calc-alkalic suites of most island and Pacific-margin arcs represent parts of a continuum of andesitic types rather than defining two discrete groups. Most circum-Pacific andesite suites are less alkalic than the middle Cenozoic igneous rocks of the Western United States, but some are quite similar. For example, Eocene rocks of the Absaroka field, now approximately 1200 km from the edge of the continental shelf, are similar in chemical variations to the active volcanoes from the inner part of the Indonesian arc that are mostly 250 to 350 km from the present trench (figure 5). The plots show appreciable scatter for both volcanic rock suites, but both have similar ranges.

In the central Andes, stratovolcanoes of alkali andesite associated with voluminous rhyolitic ash-flow tuffs (Zeil & Pichler 1967; Pichler & Zeil 1969) constitute an association that is petrologically transitional between more alkalic continental-interior volcanic areas, such as the San Juan field in Colorado, and less alkalic volcanics of continental-margin belts, such as the Aleutian or Cascade chains. Figure 6 compares K_2O–SiO_2 plots of early intermediate composition lavas (figure 3*B*, early lavas) of the Oligocene San Juan field, about 1400 km from the edge of the present continental shelf, with Quaternary andesite of the Chilean Andes that are 300 to 400 km from the axis of the Peru–Chile trench. The San Juan K_2O values are slightly higher than those of the Chilean Andes but are lower than those of the inner Indonesian arc (figure 5). Comparative SiO_2-variation plots for elements other than potassium are even more similar for the Andes and San Juan suites.

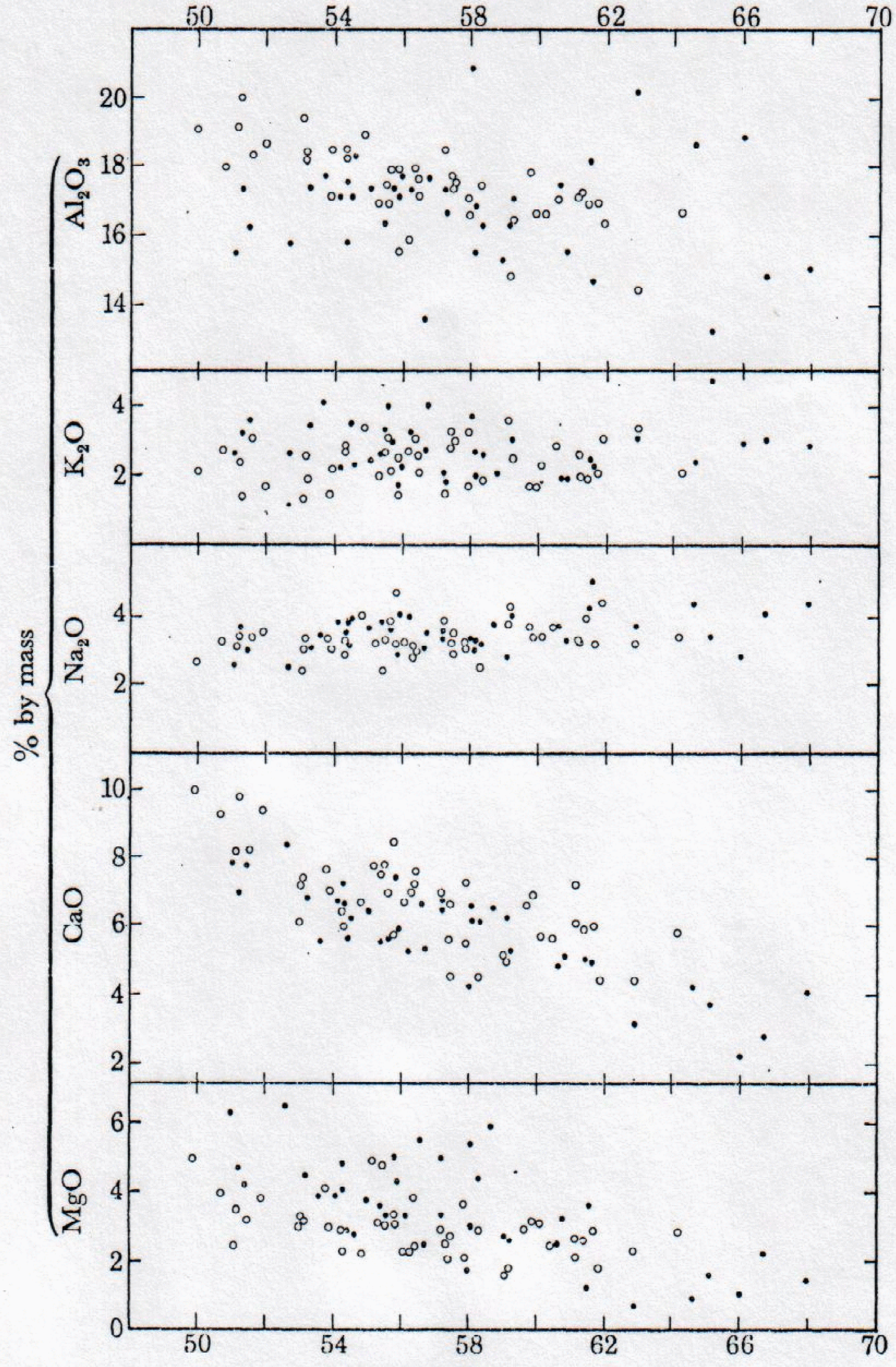


FIGURE 5. SiO_2 (Harker) variation diagrams for rocks of the Eocene Absaroka volcanic field (●) and for inner volcanoes of the modern Indonesian arc (○). Absaroka analyses are from the Sunlight, Crandall, Washburn, Sepulcher, and Gallatin centres (sources of analyses cited in table 1, nos. 55 and 56). Indonesian analyses are those available for volcanoes far from the trench: Sibajak, Sorikmarapi, Talakmau, Marapi, Tankikat, Dempo, Tangkuban Prahu, Dieng, Ungaran, Merbabu, and Bromo (from Neumann Van Padang 1951).

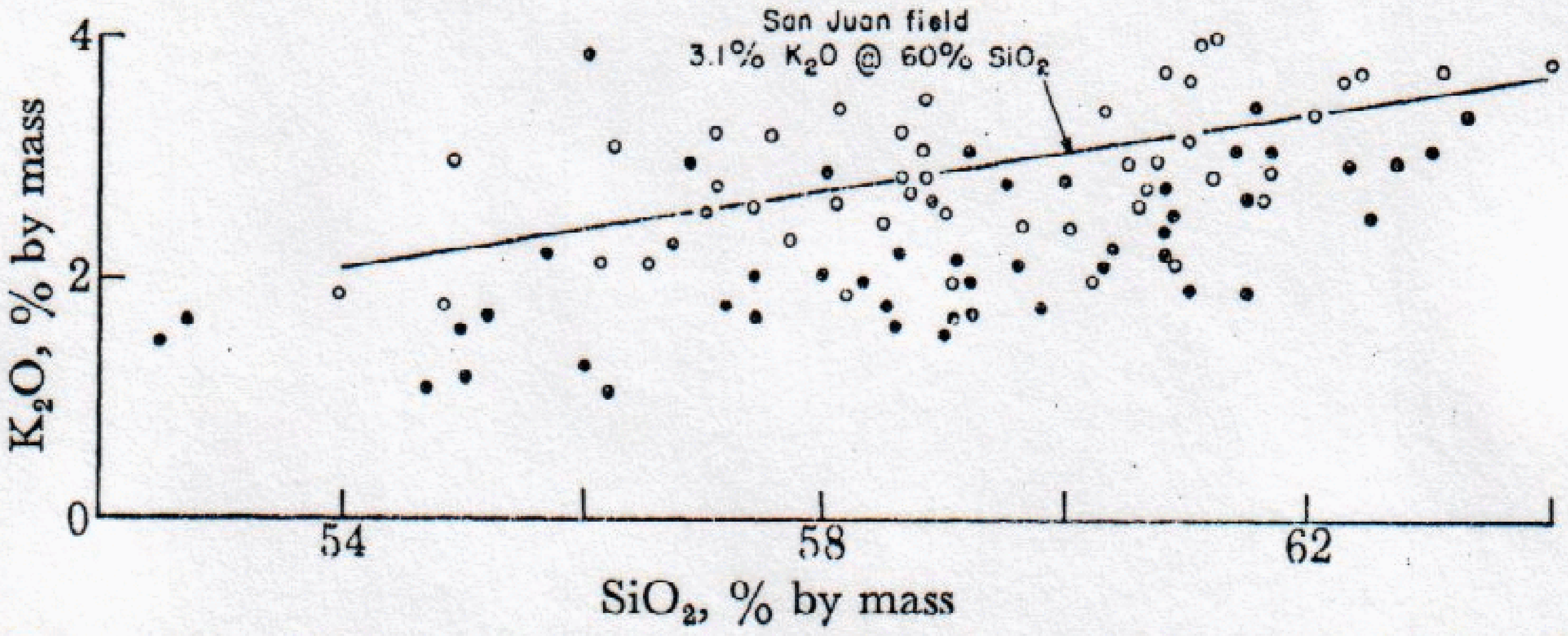


FIGURE 6. K_2O–SiO_2 variation diagram for early intermediate lavas and breccias of the Oligocene San Juan volcanic field (○) and for Quaternary andesites of the Chilean Andes (●). Trend line drawn by inspection through points for the San Juan volcanic field. San Juan analyses from sources cited in table 1, nos 18 to 21; Chilean analyses from Pichler & Zeil (1969), Katsui & Gonzales-Ferran (1968), Gonzales & Vergara (1962), Zeil & Pichler (1967), Thiele & Katsui (1969), Oyarzun & Villalobos (1969), Vergara (1969) and Vergara & Katsui (1969).

Significant regional petrologic variations, especially in alkali contents, have also been recognized for many years among Cenozoic igneous rocks of the Western United States. Merriam & Anderson (1942, figure 3) showed that the ratio of potassium to sodium in Cenozoic volcanic rocks increases fairly systematically from California eastward to Utah. Moore (1962) made contour maps of potassium–sodium ratios for Cenozoic igneous rocks of the Western United States, which also indicated general eastward increase of potassium, as well as a close correspondence to variations in Bouguer gravity maps. In addition, upper Mesozoic batholithic rocks of the Western United States generally shown an increase in K-feldspar relative to plagioclase from the Pacific Coast eastward (Lindgren 1915; Buddington 1927; Moore 1959).

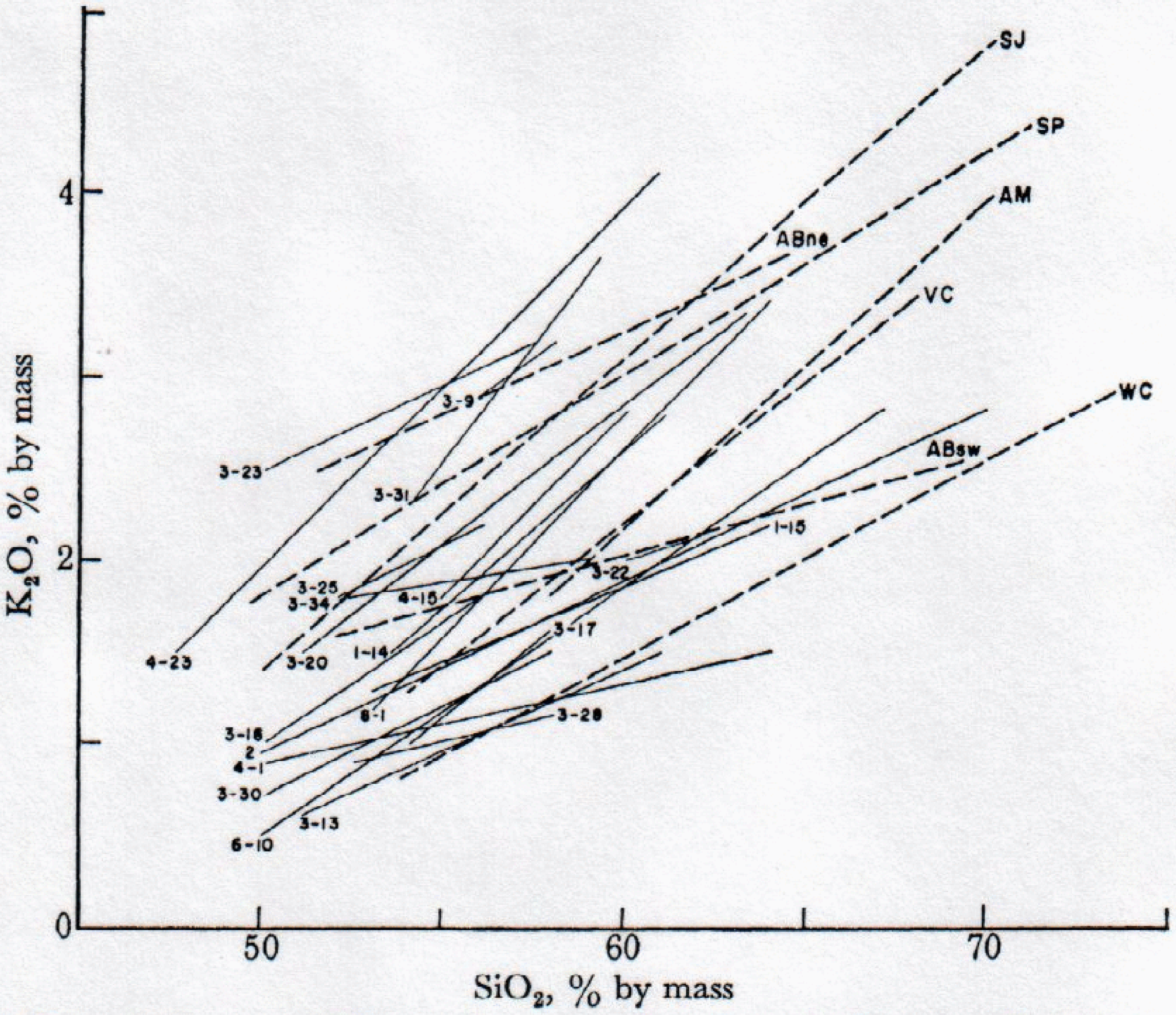


FIGURE 7. K_2O–SiO_2 variation trend lines for igneous suites from the Western United States and Indonesia. Western United States trends (– – –) are for the San Juan field (SJ), the southwestern (ABsw) and north-eastern (ABne) belts of centres in the Absaroka field, the Virginia City area (VC), the Spanish Peaks centre (SP), the western Cascade Range (WC), and the Abajo Mountains laccoliths of the Colorado Plateau (AM); for sources of data see table 1. The 20 Indonesian trends (——) are for every individual volcano listed in the *Catalogue of active volcanoes* (Neumann Van Padang 1951), for which four or more analyses are available; numbers are indices of volcanoes in the *Catalogue*. All trend lines were drawn by inspection through scatter diagrams, as illustrated in figure 6.

In conjunction with these lateral variations in alkali contents, lithologic associations locally deviate significantly from the general intermediate compositional character of the early to middle Cenozoic igneous activity. For example, continental margin calcic andesitic chains, such as the western Cascades, contain considerable basalt but relatively minor volumes of silicic rocks (Peck *et al.* 1964), in contrast with calc-alkalic continental interior fields, such as the San Juans, where basalt is rare and silicic differentiates are abundant (figure 3*B*). Alkalic foreland volcanic rocks of the continental interior, such as the west-central Montana centres, also include abundant mafic igneous rocks (table 1, nos. 61 to 64), as do the alkalic suites of the Absaroka field (figure 3*A*), and the somewhat bimodal alkalic volcanic rock associations of southwestern Texas (table 1, no. 1). Although the differentiated alkalic suites of these foreland centres therefore begin to resemble late Cenozoic suites that we describe as fundamentally basaltic, the foreland volcanism is part of a compositional spectrum of contemporaneous igneous activity that regionally is predominantly intermediate in composition. This is especially evident for the Absaroka field.

If K_2O–SiO_2 variation trends are plotted for a wide variety of Cenozoic igneous suites from the western interior United States (figure 7), they fall within the family of trends defined by circum-Pacific andesite suites, such as Indonesia, that are clearly related to active seismic zones and subduction tectonics (Hatherton & Dickinson 1969). Thus, the similarities in compositional ranges and in systematic regional variations of alkali contents, especially potassium, suggest that the middle Cenozoic andesitic rocks of the Western United States, like modern andesites around the Pacific basin, were genetically related to a subduction system.

A subduction system is also indicated for Western North America in middle Cenozoic time by the sea-floor magnetic anomalies of the northeast Pacific. Although the present boundary between the American and Pacific plates (figure 1) consists of an oblique transform system that separates two segments of the East Pacific Rise from Baja California to the Mendocino fracture zone (figure 1), the preservation of virtually continuous magnetic anomalies related to sea-floor spreading as young as anomaly 9 (approximately 30 Ma ago) requires that the East Pacific rise formed a continuous spreading system until late in Oligocene time (McKenzie & Morgan 1969; Atwater 1970). These anomalies are preserved only in the Pacific plate west of the East Pacific rise, but their presence requires that the East Pacific rise was a symmetrical spreading system and therefore that a continuous plate existed between the East Pacific rise and North America until late Oligocene time (figure 2). As only fragments of this plate, somewhat inappropriately designated the 'Farallon plate' by McKenzie & Morgan (1969),† are still preserved as the Cocos plate off Mexico and the Juan de Fuca plate off Oregon and Washington, the intervening portion must have been resorbed into the mantle along a Cenozoic subduction boundary with the American plate.

The former presence of a subduction system between the Farallon and American plates in late Mesozoic time can also be inferred from stratigraphic and structural features of the Franciscan terrane in California (Page 1969, 1970; Ernst 1970; Bailey, Blake & Jones 1970), as well as from features of the late Mesozoic batholith belt of western North America (Hamilton 1969*a*, *b*; Dickinson 1970). Geologic features in the California Coast Ranges suggest that this subduction system may have been active until Oligocene time (Page 1970).

Extrapolation of a plate-tectonic model of constant motions back to Mesozoic and early Cenozoic time led Atwater (1970, figure 18) to infer an earlier transform plate boundary at the western edge of the United States, before development of the subduction boundary in this region. Although our interpretation, based on the continental volcanic record, is in excellent agreement with the constant-motion plate model for middle and late Cenozoic time, the constant-motion extrapolation to earlier times is difficult for us to reconcile with the presence of subduction-type calc-alkalic igneous rocks of Late Cretaceous and early Cenozoic age in the northwestern United States and western Canada (Tabor, Engels & Staatz 1968; Yeats & Engels 1970; Yates & Engels 1968; Forbes & Engels 1970; Menzer 1970; Souther 1970).

† The Farallon Islands, from which the name was derived, consist mainly of Mesozoic granitic rocks that were within the American plate at the time that the 'Farallon plate' existed, but now these islands are within the Pacific plate.

Geometry of the subduction system

We have attempted to reconstruct the geometry of the inferred middle Cenozoic subduction system from chemical variations in igneous rocks of the same age. In the circum-Pacific regions of active subduction-related volcanism, the K_2O–SiO_2 ratios of andesitic rocks from many different areas have been shown to increase systematically with increasing depth to the Benioff seismic zone that marks the subduction boundary (Dickinson & Hatherton 1967; Hatherton & Dickinson 1969). If this relationship (figure 8) were also valid for former subduction zones (and we see no reason why it should not be), it would provide a method for estimating the depths to the inferred middle Cenozoic subduction system of western North America.

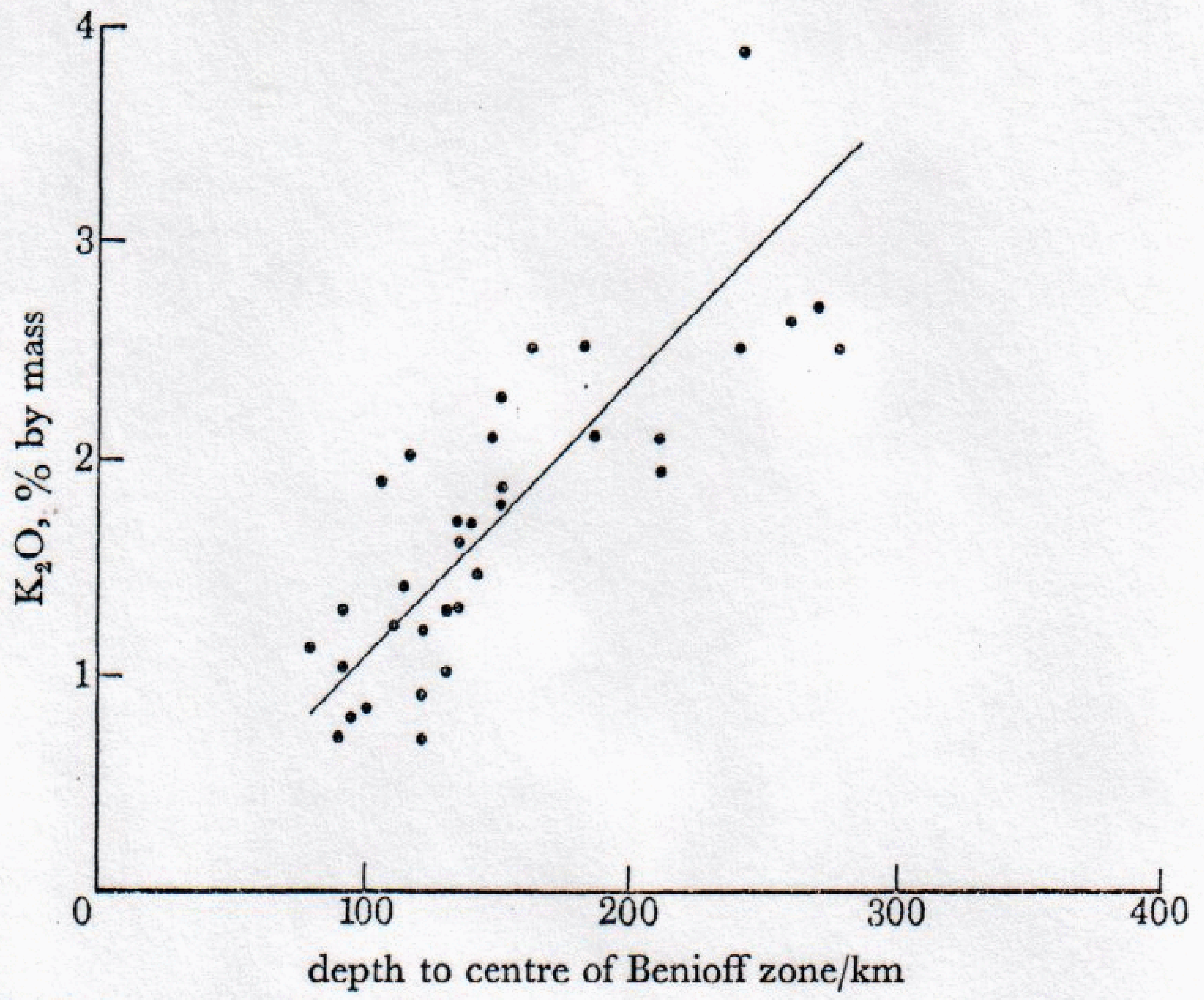


FIGURE 8. Variation in K_2O content (at 60 % SiO_2) of andesites from recently active arc volcanoes with depth to the centre of the inferred Benioff seismic zone beneath the volcanoes. Straight line is least-squares best fit (from Hatherton & Dickinson 1969, figure 6).

Accordingly, we have estimated, from simple SiO_2 variation diagrams, the K_2O contents at 60 % SiO_2 for about 70 dated middle Cenozoic volcanic rock suites for which adequate chemical data are available (table 1). Also included are K_2O contents for some areas of Miocene volcanism that predate the local beginning of basin-range extensional faulting in parts of Arizona, Nevada and California. The way K_2O content is estimated from a variation diagram is illustrated by figure 6. Obviously, there is a substantial range in the quantity and quality of data available from area to area. We have not used plots in which paucity of data or excessive scatter of points makes it difficult to define K_2O at 60 % SiO_2 within ± 0.3 %. For most plots, a much smaller shift than this in position of the trend line results in a clearly unsatisfactory fit.

The K_2O contents at 60 % SiO_2 (table 1) have been converted into depth values by reference to figure 8, plotted on a map of the Western United States and contoured at 50 km intervals (figure 9). The contours were drawn to fit points within ± 10 km depth, equivalent to approximately ± 0.1 % K_2O, which we regard as the probable precision of well-defined points. Two points, for the Thomas range in Utah and the Hot Creek–Pancake ranges in Nevada (table 1, nos. 32, 50), are anomalously low in K_2O and in inferred depth; these points have been disregarded in the contouring, although they may have special significance, as discussed later. Otherwise, the plot shows two well-defined trends of eastward-increasing depths, broken by a north-trending discontinuity across which contours cannot easily be connected. This discontinuity extends from the northern Rocky Mountains, south along the Wasatch front, to the

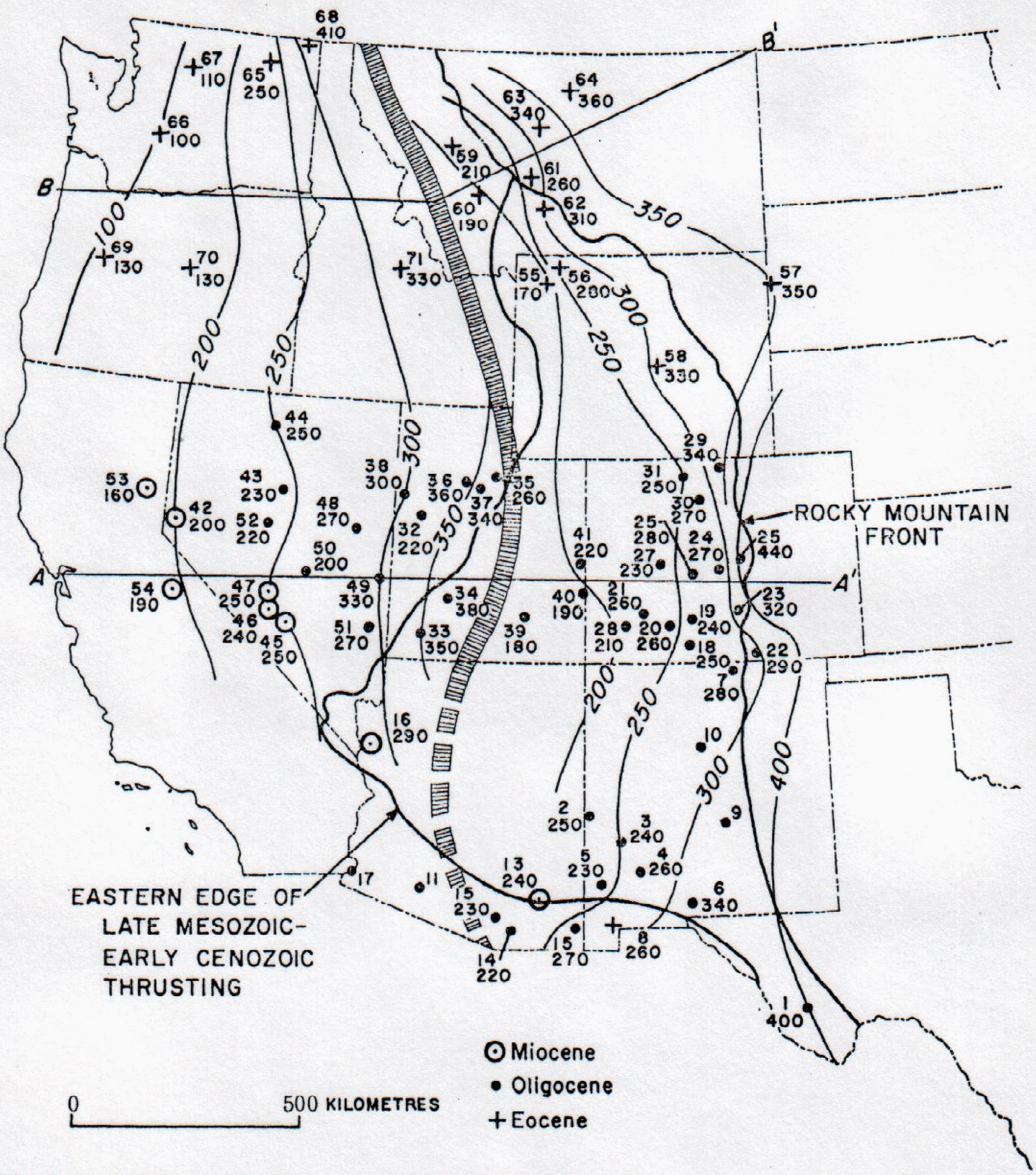


FIGURE 9. Contoured depths (km) to inferred middle Cenozoic subduction zones of Western United States. Data points from table 1: upper number at each point identifies locality; lower number gives depth to subduction zone, calibrated to the K_2O–depth plots of Hatherton & Dickinson for modern island-arc volcanic rocks. (See figure 8.) Hachured line indicates discontinuity in contours. Sections *A–A′* and *B–B′* shown on figure 10.

western Colorado Plateau; it coincides approximately with the east edge of the belt of late Mesozoic-early Tertiary thrusting (figure 9; Christiansen & Lipman, this volume, figure 4). The discontinuity in depth values is at a maximum in the west-central United States and becomes less pronounced toward the Canadian border as the width of the volcano-tectonic belt narrows (figure 1). In southwestern Canada, Cenozoic igneous rocks are absent east of the projected trend of the discontinuity (approximately coincident with the Rocky Mountain trench). Petrologic descriptions—chemical data are scarce—suggest that the discontinuity persists southwest into or through Arizona but not far into Mexico.

A general eastward increase in depth could have been predicted from the earlier studies that indicated eastward increase in potassium (Merriam & Anderson 1942; Moore 1962); however, the discontinuity is obscure on Moore's (1962, figures 2, 3) contour maps, because these maps are based on data from upper Cenozoic igneous rocks as well. Only in recent years have sufficient radiometric age data become available to make possible an attempt at systematic subdivision of Cenozoic igneous activity by age.

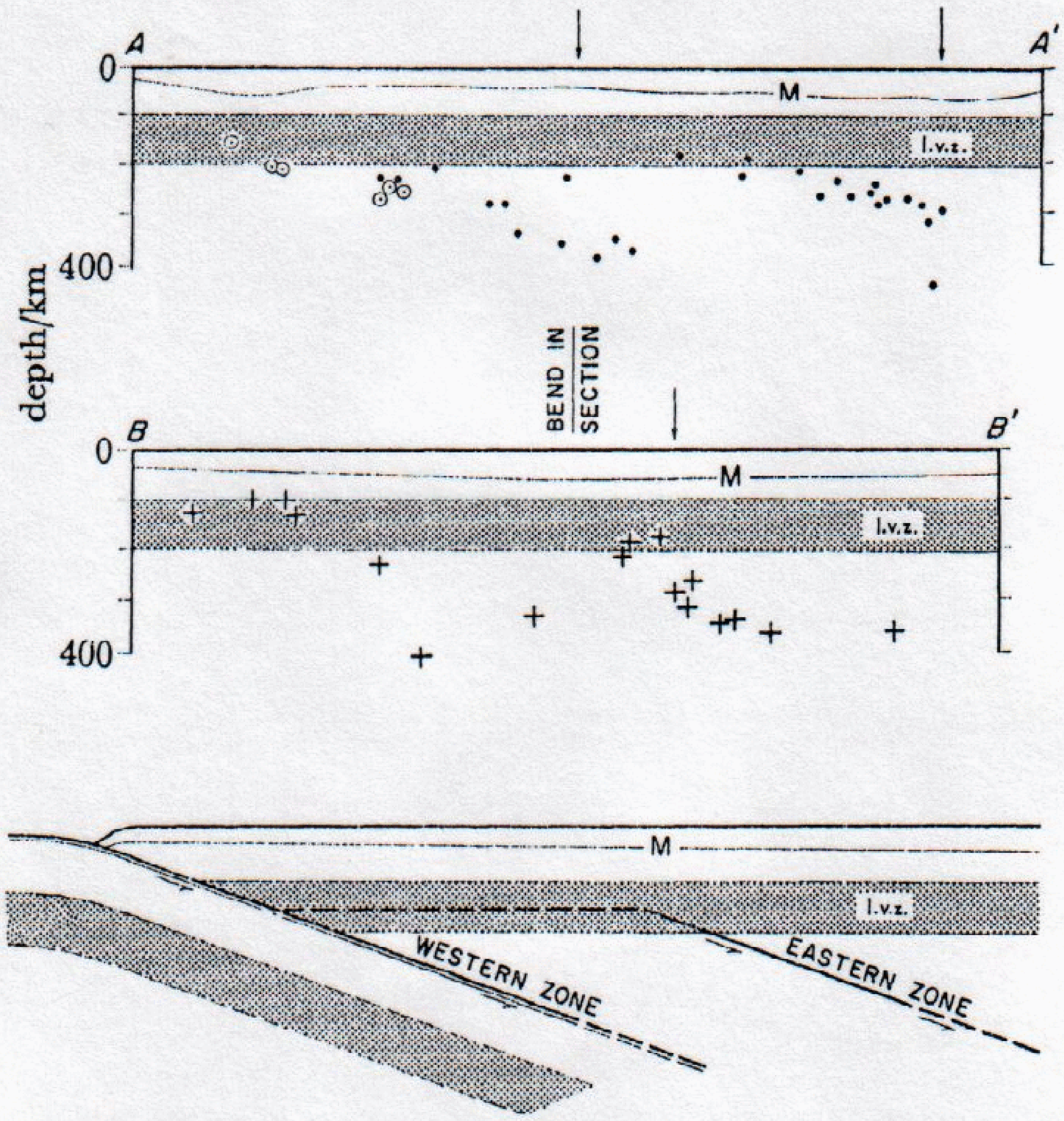


FIGURE 10. Cross-sections of inferred middle Cenozoic subduction zones across the Western United States. Locations of sections shown in figure 9. Horizontal scale equals vertical scale; surface topography not shown. All Oligocene (●) and Miocene (⊙) points within 100 km north or south of section *A–A′* are projected into this section. All points (+) for Eocene volcanic rocks in the northwestern United States are projected into section *B–B′*. M, Mohorovicic discontinuity (from Pakiser & Zeitz 1965); l.v.z., approximate location of low-velocity zone (Gutenberg 1959); arrows, structural-province boundaries from figure 9. The bottom section presents our schematic interpretation of the lithsopheric plate motions suggested by the data plotted in sections *A–A′* and *B–B′*.

In cross-section projections the depth values define two parallel zones that dip eastward at about 15–20° (figure 10), with the dips of the Eocene zones in the northwest seemingly slightly steeper than dips of Oligocene and Miocene zones in the west-central region. Because the area of the sections has been extended at least 100 km by late Cenozoic basin-range faulting and related extensional deformation (Hamilton & Myers 1966), middle Cenozoic dips of the subduction zones that are defined by these points should have been slightly steeper, probably 20 to 25°. These plots suggest that middle Cenozoic subduction of the Farallon plate beneath North America was along gently dipping imbricate zones.

The inferred 20 to 25° dip for subduction is more gentle than the dips of carefully studied active Benioff seismic zones of oceanic arcs such as Tonga-Kermadec (40 to 60°; Isacks, Sykes &

Oliver 1969), Izu-Mariana (50 to 70°; Katsumada & Sykes 1969), New Zealand (45 to 60°; Hamilton & Gale 1968), and Indonesia (45 to 55°; Hatherton & Dickinson 1969). Continental-margin seismic zones, however, seem to be less steep than island-arc zones, as first suggested by Benioff (1954), who distinguished between seismic zones related to oceanic faults and those related to continental-margin faults. Published earthquake-foci diagrams suggest that the seismic-zone dips flatten from 50 to 70° under the Mariana arc (Katsumada & Sykes 1969) to only 20 to 25° under the semi-continental crust of northern Japan (Wadati & Iwai 1956; Katsumada 1956). Although detailed earthquake-foci diagrams have not been published for the Aleutian arc, unpublished data that indicate a dip of only 17° under the Katmai peninsula are cited by Dickinson & Hatherton (1967, table 1). This area is at the eastern continental margin end of the Aleutian arc; steepening of the subduction system to the west, where the arc becomes oceanic, is suggested by decreasing distance from trench axis to volcanic front (King 1969).

The western subduction zone of middle Cenozoic North America should have emerged at the surface as a trench, in much the same way as the Peru–Chile trench parallels the coast of South America today. The inferred imbrication of the subduction zones beneath the Western United States has no middle Cenozoic surface tectonic expression that we recognize, other than volcanism, although the line of discontinuity between the two subduction zones coincides approximately with the eastern edge of the major late Mesozoic-early Cenozoic thrust belt (figure 9). There is no major through-going structural feature suggestive of a middle Cenozoic continental analogue of an oceanic trench. For this reason, we think that the imbricate down-going plates must have been largely decoupled from the overlying continental plate and that the eastern zone never emerged at the surface. The shallowest depths of the inferred eastern subduction zone to which volcanism was related are 175 to 200 km (figures 9, 10). These depths correspond with the location of the low-velocity layer of the asthenosphere (Gutenberg 1959, p. 84; Lehmann 1967), a plausible horizon along which decoupling might have occurred. The two anomalously shallow depth values cited earlier, from volcanic rocks of the Thomas range in Utah and the Hot Creek–Pancake ranges in Nevada (table 1, nos. 32, 50), plot close to the low-velocity layer in the region between the two dipping zones (figure 10, *A–A′*), and the chemistry of these volcanic rocks may be a reflection of the zone of decoupling. The cross-section at the bottom of figure 10 schematically illustrates our interpretation of the plate boundaries, with decoupling in the low-velocity layer.

No modern imbricate seismic-zone system has yet been recognized by seismologists beneath a continental plate. An oceanic feature that may be analogous in some respects is represented by the parallel Mariana and Philippine arcs, both with west-dipping subduction zones separated by the Philippine Sea. The schematic interpretation of this area by Dewey & Bird (1970, figure 2*c*) resembles our interpretation of middle Cenozoic plate relations in the Western United States (figure 10, bottom).

The presence of an imbricate, gently dipping subduction system under the Western United States in middle Cenozoic time provides a plate-tectonic interpretation for the conspicuous bulge that otherwise appears anomalous (Gilluly, 1971) in the continental margin belts of igneous and tectonic activity (figure 1). Similar reconstructions of paleo-subduction zones of other ages permit inferences regarding the geometric evolution of subduction systems along the western margin of North America through late Mesozoic and Cenozoic time.

Jurassic to Late Cretaceous igneous activity is confined to a relatively narrow region of the Western United States (Gilluly 1965, figures 4, 5). Granitic rocks as young as about 80 Ma are present in the Sierra Nevada batholith (Evernden & Kistler 1970), but Mesozoic igneous rocks of this age and older are sparse east of central Nevada. A steep subduction zone (about 50°) has been inferred for the Sierra Nevada region in late Mesozoic time (Dickinson 1970), on the basis of chemical data on batholithic rocks published by Bateman & Dodge (1970). These authors and Hamilton (1969*a*, *b*) also discuss the possible existence of a subduction zone during emplacement of the Sierra Nevada batholith. Although Dickinson's interpretation of a steep subduction zone is based on rocks whose ages range from Early Jurassic to Late Cretaceous, the inference of a steep subduction zone is supported by the narrowness and linearity of granitic sequences of more restricted age ranges within the composite batholith (Evernden & Kistler 1970). In contrast, middle Cenozoic igneous activity associated with gently dipping subduction zones extends over broad regions (figure 2).

Eastward shift of igneous activity in latest Cretaceous and early Cenozoic time was first postulated by Lindgren (1915). In the west-central United States Upper Cretaceous igneous rocks older than about 80 Ma are in California and western Nevada; latest Cretaceous (Laramide) igneous rocks (70–65 Ma) are in north-central Nevada (McKee & Silberman 1970*b*) but also extend as far east as the southern Rocky Mountains region (Pearson, Tweto, Stern & Thomas 1962). Our attempts to reconstruct the geometry of the late Cretaceous–early Cenozoic subduction zone from the chemistry of Laramide igneous rocks have been hampered by a paucity of chemical data and by the extensive alteration of many of the analysed rocks of this age. The relatively few Laramide igneous suites, for which seemingly reliable compositional trends can be drawn and subduction-zone depths inferred, fit fairly closely to the trends defined by middle Cenozoic igneous suites (figure 9). As early as about 70 Ma, Laramide igneous activity extended as far east as the subsequent middle Cenozoic volcanism (Gilluly 1965, figures 6, 7). These variations in chemistry and distribution of the igneous rocks, as well as in inferred depths to the subduction zone, suggest that the major shift in the geometry of subduction occurred fairly abruptly within about 10 Ma in Late Cretaceous time (between about 80 and 70 Ma ago).

Concurrently with the shift in distribution of igneous activity, an eastward shift in orogenic activity occurred, which was marked by initiation of Laramide foreland deformation (Sales 1968). Beginning of Laramide uplift in the central Rocky Mountains, as marked by change in sedimentary depositional environment from marine shale (Pierre Shale) to marine and continental sandstone (Fox Hills, Lance, and Laramie Formations), occurred at about the beginning of Maestrichtian time (about 70 Ma ago: Gill & Cobban 1966; G. R. Scott 1963 and oral communication 1971). We suggest that this change in distribution of volcanism and tectonism, which would be enigmatic in terms of a steeply dipping subduction boundary, as pointed out by Gilluly (1971), is a plausible consequence of imbrication and flattening of the subduction zone.

Although the middle Cenozoic discontinuity (and also a probable discontinuity of Laramide age) related to the imbricate subduction zones is near the eastern edge of the late Mesozoic–early Cenozoic thrust belt, the coincidence is only approximate (figure 9). The eastern subduction zone would project to the surface several hundred kilometres farther west (figure 10), within a region of exposed Cretaceous metamorphism in the eastern great basin (Armstrong & Hansen 1966) and the northern Rocky Mountains (Hamilton 1963). The boundary between this metamorphic region and rigid Precambrian basement to the east, which marked the eastern edge of major orogenic activity in Mesozoic time, may have helped control the location of the eastern subduction zone as it developed in Late Cretaceous time. Development of the thrust belt, which began at least as early as Jurassic time (Armstrong & Oriel 1965; Armstrong 1968; Price & Mountjoy 1970), probably is related primarily to the steep Mesozoic subduction system to the west (Hamilton 1969*a*), although thrusting continued and expanded eastward in Late Cretaceous and Paleocene time concurrently with development of the imbricate subduction system. The absence of major middle Cenozoic thrusting or foreland deformation in the Rocky Mountain region, despite voluminous subduction-related igneous activity at this time, suggests that, as the subduction system flattened, decoupling along the low-velocity layer (figure 10, bottom) became more effective.

The imbricate subduction hypothesis also provides a possible interpretation of certain puzzling features of the Colorado Plateau (Gilluly 1963). The Plateau area was a distinctive region of tectonic stability during Laramide folding, thrusting, and uplift of the surrounding region in latest Cretaceous and early Tertiary time and also during the basin-range extensional faulting of adjacent areas in late Cenozoic time (Gilluly 1963, figures 10, 18). In presently active plate-convergence systems of the Pacific basin, a nonvolcanic region of tectonic stability, 75 to 275 km wide, separates the oceanic trench that marks surface emergence of the subduction zone and the first andesitic volcanoes (the volcanic front of Sugimura, 1960). This stable region has been termed the 'arc-trench gap' by Dickinson (1971). Perhaps the relative tectonic stability of the Colorado Plateau represents an intra-continental equivalent of the 'arc-trench gap', related to shallow parts of the eastern subduction zone of the imbricate system that was active from Late Cretaceous to middle Cenozoic time. The present high topographic position of the Colorado Plateau is largely due to uplift in the late Cenozoic (Gilluly 1963; Hunt 1956, 1969); such uplift occurred widely in the Western United States in late Cenozoic time, but those areas characterized by earlier igneous activity and tectonism were attenuated by basin-range extensional deformation. We suggest in part II of this paper that this regional uplift may be related to disruption of a dynamic equilibrium maintained by the wide early and middle Cenozoic subduction system.

Reconstruction of the inferred Pleistocene palaeoseismic zone under the Oregon and Washington Cascade volcanic chain (Dickinson 1970) indicates dips of 40 to 50°. We suggest that the low-angle imbricate Eocene subduction system in the northwest (figure 10) was replaced in early Oligocene time by a steeper subduction zone that was active until recently. This shift is reflected by the termination of andesitic volcanism in the continental interior (Montana, Wyoming, Idaho) about 40 Ma ago (figure 2), and by the continuation of andesitic volcanism in the Cascade region through Quaternary time (Peck *et al.* 1964, figure 35). In southern parts of the Cascade range the belts of volcanic vents have shifted slightly eastward from early Oligocene to Quaternary time (Peck *et al.* 1964, figure 36), but compositions of andesitic rocks seem to have remained nearly constant (Peck *et al.* 1964, figure 28). For the Cascade range as a whole, the trend of late Tertiary plutons is slightly oblique to that defined by Quaternary volcanoes (McBirney 1968, figure 2). These relations suggest that the depth of magma generation may have been nearly constant and that the shifts in igneous activity reflect minor continued adjustments of the subduction zone.

In the west-central and southwestern United States the low-angle imbricate subduction system continued to operate through Oligocene time, as indicated by the distribution of Oligocene andesitic volcanism in these regions (figure 2*b*).

Termination of plate convergence

Quaternary andesitic volcanism of the type characterizing plate convergence is almost restricted to the Cascade volcanoes adjacent to the Juan de Fuca plate, and to the volcanoes of central Mexico that are opposite the active East Pacific rise and middle America trench. In between, from Baja California to the Mendocino fracture zone, the American and Pacific plates are presently in contact along an oblique transform boundary system (figure 1); in this gap Quaternary plate subduction and andesitic volcanism did not occur. Intermediate-composition volcanism terminated in the southwestern United States at about the end of the Oligocene, although voluminous andesitic rocks of the same type were erupted in Miocene time over much of western Nevada and adjacent parts of eastern California, and through the Quaternary in the Cascade Range. Initial intersection of Pacific and American plates also occurred at about the end of Oligocene time (figure 2; Atwater 1970). We interpret the enlarging gap in the late Cenozoic andesitic volcanic belt of western North America as the reflection of the growing zone along which the American and Pacific plates were in contact, and along which the intervening Farallon plate had been totally consumed and the bounding trench replaced by a transform boundary system. This shifting pattern of volcanism and its relations to changing plate-tectonic boundaries are the topic of part II of this paper (this volume, p. 249).

We have benefited greatly from discussions with participants of several meetings at which our interpretations were presented in preliminary form: the 1969 Symposium on Andesites at the Geological Society of London, the 1969 Geological Society of America Penrose Research Conference on Implications of Plate Tectonics for Orogenic Processes, and the 1970 Cordilleran Section meeting of the Geological Society of America. In addition, we are indebted to colleagues on the U.S. Geological Survey who have provided access to unpublished data. Our thinking has been especially stimulated by discussions with Warren Hamilton of the U.S. Geological Survey and William R. Dickinson of Stanford University, who also helpfully reviewed the manuscript.

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