Plutonism and Orogeny in North-Central Washington—Timing and Regional Context

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By KENNETH F. FOX, JR., C. DEAN RINEHART, *and* JOAN C. ENGELS

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PLUTONISM AND OROGENY IN NORTH-CENTRAL WASHINGTON—TIMING AND REGIONAL CONTEXT

By KENNETH F. FOX, JR., C. DEAN RINEHART, and JOAN C. ENGELS

ABSTRACT

Bedrock in north-central Washington comprises (1) weakly to moderately metamorphosed eugeosynclinal rocks of Permian, Triassic, and possibly Jurassic ages, (2) high-grade polymetamorphic rocks—gneiss, schist, and amphibolite—that are, at least in part, more highly metamorphosed derivatives of the rocks of the previous category, (3) Mesozoic and Cenozoic plutonic rocks, and (4) Cenozoic lavas and continental sedimentary deposits.

A review of the radiometric ages from north-central Washington suggests a complex history of plutonism and metamorphism beginning in Late Triassic and extending to Eocene time. However, ages of coexisting minerals from single samples typically are moderately to highly discordant. The discordance reaches a maximum along a zone flanking the Okanogan gneiss dome on the west. This dome forms the southwestern extremity of the Omineca crystalline belt, a north-trending orogenic subprovince about 250 km (150 mi) long and 55 km (35 mi) wide in British Columbia and Washington characterized by the presence of Shuswap (Monashee Group) terrane, gneiss domes, and allied metamorphic rocks.

The Okanogan gneiss dome and the Shuswap are believed to be products of metamorphism and deformation, in part at least, of Late Cretaceous age. The development of the gneiss domes indicates that metamorphism within the Omineca reached sufficient intensity at depth to cause incipient anatexis and mobilization of the infrastructure. The discordance west of the gneiss dome and the Omineca belt is attributed to weak thermal metamorphism that developed above and west of the zone of more intense metamorphism.

Ages from within the Okanogan gneiss dome range from Late Cretaceous (lead-uranium, zircon) to Eocene (K–Ar, biotite and muscovite; fission track, apatite). Their discordance is attributed to slow cooling after the climax of metamorphism in the Late Cretaceous.

Orogeny and plutonism in the north-central area of Washington began almost simultaneously during the Late Triassic with folding of the Permian and Triassic bedded rocks and their intrusion by the 195-million-year-old Loomis pluton. Thermal events subsequent to Late Triassic cannot be linked with specific orogenic deformation, except for deformation associated with the hypothesized Late Cretaceous metamorphism and mobilization of the Shuswap and the gneiss domes within it. This event is temporally associated with westward-directed overthrusting along the Shuksan thrust to the west, attributed to mid-Cretaceous orogeny, and with eastward-directed overthrusting to the east along the Cordilleran thrust belt, attributed to the Laramide orogeny, which ended in the Late Cretaceous or early Tertiary. The Shuswap terrane and associated gneiss domes appear to occupy the axial zone between convergent thrusts that show an aggregate crustal contraction of possibly 250 km (150 mi).

These relations suggest a genetic model as follows: The Permian volcanic and pyroclastic rocks of the region were probably deposited in island-arc and back-arc basins located east of an east-dipping subduction zone. We speculate that in Late Triassic, the continental plate overrode a rise system embedded in the oceanic plate, after which the eugeosynclinal prisms were invaded by calc-alkalic magmas derived through partial melting of upper mantle and lower crust within a zone of high heat flow above the overridden rise. The residue remaining in the zone of partial melting probably became progressively more dense as the hyperfusible part was removed and in Late Cretaceous catastrophically sank into the asthenosphere, forming a short-lived convection cell whose axis lay beneath the Okanogan region. The overlying crust was dragged toward this axis which caused thrust faulting to the east (Cordilleran thrusts) and west (Shuksan thrust) and thickening of the crust over the cell by stacking of thrust sheets and plastic flow. Concurrently, elements of the infrastructure were mobilized and penetrated higher levels in the crust, with their culminations forming the Okanogan gneiss dome and other gneiss domes within the Shuswap. After the demise of the convection cell in latest Cretaceous, the thickened crust isostatically rebounded. Upper crustal levels over the Omineca were rapidly eroded away, and elements of the Late Cretaceous infrastructure (the gneiss domes and the Shuswap) were exposed in the Eocene.

INTRODUCTION

Bedrock in north-central Washington (fig. 1) comprises (1) weakly to moderately metamorphosed eugeosynclinal rocks of Permian, Triassic, and possibly Jurassic age, (2) high-grade polymetamorphic rocks—gneiss, schist, and amphibolite—that are, at least in part, more highly metamorphosed derivatives of the rocks of the previous category, (3) Mesozoic and Cenozoic plutonic rocks, and (4) Cenozoic lavas and continental sedimentary deposits that patchily overlie the older rocks. The lavas are products of regional episodes of volcanism, the older of which is believed to be Eocene and the younger late Miocene to early Pliocene. The age of plutonism and metamorphism in the Okanogan area is bracketed by the Permian to Triassic age of the eugeosynclinal host rocks and the early Cenozoic age of the overlying continental deposits. Fossils are so rare in the eugeosynclinal deposits, however, that the older boundary is quite imprecise.

Previously reported radiometric ages of rocks from within the study area (fig. 2; table 1) range from Late Triassic to Eocene and taken collectively indicate that the history of magmatism and metamorphism has been

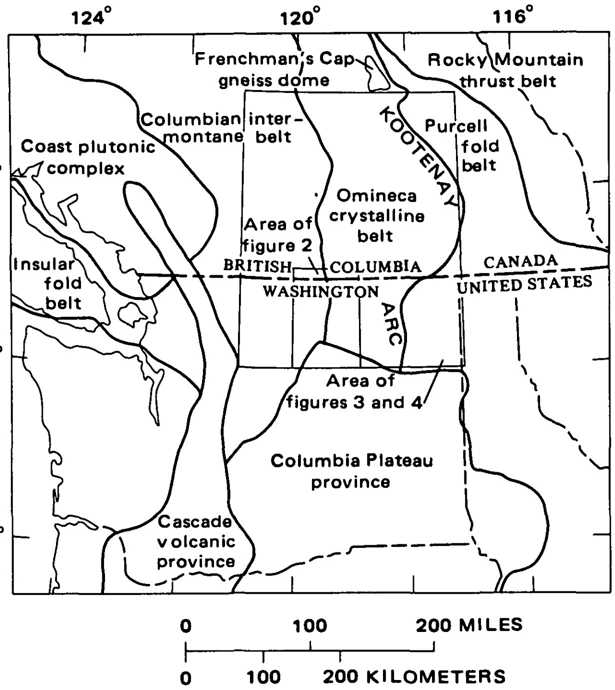


FIGURE 1.—Index map showing orogenic provinces in Washington and southern British Columbia. Modified and extended mainly from Wheeler (1970, p. 1).

complex. However, the age of many of the plutons in the study area, including some that have been studied intensively, has not been unambiguously determined. These rocks typically show substantial discordance between age determinations on mineral pairs from individual samples and discordance from sample to sample within individual plutons. In general, we assume that this discordance reflects varied retention of fission tracks, argon, rubidium, strontium, lead, or uranium, either during protracted cooling after initial emplacement of the pluton or varied retention during and after later thermal metamorphism. The temperature of a cooling mineral, at which the rate of loss of the radiogenic daughter element by diffusion becomes negligible compared to the rate of its accumulation, is defined as the critical blocking temperature (Macintyre and others, 1967, p. 822). This temperature may be a function of various factors, notably grain size, the pressure of argon within the host but external to the dated mineral, and the length of time during which that particular temperature is maintained. The blocking temperature for radiogenic argon is higher in hornblende than in coexisting biotite, and as a result, in thermally metamorphosed terranes the hornblende commonly yields an older age than the coexisting biotite (Hart, 1964; Hanson and Gast, 1967). The concept of a critical blocking temperature can also be extended to the rate of erasure of fission tracks relative to their rate of formation. Thus, the minimum age of crystallization of plutonic rocks yielding discordant K–Ar or fission-track ages is here assumed to be the older of a mineral pair.

This paper refines the plutonic and metamorphic history of part of the Okanogan Highlands, mainly through review of the radiometric ages and consideration of the implications of their discordance, and relates this history to the depositional and orogenic history of the region. The eastern part of the region includes stratified rocks of Precambrian and early Paleozoic age that are not found within the study area. The distribution of these rocks, along with that of the more widespread late Paleozoic and Mesozoic stratified rocks, and their various dynamically metamorphosed derivatives provides the basis for dividing the region into depositional and orogenic provinces. We define these provinces and then discuss the Mesozoic and Cenozoic geologic history of the study area, where possible in the broader context of the geologic history of the region.

We are grateful to R. G. Yates, who introduced us to the geologic problems of the region and who provided advice and encouragement throughout the duration of the study. We are also indebted to M. D. Crittenden for numerous stimulating discussions on tectonic problems of the region.

GEOLOGIC PROVINCES

The region encompassed by figure 3 comprises parts of two depositional provinces of differing ages. The pre-Tertiary stratified rocks of that part of northeastern Washington lying roughly east of long 118° W. are chiefly Precambrian and lower Paleozoic miogeoclinal (Dietz and Holden, 1966) rocks, in contrast to those to the west, which are upper Paleozoic and Mesozoic eugeosynclinal rocks.

In Washington, the contacts between the eugeosynclinal rocks and the miogeoclinal strata to the east are obscured by faulting and folding along much of their extent (Yates and others, 1966, p. 54). According to Campbell (1964), the eugeosynclinal deposits in the Hunters quadrangle and vicinity (south of Kettle Falls, Wash.) have been thrust eastward over the miogeoclinal rocks. In British Columbia however, rocks of the eugeosyncline (Milford Group) concordantly but unconformably overlie rocks probably belonging to the miogeocline (Lardeau Group, Cairnes, 1934, p. 36–38).

In addition to the depositional provinces discussed above, the southern British Columbia and north-central Washington region can also be assigned to three orogenic provinces, the Purcell foldbelt, the Omineca crystalline belt, and the Columbian intermontane belt (fig. 1).

The Omineca crystalline belt is distinguished chiefly by the presence of medium- to high-grade gneiss and schist of the Monashee Group of the Shuswap terrane of Jones (1959). Several gneiss domes have been identified within the rocks of the Monashee Group and its probable correlatives. The gneiss domes include among others the Valhalla and Passmore domes in British Columbia (Reesor, 1965) and the Okanogan dome in Washington (Fox and Rinehart, 1971) (fig. 1).

The eastern limit of the Monashee Group and Monashee-like rocks within the region serves as the boundary with the Purcell foldbelt, and the western limit serves as the boundary with the Columbian intermontane belt. The contact between the Purcell foldbelt and the Omineca crystalline belt (figs. 1, 3) nearly coincides with the contact between rocks of the eugeosynclinal facies and the rocks of the miogeoclinal facies.

GEOLOGIC HISTORY

PERMIAN AND TRIASSIC HISTORY

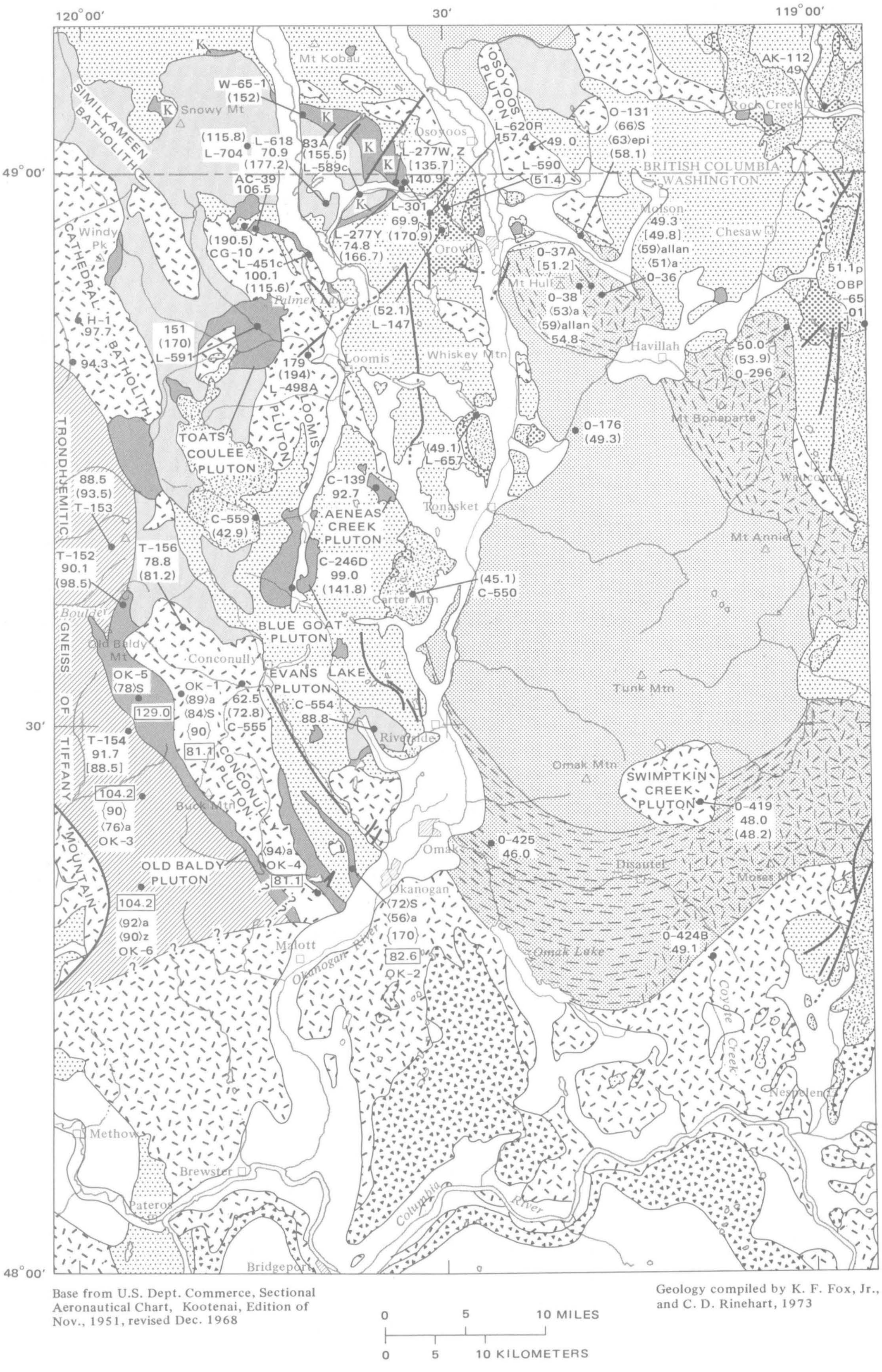
The eugeosynclinal province within the area of figure 3 is chiefly floored by sparsely fossiliferous rocks, probably of late Paleozoic age. Within the study area (fig. 2), these rocks have been assigned to the Anarchist Group (Rinehart and Fox, 1972, p. 8–11), which is composed of variably metamorphosed complexly interfingering and intergrading deposits of argillite, siltstone, sharpstone conglomerate, graywacke, sandstone, and limestone. Deposits exclusively of volcanic origin—metamorphosed lava or pyroclastic material—are also present but are not abundant. However, clasts of volcanic rock are a minor but apparently widespread constituent of the coarser grained clastic rocks, such as the sharpstone conglomerate and graywacke, suggesting that material of volcanic origin may be a significant constituent of the finer grained clastic rocks as well. The Anarchist is believed to have an overall thickness in excess of 4,500 m (15,000 ft). Fossils are rare. Those found, which are probably restricted to a narrow stratigraphic interval in the upper middle of the group, have been dated as Permian, perhaps Late Permian (Rinehart and Fox, 1972, p. 9–10; Waters and Krauskopf, 1941, p. 1364). The Anarchist is therefore, at least in part a temporal equivalent of the lithologically similar Cache Creek Group of south-central British Columbia (fig. 4), from which fossils of Mississippian(?), Pennsylvanian, and Permian ages have been reported (Cockfield, 1961, p. 9–11).

Within the northern part of the study area (fig. 2), the Kobau Formation and Palmer Mountain Greenstone overlie the Anarchist Group disconformably or along a slight unconformity (Rinehart and Fox, 1972, p. 11–12, 22). The Palmer Mountain Greenstone consists of metavolcanic rocks estimated to be locally at least 2,100 m (7,000 ft) thick, consisting chiefly of greenstone and metadiabase. The formation probably originated as a pile of mafic extrusive rocks—a volcano—into which mafic magma was abundantly intruded.

The Kobau Formation is composed of an unfossiliferous sequence—locally at least 3,600 m (12,000 ft) thick—of phyllite, greenstone, and massive metachert. The lower parts of the Kobau apparently interfinger with the upper part of the Palmer Mountain, leading to the hypothesis that the Kobau originated "as a belt of marine deposits that formed satellitic to, and partly as a result of, the volcanic activity represented by the Palmer Mountain Greenstone" (Rinehart and Fox, 1972, p. 22). The maximum possible age of the Kobau and Palmer Mountain, as established by the fossils in the subjacent Anarchist, is Permian, probably Late Permian. Okulitch (1973, p. 1514) rejected this conclusion and suggested instead that the Kobau Formation as mapped by us is in part older than the Anarchist because the Kobau contains rootless isoclinal folds that he believed were related to a deformational event that predates the Anarchist. In his opinion (p. 1516), the Kobau is mid-Paleozoic. Our mapping (Rinehart and Fox, 1972) revealed no evidence supporting his speculations, and in our view the disconformable or unconformable contact of Kobau over Anarchist is well established.

In the northwest corner of the region, near Kamloops (fig. 4), the Cache Creek Group is overlain along an erosional unconformity by the Nicola Group, a sequence of volcanic rocks composed mainly of greenstone and minor sedimentary rocks containing Triassic—locally Late Triassic—fossils (Cockfield, 1961, p. 8–15). The Nicola crops out over an extensive area within south-central British Columbia. The stratigraphic positions of the Nicola over the Cache Creek and Kobau over the Anarchist thus seem comparable and suggest that these two chiefly volcanic and volcaniclastic assemblages—Nicola and Kobau—are correlative. However, Dawson (1879, p. 87B) considered the rocks now referred to the Kobau (along the route of his reconnaissance between the Similkameen and Okanogan Rivers and a few kilometers north of the international boundary) to be Cache Creek, although he recognized that the sequence contained members resembling the Nicola. In view of the relations noted above, the Kobau could be either the temporal equivalent of the Nicola—differing from the Nicola of the type area in that volcaniclastic rocks and bedded chert predominate over lava—or alternatively, older than the Nicola but not present or not recognized in the type areas of the Nicola and the Cache Creek.

In the southern part of the study area, the Anarchist is unconformably overlain by the Cave Mountain Formation (Rinehart and Fox, 1976), a sequence of



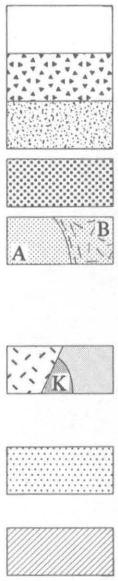
Base from U.S. Dept. Commerce, Sectional Aeronautical Chart, Kootenai, Edition of Nov., 1951, revised Dec. 1968

Geology compiled by K. F. Fox, Jr., and C. D. Rinehart, 1973

FIGURE 2.—Geologic map of Okanogan Highlands and vicinity showing radiometric ages.

weakly to highly metamorphosed limestone, dolomite, siltstone, slate, and capping basaltic volcanic rock. One member contains fossils dated as Late Triassic (Misch, 1966, p. 118). These rocks are similar to a sequence of interlayered quartzite and limestone at Hedley, British Columbia, originally described by Camsell (1910) that also contains Late Triassic fossils. The nomenclature and lithology of these rocks have been reviewed by Rice (1947, p. 12–14), who considers them to be part of a dominantly sedimentary facies of the Nicola. Within

EXPLANATION



SURFICIAL DEPOSITS (Quaternary) – Alluvium and glacial drift

COLUMBIA RIVER BASALT GROUP (upper Miocene)

VOLCANIC ROCKS(Eocene) – Dacitic , andesitic, and minor trachytic lava flows and hypabyssal intrusive rocks

EPICLASTIC AND VOLCANICLASTIC ROCKS (Eocene) – Sandstone, graywacke, tuff, conglomerate, and shale

PARAGNEISS, ORTHOGNEISS, AND ASSOCIATED GRANITIC ROCKS OF THE OKANOGAN GNEISS DOME (age of metamorphism and mobilization probably late Mesozoic) – Includes similar rocks of the Monashee Group of the Shuswap terrane of Jones (1959) in British Columbia.A, paragneiss; B, orthogneiss and associated granitic rocks

GRANITIC ROCKS (Triassic to Tertiary) – Includes granodiorite, quartz monzonite, quartz diorite, diorite, monzonite, and minor nepheline syenite, shonkinite, malignite, alkali pyroxenite, peridotite, hornblendite, and serpentinite. Interplutonic contacts shown where known. Kruger Alkalic Complex identified by "K"

EUGEOSYNCLINAL DEPOSITS (Permian to Cretaceous) – Greenstone, greenschist, slate, phyllite, schist, metawacke, metasharpstone conglomerate, marble, and metachert

GRANITOID GNEISS OF OKANOGAN RANGE (age unknown) – Trondhjemitic orthogneiss and paragneiss

?

Contact
Queried where projected through unmapped areas

Fault
Dotted where concealed

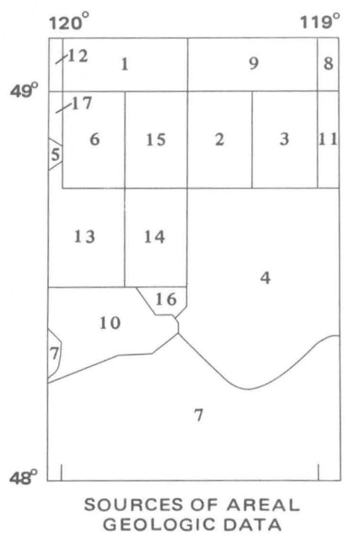
KEY TO RADIOMETRIC AGES
(Ages compiled from sources listed in table 1)

Locality L-618 ●
K-Ar age
- Biotite 179
- Muscovite [51.2]
- Hornblende (194)
- Plagioclase 51.1p

Fission track age
- Apatite ⟨53⟩a
- Sphene ⟨78⟩s
- Allanite ⟨59⟩allan
- Zircon ⟨90⟩z
- Epidote ⟨63⟩epi

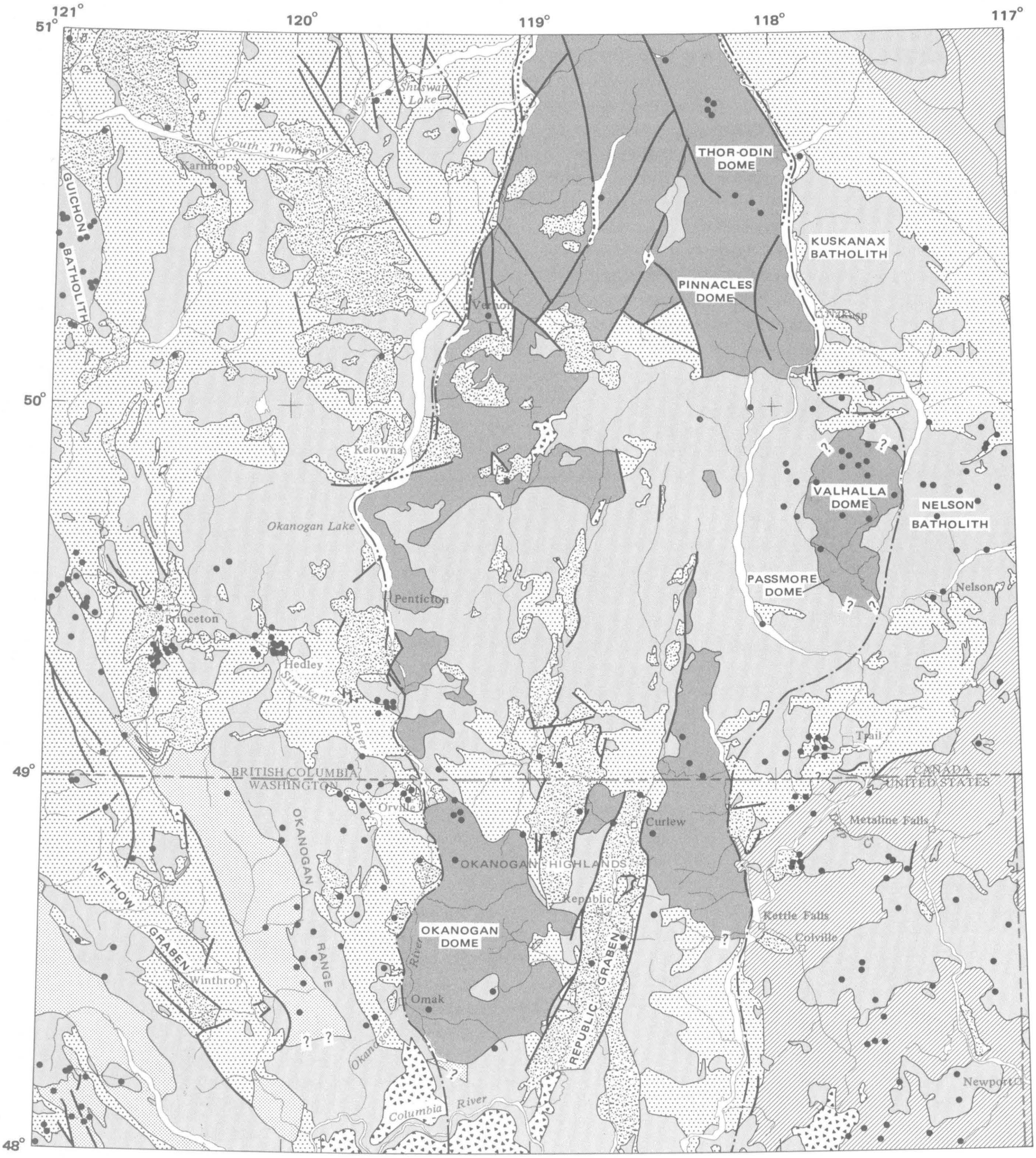
Rb-Sr age [83]

Pb-α age {170}



1. Bostock (1940)
2. Fox (1970)
3. Fox (unpublished mapping)
4. Fox and Rinehart (unpublished mapping)
5. Hawkins (1968)
6. Hibbard (1971)
7. Huntting and others (1961)
8. Little (1957)
9. Little (1961)
10. Menzer (1964, 1970)
11. Pearson (1967)
12. Rice (1947)
13. Rinehart (unpublished mapping)
14. Rinehart and Fox (1970)
15. Rinehart and Fox (1972)
16. Roberts, R. J., and Hobbs, S. W. (unpublished mapping)
17. Staatz and others (1971)

FIGURE 2.—Continued.



Base from U.S. Dept. Commerce, Navigational Chart E 16 Edition 5, 1968

Geology compiled by K. F. Fox, Jr., and C. D. Rinehart, 1973

0 15 30 MILES

0 15 30 KILOMETERS

FIGURE 3.—Geologic map of part of northeastern Washington and southern British Columbia showing distribution of localities at which age determinations have been made.

EXPLANATION

SEDIMENTARY, VOLCANIC, AND LOW-GRADE METAMORPHIC ROCKS

BASALT (upper Miocene and lower Pliocene)

EPICLASTIC SEDIMENTARY ROCKS AND VOLCANIC ROCKS (Eocene) – Basaltic, andesitic, dacitic, rhyolitic, and trachytic lava flows, hypabyssal intrusive rocks, pyroclastic rocks, sandstone, graywacke, shale, and conglomerate

EUGEOSYNCLINAL DEPOSITS (Cambrian? to Cretaceous) – Sandstone, graywacke, argillite, conglomerate, lava flows, pyroclastic rock, greenstone, greenschist, slate, phyllite, schist, metawacke, metasharpstone conglomerate, marble, and metachert

MIOGEOCLINAL DEPOSITS (Precambrian to Mississippian?) – Limestone, dolomite, argillite, slate, phyllite, schist, marble, quartzite, siltite, and minor greenstone and conglomerate

MEDIUM- AND HIGH-GRADE METAMORPHIC ROCKS

PARAGNEISS, SCHIST, AND AMPHIBOLITE (Age not specified)

MONASHEE GROUP OF SHUSWAP TERRANE OF JONES (1959), AND SIMILAR ROCKS (age of metamorphism probably Mesozoic) – Paragneiss, orthogneiss, and associated granitic rocks, probably in part originating as diapiric intrusions. Includes rock within Thor-Odin, Pinnacles, Valhalla, Passmore, and Okanogan gneiss domes

PLUTONIC ROCKS

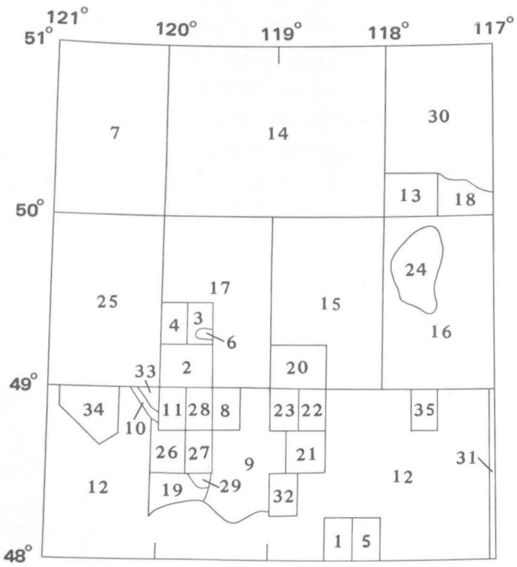
GRANITIC ROCKS (Triassic to Tertiary) – Includes granodiorite, quartz monzonite, syenite, monzonite, quartz diorite, diorite, granite, and minor nepheline syenite, shonkinite, malignite, alkali pyroxenite, gabbro, peridotite, dunite, hornblendite, and serpentinite

Radiometric age locality
Compiled from sources listed in tables 1 and 2

Boundary of orogenic province

Contact
Queried where projected through unmapped areas

Fault
Dotted where concealed



SOURCES OF AREAL GEOLOGIC DATA

1. Becraft (1966)
2. Bostock (1940)
3. Bostock (1941a)
4. Bostock (1941b)
5. Campbell and Raup (1964)
6. Church (1971)
7. Cockfield (1961)
8. Fox (1970)
9. Fox and Rinehart (unpublished mapping)
10. Hawkins (1968)
11. Hibbard (1971)
12. Huntting and others (1961)
13. Hyndman (1968b)
14. Jones (1959)
15. Little (1957)
16. Little (1960)
17. Little (1961)
18. Little (1962)
19. Menzer (1964, 1970)
20. Monger (1968)
21. Muessig (1967)
22. Parker and Calkins (1964)
23. Pearson (1967)
24. Reesor (1965)
25. Rice (1947)
26. Rinehart (unpublished mapping)
27. Rinehart and Fox (1970)
28. Rinehart and Fox (1972)
29. Roberts, R. J. and Hobbs, S. W., (unpublished mapping)
30. Ross (1970)
31. Savage (1967)
32. Staatz (1964)
33. Staatz and others (1971)
34. Tabor and others (1968)
35. Yates (1964)

FIGURE 3.—Continued.

TABLE 1.—*Sources of radiometric age data compiled in figures 2, 3, 5, and 6*

Figure	*References*
2, 5	Cannon, 1966
	Engels, 1971
	Engels, Tabor, Miller, and Obradovich, 1976
	Fox, Rinehart, and Engels, 1975
	Fox, Rinehart, Engels, and Stern, 1976
	Hawkins, 1968
	Hibbard, 1970, 1971
	Mathews, 1964
	Menzer, 1970
	Naeser, Engels, and Dodge, 1970
	Rinehart and Fox, 1972
	Rinehart and Fox, 1976
3, 6	All above, plus the following:
	Baadsgaard, 1961
	Engels, 1975
	Hills and Baadsgaard, 1967
	Leech, Lowden, Stockwell, and Wanless, 1963
	Lowdon, 1960, 1961
	Lowdon, Stockwell, Tipper, and Wanless, 1963
	Nguyen, Sinclair, and Libby, 1968
	Preto, White, and Harakal, 1971
	Roddick and Farrar, 1971, 1972
	Roddick, Farrar, and Procyshyn, 1972
	Sinclair and White, 1968
	Tabor, Engels, and Staatz, 1968
	Wanless, Stevens, Lachance, and Rimšaite, 1965
	Wanless, Stevens, Lachance, and Edmonds, 1968
	White and others, 1967
	Yates and Engels, 1968

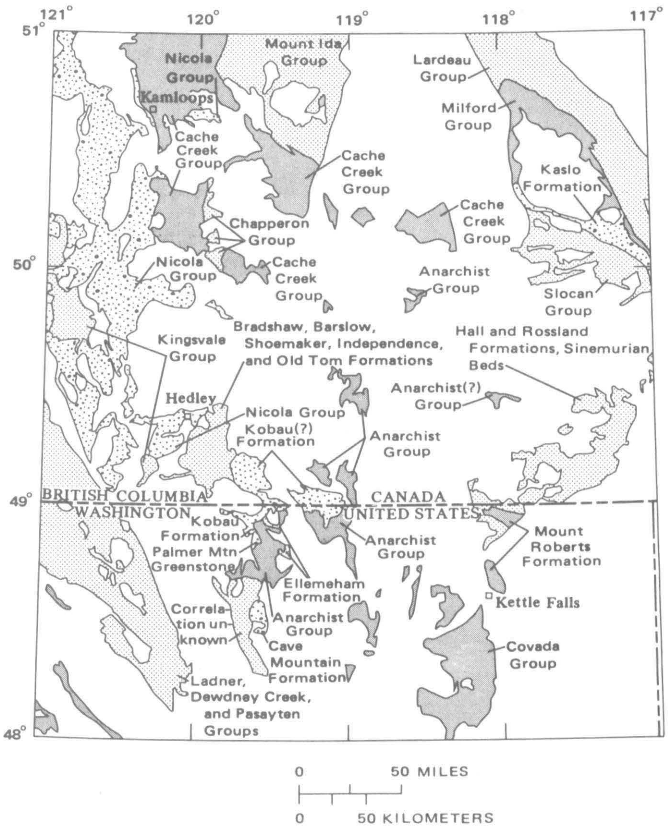


FIGURE 4.—Late Paleozoic and Mesozoic stratified rocks of the eugeosynclinal province in northeastern Washington and southern British Columbia. Contrasting patterns used to aid in distinction of map units. Data modified slightly from same sources as figure 3.

the study area (fig. 2), both the Anarchist Group and Kobau Formation were strongly folded along north-northwest axes and afterwards intruded and metamorphosed by the Loomis pluton, a quartz diorite–granodiorite batholith. The Loomis pluton is the oldest pluton so far recognized within the study area (fig. 2). K–Ar ages of 194 m.y. and 179 m.y. were measured on coexisting hornblende and biotite (table 2), respectively. The older of these is regarded as a minimum age—and probably close to the true age of the batholith, in view of the modest amount of the discordance.

The relations within the study area (fig. 2) indicate that deposition of eugeosynclinal volcaniclastic sediments, carbonate sediments, and minor intercalated lavas was terminated during the Permian. After an interval of minor tilting, and possibly of minor erosion, voluminous lavas were erupted and deposited along with associated pyroclastic sediments. This phase of volcanism was terminated in Late Triassic by uplift and strong folding, followed by batholithic intrusion about 195 m.y. ago.

JURASSIC TO PALEOCENE HISTORY

The folded Kobau Formation and Anarchist Group were beveled and buried beneath the Ellemeham Formation, a unit composed of a basal member of greenstone and tuffaceous metasiltstone, a middle member of monolithologic greenstone conglomerate, and an upper conglomeratic member composed of heterogeneous fragments of metamorphic rock. The Ellemeham Formation is locally at least 850 m (3,000 ft) thick (Rinehart and Fox, 1972, p. 22–25). The Ellemeham is younger than the subjacent Kobau, probably younger than the Late Triassic Loomis pluton, and is clearly older than 157 m.y., the age of biotite (sample L62OR, table 2) in a hornfels formed at the contact of a crosscutting alkalic intrusive body; hence it appears to be of Late Triassic or Jurassic age.

With the possible exception of the Ellemeham Formation, there are no known depositional units of Jurassic to Paleocene age within the study area. Rocks older than Ellemeham were eroded to levels deep enough to expose the plutonic rocks, then buried by the lavas, pyroclastic deposits, breccia, and conglomerate of the Ellemeham.

Deposition was interrupted in the central and western parts of the eugeosynclinal province during the Late Triassic tectonism, but on the east flank of the province,

TABLE 2.—*Interpretation of radiometric age determinations in part of northern Okanogan Highlands*

Unit	Sample	Mineral	Method of age determination[1]	Mineral age (m.y.)	Source of mineral age	Inferred age of event (m.y.)	Remarks
Swimptkin Creek pluton.	0–419	Biotite	K–Ar	48.0±1.5	Fox, Rinehart, Engels, and Stern, 1976.	Intrusion, rapid cooling at ~48	Swimptkin Creek pluton cuts Okanogan gneiss dome; contains xenoliths of lineated gneiss of dome and of low-grade metamorphic rocks believed to have been part of roof of dome when cut by Swimptkin Creek. See Newcomb (1937). Contains autoliths and fine-grained marginal phases.
		Hornblende	do	48.2±1.5	do.		
Andesite of Whitestone Mountain.	L–657	Hornblende	K–Ar	49.1±1.8	Rinehart and Fox, 1972.	Volcanism, ~50.	Volcanic rocks cut or overlie associated sedimentary rocks that contain probably Eocene flora (Rinehart and Fox, 1972, p. 61). Although volcanic rocks fringe gneiss dome, they do not directly overlie it. Intercalated conglomerate contains abundant detrital granitic rock, which northeast of dome resembles rocks of gneiss dome.
Andesite-dacite plug	L–590	do	do	51.4±2.6	do.		
Do	L–147	do	do	52.1±2.3	do.		
Dacite of Carter Mountain.	C–550	do	do	45.1±2.0	Rinehart and Fox, 1976.		
Marron Formation (volcanics).	AK–112	do	do	49	Mathews, 1964, p. 465.		
Dacite of Twin Peaks.	C–559	do	do	42.9±1.3	Engels, Tabor, Miller, and Obradovich, 1976.		
Sanpoil Volcanics	OBP–65–01	Plagioclase	do	51.1	do		
Okanogan gneiss dome.	0–36A	Biotite	K–Ar	49.3±1.6	Fox, Rinehart, Engels, and Stern, 1976.	Intrusion and metamorphism in Late Cretaceous; uplift and slow cooling to retentivity threshold of Ar in biotite at ~50; unroofing of dome after 50.	Samples 0–131, 0–296 of alkalic border phase (monzonite and syenodiorite) at north edge of dome, interpreted as metasomatite formed during emplacement of dome. Sample 0–176D and 0–176E are of layered paragneiss, and remaining samples are of granodioritic gneiss. Volcanic rocks unconformably overlie country rocks dynamically metamorphosed during forceful intrusion of dome; therefore, volcanics are younger than age of emplacement of dome. The $^{206}Pb/^{238}U$ age of 87.3 m.y. is the most precise of the Pb-U ages. The cataclastic textures of the rocks of the gneiss dome indicate that pervasive and penetrative deformation continued well after most mineral constituents including zircon had crystallized. We conclude that deformation and mobilization terminated in Late Cretaceous, after about 87 m.y. ago. Younger ages of other minerals reflect slow cooling. It is less likely, in our opinion, that discordance could be due wholly or in part to thermal metamorphism of gneiss dome at ~50 m.y.
		Muscovite	K–Ar	49.8±1.6	do.		
		Allanite	F.T	59±6	Naeser, Engels, and Dodge, 1970.		
		Apatite	F.T	51±5	do.		
	0–37A	Muscovite	K–Ar	49.0±2.2	Fox, Rinehart, Engels, and Stern, 1976.		
				53.8±1.6	do.		
				51.2±1.6	do.		
				50.8±1.6	do.		
	0.38A	Biotite		54.8±1.7	do.		
		Allanite	F.T	59±2	Naeser, Engels, and Dodge, 1970.		
		Apatite	F.T	53±5	do		
	0–131	Hornblende	K–Ar	58.1±1.7	Fox, Rinehart, Engels, and Stern, 1976.		
		Epidote	F.T	63±3	Naeser, Engels, and Dodge, 1970		
		Sphene	F.T	66±7	do.		
	0–176D	Hornblende	K–Ar	49.3±1.7	Fox, Rinehart, Engels, and Stern, 1976.		
	0–176E	Zircon	$^{206}Pb/^{238}U$	87.3	do.		
			$^{207}Pb/^{235}U$	100.0	do.		
			$^{208}Pb/^{232}Th$	94.0	do.		
	0–296D	Biotite	K–Ar	50.0±1.5	do.		
		Hornblende	do	53.9±1.6	do.		
	0–425	Biotite	do	46.0±1.4	do.		
Coyote Creek pluton	0–424B	Biotite	K–Ar	49.1±1.5	Fox, Rinehart, Engels, and Stern, 1976	~49(?)	Granodiorite porphyry of Coyote Creek pluton bounds Okanogan gneiss dome on southeast, not delineated separately in figure 2. Probably younger than gneiss dome, since only slight cataclasis apparent, but it is possible that granodiorite porphyry is marginal phase grading inward to gneissic quartz diorite of gneiss dome.
Conconully pluton	C–555	Biotite	K–Ar	62.5±2.2	Rinehart and Fox, 1976.	Intrusion prior to 80, probably at about 90. Local thermal metamorphism at or subsequent to 63.	Menzer (1970, p. 576) concludes that this event is 81 m.y. old, relying on the Rb-Sr isochron. However, the F.T. and Pb-α ages average 89, suggesting that the pluton is older than 81. The discordance between the hornblende and biotite K-Ar ages at C–555 and T–156, and between hornblende at C–555 and T–156, and between hornblende at C–555 and the mineral ages of OK–1 and OK–4, probably reflect metamorphism by a later thermal event. A steep east-west gradient in discordancy is required between C–555 and OK–1, and between C–555 and T–156 (fig. 2), suggesting metamorphism by a nearby source to the east concealed in the subsurface.
		Hornblende	do	72.8±4.6	do.		
	T–156	Biotite	do	78.8±2.4	Engels, Tabor, Miller, and Obradovich, 1976		
		Hornblende	do	81.2±2.4	do.		
	OK–1	Apatite	F.T	89±9	Menzer (1970).		
		Sphene	do	84±8	do.		
		Zircon	Pb–α	90±20	do.		
	OK–4	Apatite	F.T	94±12	do.		
	OK–1 OK–4	Apatite, biotite. Whole rock, muscovite, biotite.	Rb–Sr	81.1±0.8	do.		
Trondhjemitic gneiss of Tiffany Mountain.	T–153	Biotite	K–Ar	88.5±2.7	Engels, Tabor, Miller, and Obradovich, 1976.	Intrusion prior to intrusion of Old Baldy pluton (129 or before)	Trondhjemitic gneiss is enclosed in, and is thus older than Old Baldy pluton ("granodioritic gneiss," Menzer, 1970, p. 576). Discordancy possibly due to cumulative metamorphism related to intrusion of Conconully pluton about 90 m.y. ago and to Late Cretaceous and early Tertiary thermal event. Menzer ascribes its anomalously younger isotopic age, relative to Old Baldy pluton, as possibly due to "penetrative rock deformation that pervades the unit" (p. 577). Thus far, however, our observations reveal no consistent differences between the two units in age of deformation, its intensity, or its style.
		Hornblende	do	93.5±2.8	do.		
	T–154	Biotite	do	91.7±2.8	do.		
		Muscovite	do	88.5±2.7	do.		
	T–155	Biotite	do	108±3	do.		
		Muscovite	do	94.6±2.8	do.		
("Trondhjemitic gneiss"; Menzer, 1970)	OK–3	Apatite	F.T	76±8	Menzer (1970).		
		Zircon	Pb–α	90±10	do.		
	OK–6	Apatite	F.T	92±9	do.		
		Zircon	do	90±9	do.		
	OK–3 OK–6	Whole rock, biotite. do	Rb–Sr isochron	104.2±0.5	do.		

TABLE 2.—*Interpretation of radiometric age determinations in part of northern Okanogan Highlands*—Continued

Unit	Sample	Mineral	Method of age determination[1]	Mineral age (m.y.)	Source of mineral age	Inferred age of event (m.y.)	Remarks
Old Baldy pluton	T-152	Biotite	K-Ar	90.1±2.7	Engels, Tabor, Miller, and Obradovich, 1976.	Intrusion at 129 or before.	The Old Baldy pluton is cut by the Conconully pluton. Discordancy due to cumulative metamorphism including that associated with intrusion of Conconully pluton about 90 m.y. ago and that due to Late Cretaceous and early Tertiary regional thermal event.
		Hornblende	do	98.5±3.0	do.		
("Granodioritic gneiss" of Menzer, 1970).	OK-5	Sphene	F.T	78±8	Menzer (1970).		
		Whole rock, biotite.	Rb-Sr isochron.	129.0±1.8	do.		
Cathedral batholith	H-1	Biotite	K-Ar	97.7±2.9	Engels, Tabor, Miller, and Obradovich, 1976.	Intrusion at ~95(?).	K-Ar ages of biotite provisionally accepted, since no younger rocks known in vicinity, and sample locations appear to be west of area of influence of Late Cretaceous and early Tertiary regional thermal event.
	(2)	do	do	94.3±2.8	Hawkins, 1968.		
	(2)	do	do	98	Misch (cited by Hawkins, 1968).		
Anderson Creek pluton.	L-451C	Biotite	K-Ar	100.1±3.0	Engels, Tabor, Miller, and Obradovich, 1976.	Intrusion at ~115.	Intrudes Loomis pluton (Rinehart and Fox, 1972). No younger plutons known in vicinity; therefore, discordancy between biotite and hornblende ages attributed to Late Cretaceous and early Tertiary regional thermal event.
		Hornblende	do	115.6±3.0	do.		
	AC39	Biotite	do	106.5±6.8	Hibbard, 1971		
Aeneas Creek pluton.	C-139	Biotite	K-Ar	92.7±6.6	Rinehart and Fox, 1976.	Intrusion prior to 93.	Well within area of influence of Late Cretaceous and early Tertiary regional thermal event; therefore, biotite age probably substantially less than actual age of intrusion of pluton.
Evans Lake pluton	C-554	Biotite	K-Ar	88.8±2.8	Rinehart and Fox, 1976.	Intrusion before 89.	Well within area of influence of Late Cretaceous and early Tertiary regional thermal event; therefore, biotite age probably substantially less than actual age of intrusion of pluton.
Osoyoos pluton	Osoyoos #2	Biotite	K-Ar	49.0±1.5	Fox, Rinehart, Engels, and Stern, 1976.	Intrusion before 50 m.y. ago.	Well within area of influence of Late Cretaceous and early Tertiary regional thermal event; therefore, biotite age probably substantially less than actual age of intrusion of pluton. Dynamic metamorphism along south border attributed to intrusion of Okanogan gneiss dome, therefore older than dome.
Blue Goat pluton	C-246D	Biotite	K-Ar	99.0±3.0	Rinehart and Fox, 1976.	Intrusion at or before 142.	Discordance attributed to metamorphism during Late Cretaceous and early Tertiary regional thermal event.
		Hornblende	do	141.8±8.2	do.		
Quartz diorite gneiss (Menzer, 1970).	OK-2	Apatite	F.T	56±6	Menzer, 1970.	Intrusion before 170.	Discordance attributed to cumulative effect of metamorphism during Late Cretaceous and early Tertiary regional event and metamorphism associated with intrusion of Conconully pluton about 90 m.y. ago. Menzer implies that this rock is a marginal facies of a granodioritic gneiss unit (1970, p. 573); latter is Old Baldy pluton of this paper.
		Sphene	do	72±7	do.		
		Zircon	Pb-α	170±20	do.		
		Whole rock biotite.	Rb-Sr isochron.	82.6±0.3	do.		
Toats Coulee pluton.	L-591	Biotite	K-Ar	151±5	Rinehart and Fox, 1972.	Intrusion between 170 and 195.	Intrudes Loomis pluton (Rinehart and Fox, 1972). Discordance attributed to cumulative effect of metamorphism during Late Cretaceous and early Tertiary regional thermal event and metamorphism associated with intrusion of Cathedral batholith about 95 m.y. ago.
		Hornblende	do	170±5	do.		

if the age of the Slocan Group overlaps the Late Triassic and Early Jurassic as indicated by Little (1960, p. 56–57), deposition there must have been roughly concurrent with deformation and plutonism in the central and western parts—possibly contemporaneous with deposition of the Ellemeham Formation in the study area. In Middle and probably Late Jurassic time, however, deposits were accumulating on both flanks of the province. Those on the east side were primarily lavas and related rocks (Rossland Group; Little, 1960, p. 62–71; Frebold and Little, 1962, p. 3–9), whereas those on the west side (Ladner and Dewdney Creek Groups) were primarily marine volcaniclastic sediments derived from erosion of the Nicola terrane in the central part of the region (Coates, 1970, p. 150–151) and deposited in a subsiding fault-bounded trough (White, 1959, p. 77), the Methow graben.

Sedimentation in the western part of the eugeosynclinal province within the region (fig. 3) continued through the Early Cretaceous marked by the appear-

TABLE 2.—*Interpretation of radiometric age determinations in part of northern Okanogan Highlands*—Continued

Unit	Sample	Mineral	Method of age determination[1]	Mineral age (m.y.)	Source of mineral age	Inferred age of event (m.y.)	Remarks
Shankers Bend alkalic complex.	L–620R	Biotite	K–Ar	157.4±4.7	Engels, Tabor, Miller, and Obradovich, 1976.	Intrusion before 157.	Sample of hornfelsed rock of Ellemeham Formation at contact; mid-Jurassic age surprising, since locality is well within area influenced by Late Cretaceous and early Tertiary regional thermal event. Ellemeham Formation probably Upper Triassic, Lower Jurassic, or Middle Jurassic.
Similkameen composite pluton. (Comprises Similkameen batholith (Daly, 1906, 1912) and Kruger Alkalic Complex (Daly, 1906, 1912; Campbell, 1939; Fox, Rinehart, and Engels, 1975).	L–277Y (hornfelsed Koban at contact)	Biotite	K–Ar	74.8±2.2	Fox, Rinehart, and Engels, 1975.	Intrusion between 191 (isochron age of hornblendes; see Fox, Rinehart, and Engels, 1975) and 177; thermal metamorphism between 70 and 50.	The batholith is zoned, grading outward from quartz monzonite and granodiorite to monzonite. The monzonite in turn grades outward to malignite and shonkinite of the bordering Kruger Alkalic Complex. This gradation suggests the batholith and alkalic complex are roughly coeval. The complex sharply crosscuts wallrock of the Kobau Formation, of Permian or Triassic age. Discordance attributed to Late-Cretaceous or early Tertiary regional thermal metamorphism.
		Hornblende	do	166.7±5.0	do.		
	L–277W (pegmatite-alaskite dike cuts pyroxenite phase)	Muscovite	do	135.7±4.1	do.		
					do.		
	L–277Z (pyroxenite phase)	Biotite	do	140.9±4.2	do.		
	L–589C (granodiorite interior phase)	do	do	83.4±2.5	do.		
		Hornblende	do	155.5±4.7	do.		
	L–301 (shonkinite alkalic border phase)	Biotite	do	69.9±2.1	Engels, 1971.		
		Hornblende (hastingsite).	do	170.9±5.1	do.		
	L–618 (granodiorite, interior phase)	Biotite	do	[3]70.9±2.1	do.		
		Hornblende	do	[3]177.2±5.3	do.		
	L–704 (granodiorite, interior phase)	do	do	115.8±3.6	Fox, Rinehart, and Engels, 1975.		
					do.		
	W–65–1 (Kruger syenite; shonkinite from alkalic border phase)	Hornblende-augite mixture.	do	152±9	Cannon, 1966.		
Loomis pluton	L–498A	Biotite	K–Ar	179±5	Rinehart and Fox, 1972	Intrusion about 195 m.y.	Discordance attributed to cumulative effect of metamorphism during Late Cretaceous and early Tertiary regional thermal event, and to metamorphism associated with intrusion of the Toats Coulee between 195 and 170 m.y. ago. and the metamorphism associated with intrusion of Cathedral batholith about 95 m.y. ago.
		Hornblende	do	194±6	do.		
Chopaka Intrusive Complex of Hibbard (1971).	CG10	Actinolitic hornblende.	K–Ar	190.5±15.6	Hibbard, 1971.	Metamorphism about 195 m.y. ago.	Probably the age of metamorphism associated with intrusion of the nearby Loomis pluton.

[1]F.T., fission track.
[2]Sample number unknown.
[3]Corrected for cross-contamination of biotite and hornblende in mineral separates (see Engels, 1971).

ance of material eroded from sources west of the basin as well as from the east and by the appearance of granitic detritus (Coates, 1970, p. 151). The central part of the province, which includes the study area (fig. 2), was apparently positive and was the site of continued plutonism through the Jurassic and Early Cretaceous. By mid-Cretaceous most of the province had become positive, and the area of plutonism had spread eastward and westward through the entire extent of the province within the region (fig. 3).

Within the study area (fig. 2), the Late Triassic Loomis pluton is cut by the Toats Coulee pluton on the west and the Anderson Creek pluton on the east. Coexisting hornblende and biotite from the Toats Coulee were dated at 170 m.y. and 151 m.y., respectively, and coexisting hornblende and biotite from the Anderson Creek were dated at 115.6 m.y. and 100.1 m.y., respectively (sample L–451C, fig. 2 and table 2). The discordance between ages of coexisting minerals in other plutons increases dramatically to the north and east of the Loomis. The discordance between hornblende and biotite ages from some samples of the Similkameen batholith and the coeval Kruger Alkalic Complex is as much as 106 m.y., hornblende being invariably the older of the mineral pair. In some samples of the Similkameen and Kruger, hornblende shows maximum ages of 155 to 177 m.y. (table 2).

In fact, all the other plutons within the study area (fig. 2) from which coexisting minerals have been dated show significant discordance, except for the Swimptkin Creek pluton of Eocene age, which yielded hornblende and biotite ages of 48.2 m.y. and 48.0 m.y., respectively. The data summarized in table 2 suggest that the plutonism in the study area, which had begun with intrusion of the Loomis pluton, continued sporadically through the Jurassic, Cretaceous, and early Tertiary.

DISCORDANT AGES

The discordancy between apparent ages of coexisting minerals reaches a maximum in a zone that borders the Okanogan gneiss dome on the west (fig.5).

The Okanogan gneiss dome occupies 2,500 km^2 (950 mi^2) within the study area (fig. 2). Rocks of the gneiss dome previously have been considered—at least in part—as part of the Colville batholith (Pardee, 1918; Waters and Krauskopf, 1941; Staatz, 1964). The gneiss dome consists chiefly of layered paragneiss, augen gneiss, and granodiorite, marked over most of the area by a penetrative cataclastic fabric whose chief elements are a very regular low-dipping foliation and a persistent west-northwest-trending lineation. Country rocks adjacent to the dome at the western, northwestern, and southwestern contacts have been crushed and broken, and rocks of the dome at and near its upper surface along the western contact are mylonitized.

The mylonitization and gneissic fabric of the gneiss dome and the brecciation of the wallrock were attributed by Waters and Krauskopf (1941) to emplacement of the mass as a protoclastic batholith. However, Snook (1965, p. 775) concluded that these features resulted from development of a distributed flat thrust at depth within a regionally metamorphosed sedimentary or volcanic terrane that was later upfaulted against the lower grade metamorphic rock on the west. We have suggested a third alternative—that the gneiss dome formed through metamorphism of volcanic and sedimentary strata at depth, a metamorphism that culminated in the mobilization and diapiric emplacement of the dome (Fox and Rinehart, 1971).

Ages from within the gneiss dome are themselves discordant, with the 21 age determinations previously reported (Fox and others, 1976) ranging from 100 m.y. ($^{207}Pb/^{235}U$, zircon) down to 46.0 m.y. (K–Ar, biotite).

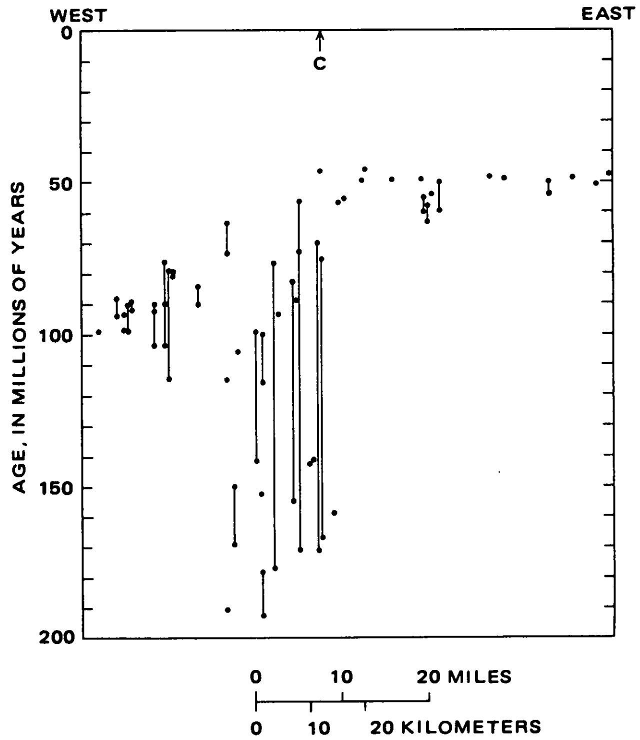


FIGURE 5.—Plot of age determinations (exclusive of uranium-thorium-lead ages of sample 0–176E, table 2) within study area (fig. 2) projected to east-west transect. Ages of coexisting minerals connected by vertical lines. Rb–Sr ages disregarded where less than fission-track ages of coexisting apatite, sphene, or zircon. Westward extremity of Okanogan gneiss dome at point *C*.

The terminal stages of metamorphism and emplacement of the dome are believed to have been Late Cretaceous, after which the dome cooled slowly through the successive temperature thresholds (blocking temperatures) of the dated minerals, which included zircon, sphene, epidote, allanite, hornblende, apatite, muscovite, and biotite. The possibility that these minerals were variably degraded as a result of later mild reheating of the dome during a thermal event associated with regionwide Eocene volcanism can be evaluated through comparison of age data from elsewhere within the region.

A compilation of age data shows that Tertiary ages are abundant within the three orogenic provinces represented within the region (figs. 1, 3)—from east to west, the Purcell foldbelt, the Omineca crystalline belt, and the Columbian intermontane belt. However, pre-Tertiary ages, which are abundant within both the Purcell foldbelt and the Columbian intermontane belt, are strikingly rare in the Omineca crystalline belt. The contrast is apparent in figure 6, a plot of ages projected to an idealized east-west transect across the region shown in figure 3.

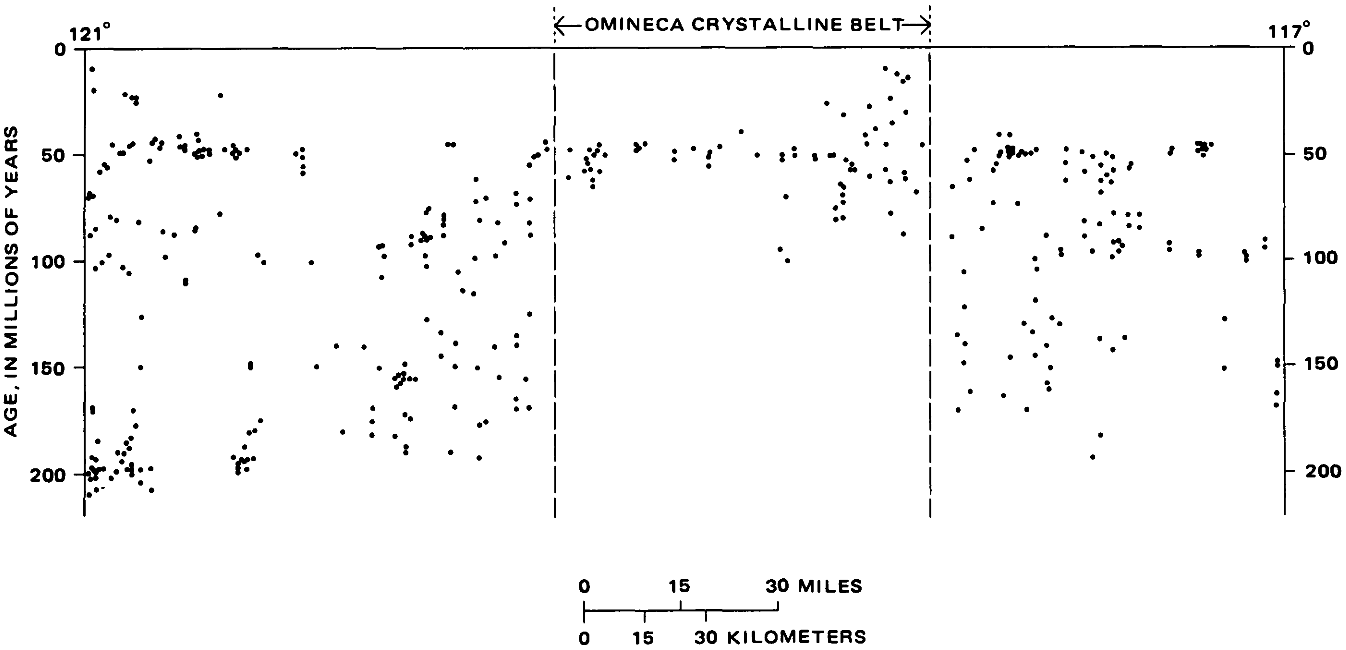


FIGURE 6.—Mesozoic and Cenozoic age determinations within region shown in figure 3 plotted with respect to the approximate boundaries of the Omineca crystalline belt and projected to the 49th parallel.

Apparently the zone of discordant ages parallel to the west side of the Okanogan gneiss dome (fig. 5) is part of a regional pattern of discordance associated with the Omineca crystalline belt.

AGE OF OMINECA CRYSTALLINE BELT

The age of the Omineca crystalline belt refers to the ages of the lithologic elements by which it is defined—the Monashee Group of the Shuswap terrane of Jones (1959) (and correlative rocks) and the gneiss domes included within that terrane. Three ages must be considered: (1) the age of the parent material; (2) the age of metamorphism, deformation, and mobilization; and (3) the age of cooling through the blocking temperatures of the dated minerals.

The Okanogan gneiss dome is one of a family of gneiss domes located within the Shuswap—or Shuswap-like—terrane of the Omineca. These include, from north to south, Frenchman's Cap (Wheeler, 1965), Thor-Odin (Froese, 1970; Reesor and Moore, 1971), Pinnacles (Reesor and Froese, 1968), Valhalla (Reesor, 1965), and Passmore (Reesor, 1965) gneiss domes (figs. 1, 3).

As redefined by Jones (1959), the Shuswap, in addition to the Monashee Group, also includes on the west flank two outliers of low-grade metamorphic rocks named the Mount Ida Group and the Chapperon Group. Their lower metamorphic grade and lack of evidence of profound tectonic involvement set them in marked contrast with rocks of typical Monashee terrane, and for that reason they are here considered part of the Columbian intermontane belt, rather than part of the Omineca. The Mount Ida Group is also lithologically dissimilar to the Monashee Group (Jones, 1959, p. 30). Thus, the Mount Ida and Chapperon Groups may not share a common age and origin with the Monashee Group. Hence, age constraints imposed on the Mount Ida and Chapperon Group do not necessarily apply to the Monashee. We refer to the K–Ar biotite ages of 135, 140, and 140 m.y. (samples GSC 61–1, 2, 3; Lowdon and others, 1963) from rocks of the Mount Ida Group and to the fact that the Mount Ida and Chapperon both are unconformably overlain by the Permian Cache Creek Group—the Mount Ida at the Glenemma(?) unconformity (Jones, 1959, p. 48) and the Chapperon at the Salmon River and Dome Rock unconformities (Jones, 1959, p. 28, 29)—thus placing firm younger age limits on both groups. Again, these restrictions do not place a younger limit on either the age of the parent material or the age of metamorphism and mobilization of the gneiss domes or the Monashee Group.

The material composing the Frenchman's Cap dome lithologically resembles part of the late Precambrian, Cambrian, and post-Cambrian succession (Wheeler, 1965, p. 9), as does that of the Thor-Odin (Froese, 1970, p. 173; Reesor and Moore, 1971, p. 113). Reesor (1970, p. 85) suggests that Pinnacles dome may be composed in part of rocks equivalent to the Milford Group (Mississippian to Triassic?), and Giovanella (1968) suggests that the Malton gneiss may include small bodies of late Precambrian Kaza rocks. Reesor (1970, p. 85) notes that changes in sedimentation from Windemere pelite, grit, and pebble conglomerate to Cambrian quartzite, pelite, and limestone can be tentatively identified by parallel changes in composition of rocks within the gneiss domes. Thus, the age of the material composing some of the domes appears to be late Precambrian to late Paleozoic.

The age of mobilization and emplacement of the domes has not been narrowly bracketed. Rocks referred to the Shuswap Series east of the Frenchman's Cap dome in the "Clachnacudainn Salient" are cut by a stock giving a K–Ar biotite age of 110 m.y. (Wheeler, 1965, p. 15), and north-trending joints in the dome itself are occupied by lamprophyre dikes dated at 41 m.y. (Wheeler, 1965, p. 16).

Unconformable contacts of the Cache Creek Group over the Monashee Group at Lavington and at B. X. Creek were reported by Jones (1959, p. 47–48), but the contacts have been reinterpreted as probable faults by Preto (1965). The Monashee Group, however, is patchily overlain by early Tertiary epiclastic deposits and volcanic rocks (Jones, 1959, p. 52).

The Monashee Group has been contact metamorphosed by a 69-m.y.-old pluton in the Nakusp area (Hyndman, 1968b, p. 68–69). Hyndman (1968a) concluded that the age of deformation of the Monashee Group in the Nakusp area was the same as that of the adjacent rocks of the Triassic Slocan Group because the orientation of the most prominent axis of folding of schistosity and cleavage is similar in both groups. Hyndman noted that the foliation in the Slocan predates contact metamorphism by Cretaceous plutons and therefore concluded that the age of this foliation, and hence the age of metamorphism and deformation of both the Slocan and Monashee Groups, is Jurassic.

A contrary view—namely that the main deformation of the Monashee Group did not involve the Slocan Group—has been put forward by Ross (1970). He suggests that in the Kootenay arc three phases of deformation can be discerned, each producing morphologically distinct folds and associated foliation and lineation. The earliest folds are isoclinal and formed easterly verging allocthonous nappes cored with Shuswap gneiss. Since in his opinion these folds affect only pre-Slocan rocks, he believes they are of pre-Triassic age. They were refolded during a second phase of deformation, also of pre-Triassic age, and again during a third and final phase of deformation. The folds produced in the third phase of deformation are open structures that involve rocks in the Slocan Group, as well as older rocks. Since the deformed Slocan is intruded by the undeformed Nelson batholith, isotopically dated at about 164 m.y. (Nguyen and others, 1968), this third phase of deformation is bracketed between Late Triassic and Early Jurassic. Ross attributes the refolded isoclinal folds of rocks of the Slocan Group, reported by Fyles (1967, p. 34–35) and Hyndman (1968a), to local deformation associated with emplacement of the Kuskanax and Nelson batholiths.

The K–Ar ages of 17 biotite samples from the Valhalla gneiss dome range from 11 to 66 m.y. (Reesor, 1965, p. 51), and K–Ar ages of both biotite and hornblende from the core zone of the Thor-Odin gneiss dome range from 60 to 70 m.y. (Reesor, 1970, p. 86).

K–Ar ages of 15 samples of the Shuswap Series (Monashee Group) reported by the Geological Survey of Canada are as follows:

Age in m.y.	*Dated mineral*	*Sample number*	*Reference*
62	Biotite	GSC 60–1	Lowdon, 1961
57	Do.	61–4	Lowdon, Stockwell, Tipper, and Wanless, 1963
52	Do.	61–5	Do.
71	Do.	61–6	Do.
102	Do.	61–7	Do.
81	Muscovite	61–8	Do.
89	Biotite	62–36	Leech, Lowden, Stockwell, and Wanless, 1963
64	Do.	62–35	Do.
70	Do.	62–44	Do.
65	Muscovite	62–45	Do.
73	Do.	62–46	Do.
76	Biotite	62–47	Do.
81	Do.	62–48	Do.
61	Hornblende	66–43	Wanless, Stevens, Lachance, and Edmonds, 1968
79	Do.	66–44	Do.

The apparent cooling age of the Monashee Group thus spans an interval from middle Cretaceous to Eocene, overlapping the cooling age of the gneiss domes. However, the average of the ten biotite ages listed above of about 72 m.y. significantly exceeds that of the six biotite ages from the Okanogan gneiss dome (table 2), which range from 46 m.y. to 56.8 m.y. and average about 50 m.y.

One striking feature of the Omineca crystalline belt is the contrast in metamorphic grade between the Monashee and Monashee-like rocks and the neighboring Permian and Triassic metavolcanic and metasedimentary rocks. Jones (1959, p. 130) notes that "unmetamorphosed Cache Creek rocks in many places lie in direct contact with sillimanite gneisses of the Monashee Group***." The transition from fine-grained greenchist facies country rock to coarsely recrystallized amphibolite facies rocks of the Okanogan gneiss dome is also abrupt.

Rocks typical of the Okanogan dome near this contact are mylonitized quartz-oligoclase-orthoclase-biotite gneiss or diopside-hornblende-plagioclase-biotite gneiss, or granodiorite. Alumino-silicate minerals are rare, but locally, toward the interior of the dome, the gneiss is composed of sillimanite, cordierite, orthoclase, biotite, garnet, quartz, and plagioclase and in places contains accessory muscovite. This assemblage is referable to the sillimanite-cordierite-muscovite-almandine subfacies of the amphibolite facies, Abukuma-type facies series (Winkler, 1967, p. 121), indicating metamorphism under moderate pressures, estimated by Winkler (1967, p. 187) as 3–5 kilobars.

Country rocks at the west contact show little evidence of contact metamorphism, a point emphasized by Waters and Krauskopf (1941, p. 1376). However, there is a metamorphic halo from 100 m (110 yd) to at least 2 km (1.2 mi) wide bordering the dome on the north and northeast, within which garnet, biotite, and staurolite are present. Nevertheless it is clear from the contrast in metamorphic grade at the contact that the rocks of the dome could not have been metamorphically formed in situ.

The rocks 10 km (6 mi) west of Curlew (fig. 3) are an exception to the above generalization that there is a sharp contrast in metamorphic grade at contacts of the Shuswap (Monashee Group) and its country rock. Parker and Calkins (1964, p. 5–24) described these rocks as the metamorphic rocks of Tenas Mary Creek and suggested their correlation with the Shuswap (Monashee Group). The Tenas Mary Creek rocks form a homoclinal sequence estimated by them to be about 5,200 m (17,000 ft) thick that dips northward about 25°. The sequence shows a gradual stratigraphic upward decrease in degree of metamorphism from orthoclase-quartz-oligoclase gneiss at the base, through sillimanite-, cordierite-, and muscovite-bearing schist and gneiss, to sericite-, chlorite-, and biotite-bearing phyllite at the top, where the sequence is conformably overlain by Permian to Triassic greenstone.

Presuming that the processes of metamorphism and penetrative deformation were at least in part simultaneous with emplacement of the Okanogan gneiss dome, the contact relations with the Triassic Cave Mountain Formation and the Eocene sedimentary and volcanic rocks place older and younger limits on the age of these processes. The cataclastic texture prevalent throughout much of the dome suggests that these processes continued well after most mineral constituents, including zircon, had crystallized. Thus, the 87-m.y. ($^{206}Pb/^{238}U$) age of zircon from one sample (table 2) is regarded as an older limit on the terminal age of penetrative deformation and mobilization of the gneiss (Fox and others, 1976).

The distribution pattern of the Eocene volcanic rocks, which are widely scattered over the entire region shown in figure 3, does not coincide with that of the Omineca crystalline belt or its bordering zone of highly discordant ages. Therefore we see no correlation between the present outcrop areas of the Eocene rocks and the belts of discordant ages and conclude that the thermal event presumably associated with the formation of these volcanic rocks is not a factor, except perhaps locally, in the origin of the discordancies. Thus, the Shuswap-like rocks of the Omineca crystalline belt probably cooled through the successive blocking temperatures for the minerals dated in Late Cretaceous and early Tertiary as a result of uplift and progressive unroofing by erosion.

The many gneiss domes and the Shuswap (Monashee) are products of an intense regional deformational and metamorphic event that was localized along the Omineca crystalline belt. It follows that the discordance in the ages of plutonic rocks bordering the Omineca is at least in part a manifestation of thermal degradation caused by reheating during this deformational and metamorphic event. Locally, however, the effect of contact metamorphism by younger plutonic rocks seems to be a factor in the origin of the discordance. For example, east of the Omineca crystalline belt in northern Washington and Idaho, the biotite and muscovite ages of pre-Tertiary plutonic rocks show increasing degradation towards areas of extensive Eocene plutonic rocks (Miller and Engels, 1975).

The Omineca metamorphic event could have begun in British Columbia prior to 110 m.y. ago, if the K–Ar age determination of biotite in the stock cutting the Shuswap in the Clachnacudainn Salient reported by Wheeler (1965, p. 15) is accurate. In the Okanogan area, that event persisted into the Late Cretaceous, when both the mobilized elements of the metamorphosed rocks and also a thermal front of sufficient intensity to affect the K–Ar age of some minerals in the rocks of the terrane west of the belt reached high levels in the crust. The metamorphic event had been concluded and the high-grade metamorphic rocks had cooled through the blocking temperature of biotite by the end of the Cretaceous in British Columbia and by Eocene in Washington.

EOCENE TO MIOCENE HISTORY

The Tertiary volcanic rocks of the region (fig. 3) have been shown by Mathews and Rouse (1963), Rouse and Mathews (1961), Mathews (1964), and Hills and Baadsgaard (1967) to include two groups, one Eocene and the other late Miocene and early Pliocene. Most of the Eocene volcanic rocks within the region were formed during a thermal episode climaxing about 50 m.y. ago.

The Eocene episode began with roughly contemporaneous deposition in local basins of arkose, wacke, and conglomerate. At the base, these deposits consist of quartzo-feldspathic material eroded from nearby sources, but higher in the section these deposits contain pyroclastic material in increasing proportions. The sedimentary beds are typically overlain by pyroclastic rocks and lava flows and intruded by the hypabyssal intrusive equivalents of the volcanic rocks. The Eocene rocks were deposited on a profound angular unconformity beveled on older rocks.

Eocene volcanic rocks in the southeastern and northeastern parts of the study area within Wasington include both the flows of quartz latite, rhyodacite, dacite, and interbedded tuff of the Sanpoil Volcanics and their rhyodacitic and quartz latitic hypabyssal intrusive equivalents, the Scatter Creek Formation (Muessig, 1962; Staatz, 1964; Parker and Calkins, 1964; Muessig, 1967). Coextensive rocks in British Columbia include andesite, trachyandesite, and sodic trachyte of the Marron Formation (Monger, 1968). Correlative volcanic rocks to the west include dacitic and andesitic flows and their hypabyssal intrusive equivalents, which are patchily distributed along the axis of the Okanogan Valley and also occupy a large tract farther west underlying the much dissected summit plateau of the Okanogan Range.

A fossil flora in sedimentary deposits cut by intrusive equivalents of the volcanic rocks northwest of Oroville is regarded as indicating a probable early Eocene age by J. A. Wolfe (in Rinehart and Fox, 1972, p. 61). The fossil flora of the O'Brien Creek Formation, a volcaniclastic unit that underlies the Sanpoil Volcanics near Republic (fig. 3), a few kilometers east of the study area, resembles the Eocene flora of Alaska and of the lower part of the Puget Group of Washington, whereas the fossil flora of the Klondike Mountain Formation, which overlies the Sanpoil, is Oligocene and Miocene(?) (R. W. Brown, *in* Muessig, 1967).

K–Ar ages of volcanic rocks within the study area, previously reported by Rinehart and Fox (1972), Mathews (1964), and Engels, Tabor, Miller, and Obradovich (1976) range from 42.9 to 52.1 m.y.

In the Republic area, Eocene deposition was accompanied by penecontemporaneous faulting and subsidence of a north-northeast-trending graben (Muessig, 1967, p. 95–96). There and elsewhere in the region the Eocene rocks were folded and faulted, partially eroded, and later buried in places by basalt during the late Miocene. The late Miocene and early Pliocene rocks include the plateau basalts of the Columbia River Basalt Group, which overlaps the south edge of the region, and its erosional outliers, and other smaller patches of basalt distributed over the remainder of the region.

OROGENIC IMPLICATIONS
GEOLOGIC HISTORY

PREVIOUS VIEWS ON OROGENY IN THE CORDILLERA

Phanerozoic orogeny in the Cordillera was visualized by Gilluly (1965) as a virtually continuous, albeit locally episodic, process. King, while holding that in general orogeny "was concentrated in a succession of episodes during each orogenic phase***" (1969b, p. 45), concluded that in the northern Cordillera (north of the 49th parallel), Mesozoic time "was one of nearly continuous orogeny from place to place in the eugeosynclinal area***" (1969b, p. 67–68). King concludes that climactic middle Mesozoic orogeny in the central Cordillera (for example, the Sierra Nevada) was in Late Jurassic but was either earlier or later in other parts of the eugeosynclinal belt. Also, it was "progressively younger eastward across the fold belt, and in the miogeosynclinal area there is no clear separation between a middle Mesozoic orogeny and a terminal Mesozoic orogeny; most of the events were at intermediate times***" (1969b, p. 70).

White (1959) attributes Mesozoic and early Cenozoic deformation in British Columbia to three orogenies: the Cassiar orogeny, which occurred between Permian and Late Triassic (p. 72); the Coast Range orogeny, whose effects first appear in the earliest Jurassic, culminate in the Early Cretaceous, and persist well into the Late Cretaceous (p. 78); and the Rocky Mountain orogeny, which culminated in Paleocene (p. 98). White later modified these views, stating (White, 1966, p. 187) that "much of Mesozoic was a time of what might be termed, 'continuous-intermittent orogeny,' when orogenic events of one sort or another were happening in one part or another of the Western Cordillera throughout a span of perhaps 150 million years." He further states (p. 189) that "it is doubtful that Tertiary orogeny was separated from Mesozoic orogeny by any significant period of crustal stability ***." Within the study area, the Cassiar orogeny may be responsible for the unconformable contacts of Kobau over Anarchist (Rinehart and Fox, 1972), but the degree of deformation is apparently slight, probably of epeirogenic rather than orogenic proportions. No pre-Late Triassic plutonic rocks have been discovered.

LATE TRIASSIC OROGENY AND PLUTONISM

The Upper Triassic Nicola Group is cut by the Guichon batholith (White, 1959, p. 87), now known to be about 200 m.y. old (White and others, 1967). Within the study area (fig. 2), strongly folded rocks of the Kobau

Formation, a probable correlative of the Nicola Group, are similarly cut by the Loomis pluton, believed to be about 195 m.y. old (Rinehart and Fox, 1972; this report, table 2). Thus, orogeny (as defined in Gary and others, 1972, p. 500) probably occurred simultaneously in southern British Columbia and northern Washington during Late Triassic. This event evidently has no chronological counterpart in the Sierra Nevada or in the Klamath Mountains of California. The earliest comparable Mesozoic event in the Sierra Nevada was in Early or Middle Triassic, and the next was in Early or Middle Jurassic (Evernden and Kistler, 1970). Mesozoic plutonism in the Klamath Mountains began in early Middle Jurassic, where it ushered in the Nevadan orogeny (Lanphere and others, 1968, p. 1047, 1050).

The ages of plutonic rocks in the Okanogan region suggest that the eugeosynclinal province underwent sporadic plutonism without a clearcut break between the Late Triassic and early Tertiary. However, the plutonism subsequent to Late Triassic and prior to emplacement of the Okanogan gneiss dome cannot be linked with orogenic deformation in the Okanogan region by means of the evidence at hand.

ASSOCIATION OF DEFORMATION OF THE MONASHEE GROUP AND GENERATION OF GNEISS DOMES WITH CORDILLERAN THRUST FAULTING

The most important thermal and deformational—hence, orogenic—event within the region was the metamorphism climaxed by mobilization of large segments of the lithosphere within and adjacent to the Omineca crystalline belt. Understanding of this event seems pivotal to an understanding of the geologic history of the region.

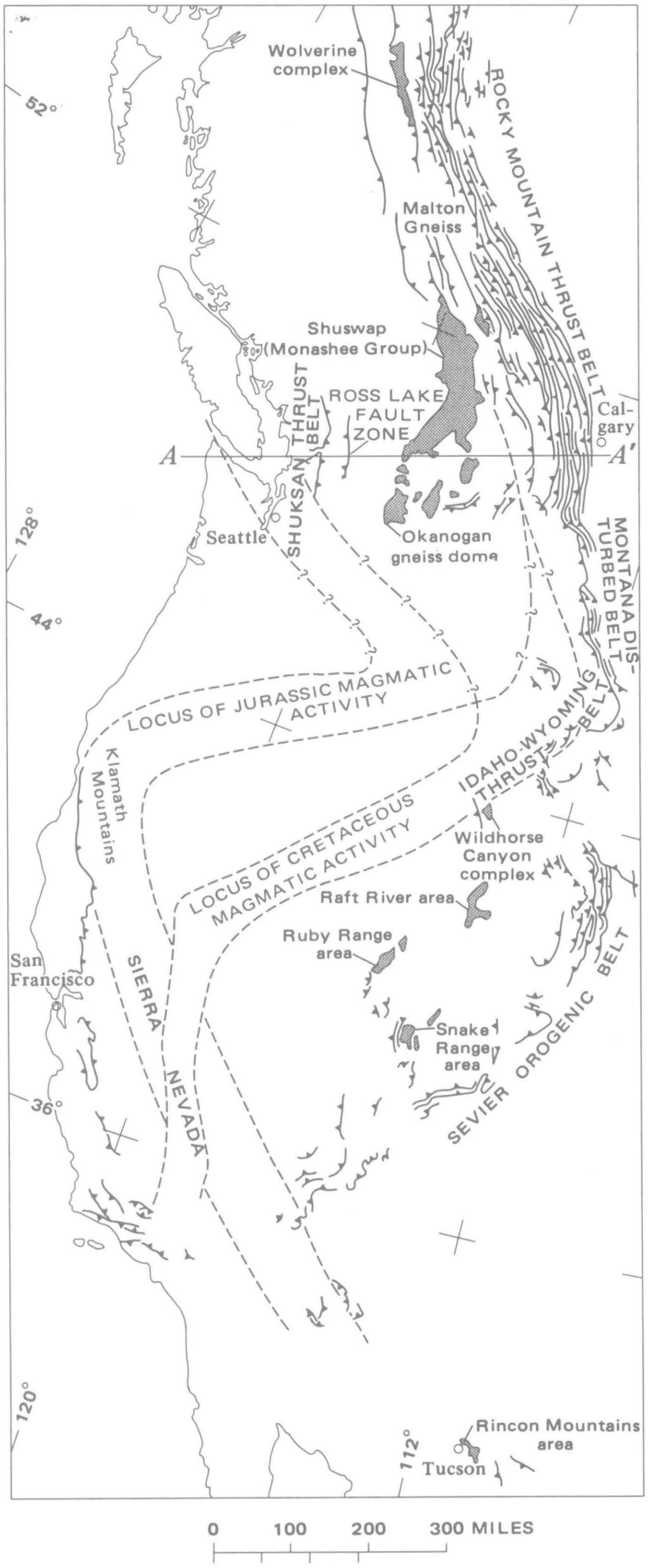
The deformation and mobilization of the Shuswap (Monashee Group) and correlative rocks may be linked to thrusting along the Rocky Mountain thrust belt to the east and along the Shuksan thrust belt of the northern Cascade Range to the west (fig. 7). According to Bally, Gordy, and Stewart (1966, p. 366–372), deformation proceeded from west to east, beginning with mobilization of gneiss domes and thrusting in the interior in Late Jurassic to Early Cretaceous and, except for later uplift, ending with the final stages of thrusting in the Rocky Mountains Foothills province in the Eocene. Price and Mountjoy (1970) postulated contemporaneous metamorphism and mobilization of the gneiss domes and thrusting in the Rocky Mountain thrust belt to the east. In their view, the origin of the Shuswap and the thrust belts can be explained by "progressive buoyant upwelling and lateral spreading of a hot mobile infrastructure beneath a relatively passive suprastructure*** and equivalent progressive northeasterly growth of the foreland thrust belt that forms the Rocky Mountains" (Price and Mountjoy, 1970, p. 7).

The crystalline core of the northern Cascade Range is flanked to the west and east by major fault zones, the Shuksan thrust and associated faults on the west and the Ross Lake fault zone on the east (Misch, 1966). The Shuksan thrust is part of Misch's (p. 128) "mid-Cretaceous Northwest Cascades System." The age of this westward thrusting is bracketed between "that of the Nooksack Group (early Early Cretaceous) and that of the Chuckanut Group (late Late Cretaceous and Paleocene)" (McTaggart, 1970, p. 138, 144).

The terrane between these two major faults consists mainly of the Cascade River Schist and the Skagit Gneiss, believed by Misch (1966, p. 103, 113) to have been metamorphosed before the Shuksan thrusting of the mid-Cretaceous orogeny in pre-Jurassic or probably pre-Mesozoic time. However, McTaggart (1970, p. 143) suggested that the Skagit migmatites and their correlative in Canada, the Custer Gneiss, might be Early to middle Cretaceous in age and closely related to the mid-Cretaceous orogeny. Mattinson (1972, p. 3778–3779) obtained ages ranging from 90 to 60 m.y. on sphene and zircon from the Skagit Gneiss and correlative rocks by Pb-U and Pb-Pb methods, indicating that the Skagit metamorphism was probably middle to Late Cretaceous. In view of the relative ages of Skagit metamorphism and Shuksan thrusting, implied by Misch (1966), it is unlikely that the Shuksan thrusting is older than Late Cretaceous.

According to Misch (1966, p. 128) yielding on the Shuksan and associated thrusts is southwestward on the south, westward near the Border, and northwestward north of it. Movement on the Ross Lake fault zone was primarily strike slip, although reverse in part with eastward-directed over-thrusting on the subsidiary Jack Mountain fault of up to about 10 km (6 mi) displacement (Misch, 1966, p. 134). The northern extension of the Ross Lake fault zone, the Hozameen fault, seems to be very steep (McTaggart, 1970; Coates, 1970). Slivers of tectonically emplaced ultramafic rock are associated with both the Shuksan and Ross Lake fault systems, causing Misch (p. 133, 134) to postulate extension of these faults to the base of the crust. In Misch's view (p. 136), a 40-km-wide (25-mi-wide) wedge of the crystalline core of the Cascades has been compressively uplifted between the two faults.

Maximum offset of the Shuksan and associated thrusts was estimated as about 50–65 km (30–40 mi) (Misch, 1966, p. 128), whereas the cumulative offset on the Rocky Mountain thrusts was variously estimated as about 190 km (120 mi) by Bally, Gordy, and Stewart (1966, p. 359) and at least 200 km (125 mi) by Price and Mountjoy (1970, p. 16). The Shuswap complex and as-



sociated gneiss domes thus appear to occupy the axial zone between convergent thrusts that show an apparent aggregate crustal contraction of possibly 250 km (150 mi).

The Shuswap (Monashee Group) is but one of the several areas of gneiss domes, or of medium- to high-grade metamorphic rock, some exhibiting Late Cretaceous to early Tertiary cooling ages, that flank the interior (western) side of the Cordilleran thrust belt in British Columbia, Washington, Idaho, Nevada, and Utah (fig. 7). These also include the Wolverine complex; Wildhorse Canyon Complex (Dover, 1969); the metamorphic complexes in Idaho, Utah and Nevada grouped by Armstrong and Hansen (1966) into three areas, namely Raft River, Ruby Range, and Snake Range; and the metamorphic complex in the Rincon Mountains area of Arizona (Waag, 1969). We speculate that some of these metamorphic terranes are either (1) uplifted elements of an abscherungzone (Armstrong and Hansen, 1966), (2) diapiric tongues of gneiss, possibly originating at an abscherungzone and intruding supracrustal strata of the overthrust plate, or (3) gneiss domes forming through mobilization of infrastructure farther to the west in the hinterland. The metamorphic terranes appear to be both temporally and geographically associated with the Late Cretaceous to early Tertiary thrust belts.

These relations compel the conclusion that the formation of the Shuswap (Monashee Group) and the gneiss domes is the regional expression of the widespread deformational and metamorphic events that culminated in the formation to the west of the westward-directed Shuksan thrust, and to the east, as advocated by Balley, Gordy, and Stewart (1966) and Price and Mountjoy (1970), of the belt of easterly directed Rocky Mountain thrusts.

LARAMIDE OROGENY, FORMATION OF GNEISS DOMES, AND THRUST FAULTING

In concept, the geographic locus of a deformational event could gradually shift through time so that the beginning and ending of the event would not be everywhere the same age. However, few Mesozoic and

FIGURE 7.—Distribution of (1) late Mesozoic and early Cenozoic thrust belts and (2) areas of medium- to high-grade metamorphic rock, including gneiss domes, that show late Mesozoic to early Cenozoic cooling ages and (or) structural fabrics suggesting extensive mobilization. Belts of Jurassic and Cretaceous magmatic activity modified after Kistler, Evernden, and Shaw (1971, p. 858). Thrust faults after King (1969a). Section $A–A'$ shown in figure 8.

Cenozoic deformational events have left a structural signature by which they can be correlated from region to region. The Laramide orogeny may be an exception. Eardley (1951, p. 285) noted that Mesozoic movements in the Rocky Mountains systems "although intense in some places, were generally precursory to the climactic ones at the close of the Cretaceous and in the early Tertiary." He indicated that the Laramide orogeny was "the great compressional disturbance that occurred in very late Cretaceous, Paleocene, and Eocene." It follows that thrusting of that approximate age along the nearly continuous Rocky Mountain thrust belt in Canada, the Montana disturbed belt, Idaho-Wyoming thrust belt, and the Sevier belt in Nevada and Utah, is referable to the Laramide orogeny, regardless of minor differences in age of thrusting from region to region.

This deduction contravenes Armstrong's (1968, p. 451) conclusion that thrusting in the Sevier belt predates and therefore is not part of the Laramide orogeny. Noting that thrusting in the Sevier belt ended in the Campanian Age of the Late Cretaceous, he introduced the term Sevier orogeny for that deformational event and relegated the Laramide orogeny to the succeeding episode of morphogenic uplift ranging in age from the Maestrichtian or Campanian Ages of Late Cretaceous through the middle Eocene. We agree that the term Sevier orogeny should be retained but suggest that it is a regional correlative of the Laramide orogeny as defined by Eardley (1951).

The deformational events that resulted in formation of the gneiss domes, Shuswap terrane, and thrust belts therefore include tectogenesis attributed to mid-Cretaceous orogeny (Misch, 1966) on the west and to Laramide orogeny on the east.

Price and Mountjoy (1970, p. 23) suggested that the thrusting and folding in the Canadian Rocky Mountains spanned the time interval from Late Jurassic to Paleocene or Eocene. Their older limit is apparently based on the supposition that orogenic development in the interior of the Rocky Mountains and deposition of the sequence of clastic wedges to the northwest had begun by Late Jurassic and on the premise that formation of the clastic wedges implies beginning of thrusting as well as orogeny. In their view, a minimum age for thrusting and doming of the gneiss of 111 m.y. is established by the age of porphyroblastic biotite (sample GSC 66–47, Wanless and others, 1968) from metamorphic rocks believed to be part of the metamorphic halo that envelopes the diapiric Malton Gneiss (see fig. 7), more than 150 km (95 mi) north of the area represented in figure 3. However, the following K–Ar ages have been reported from the gneiss itself:

Mineral	*Age, in m.y.*	*Sample number*	
Biotite	72	GSC 65–24	Wanless, Stevens, Lachance, and Edmonds, 1967
Biotite	53	67–43	Wanless, Stevens, Lachance, and Delabio, 1970
Biotite	59	67–43	Do.
Hornblende	114	67–44	Do.
Muscovite	60	70–16	Wanless, Stevens, Lachance, and Delabio, 1972
Biotite	66	70–17	Do.
Biotite	57	70–18	Do.

The cooling age of mica in the Malton Gneiss is roughly Paleocene, virtually equivalent to that of the gneiss domes to the west.

Armstrong and Oriel suggested that thrusting in the Idaho-Wyoming thrust belt spanned the interval from Late Jurassic to Eocene (1965, fig. 19, p. 1861). Evidence of thrusting prior to latest Cretaceous apparently consists of the presence of a latest Jurassic to earliest Cretaceous conglomerate 24 km (15 mi) east of the trace of the Paris thrust fault. Armstrong and Oriel reasoned that if the conglomerate is a synorogenic deposit derived from source areas to the west, as is likely, then movement on the Paris thrust is dated as latest Jurassic and earliest Cretaceous (1965, p. 1859).

Although the time at which the Laramide orogeny ended is relatively easily established by the cooling ages of the crystalline rocks and the termination of overthrusting, the time at which this orogeny began is more difficult to establish and is partly a matter of definition. We suppose that the Shuswap (Monashee Group) and correlative rocks include metamorphosed equivalents of Precambrian, Paleozoic, and even Mesozoic supracrustal rocks that were converted to infrastructure—that is, were deeply buried, heated, metamorphosed, migmatized, and ultimately formed rheomorphic bodies that later were forced upward into the suprastructure, much as visualized by Campbell (1973, p. 1614). This process may have begun in places in the region as early as the Late Triassic, which may mark a significant milestone in the orogenic cycle. However, the gneiss domes are set apart not by migmatization or metamorphic grade, but by internal structure and contact relations suggesting rheomorphic flow and diapiric intrusion. The time at which wholesale upward and lateral flow of parts of the infrastructure began therefore marks the beginning of this deformational event in the Okanogan region. The evidence within this region for the time at which this phase began is nebulous, principally because of difficulty in discriminating effects of earlier phases of metamorphism and deformation.

The earlier phases of plutonism and metamorphism may have been a necessary precursor to the later invasion of supracrustal rocks by gneiss domes and concomitant overthrusting. However, the diapirism and thrusting seemingly mark a radical, even catastrophic, change in the tenor of compressional tectonics and thereby stand in contrast to the previous history. The rough simultaneity in the cooling age of the gneiss domes and the age of terminal thrusting indicates a similar equivalence in the age of their birth, assuming that the thrusts and gneiss domes are genetically related and that both originate at least in part in response to a particular episode of regional compression. If the gneiss domes and thrusts are indeed roughly coeval, the probable older limit on the age of the Shuksan thrust of Late Cretaceous suggests a comparable older limit on the age of the gneiss domes.

DISCUSSION

A stable—and perhaps at times extensional—margin bordered the continental plate during building of the late Precambrian and early Paleozoic depositional wedge now exposed in the miogeoclinal province (Stewart, 1972; Gabrielse, 1972, p. 533). The tectonic stability that marked this period was succeeded in late Paleozoic or early Mesozoic time by the strongly convergent regime that prevailed throughout much of the Mesozoic and Cenozoic, while large areas of oceanic plate were subducted beneath the western side of the North American plate (Hamilton, 1969b).

The late Paleozoic Cache Creek strata may represent sediments deposited upon the oceanic plate seaward of an eastwardly dipping subduction zone (Monger and others, 1972, fig. 6). However, the temporal counterpart of these strata in the study area, the Anarchist Group, contains limestone and clastic detritus more compatible with sedimentation in a back-arc basin close to a nearby landmass than with sedimentation far from land at the abyssal depths of the ocean floor. In any case, the predominantly volcanic strata of the overlying Middle to Late Triassic formations and their probable correlative in the study area, the Kobau Formation, mark the abrupt onset of a period of (1) extrusion of voluminous lavas, (2) intrusion of batholiths and lesser plutons, (3) strong folding, (4) uplift and subaerial erosion, and deposition of detritus in successor basins, and (5) regional metamorphism. Most of this activity can be ascribed to processes occurring above an eastwardly dipping subduction zone or zones, located along the continental margin to the west (Monger and others, 1972, p. 592–593).

Perhaps the most enigmatic of the problems discussed in the preceding pages are those of the origin of the gneiss domes and the thermal metamorphism of the Omineca crystalline belt and the reason for the apparent association of these features with the Cordilleran thrust faults.

In the Canadian Cordillera, the spatial and temporal association of the gneiss domes and thrust faults with the Mesozoic plutonic belts and with the presumed convergent plate boundary suggest that the gneiss domes and thrust faults are in some way related to Mesozoic subduction to the west (Campbell, 1973, p. 1618). But it is puzzling that south of the Canadian Cordillera the belt of gneiss domes and the track of the eastern thrust belts diverge from parallelism with the continental boundary (as marked roughly by the present coastline, fig. 7) and swing far inland of the Jurassic and Cretaceous belts of plutonic activity delimited by Kistler, Evernden, and Shaw (1971, p. 858).

Coney (1972, p. 620) stresses the probable tectonic impact of the worldwide change in spreading directions that occurred about 80 m.y. ago. He proposed (1972, p. 619–620) that the Late Cretaceous orogeny in the Sevier belt of Armstrong (1968) resulted from failure of the leading edge of the North American plate as it was being driven northwestwardly over a Benioff zone entrained in the oceanic plate below. Later, after the change in spreading directions, the early Tertiary orogeny resulted from failure of the leading edge of the North American plate as it was driven westwardly to southwestwardly, also over a subduction zone or zones entrained in the subjacent oceanic plate. The latter is a vital stipulation because it frees the process from the possible requirement that subduction-related phenomena occur reasonably near the margin of the overriding plate.

The cause of the change in spreading directions is unclear; in our opinion, it is as likely that the orogeny caused the change in spreading directions as vice versa. We do not dispute the geographic association of the tectonic and plutonic features of Laramide orogeny with the generally convergent plate boundary to the west, but this observation falls short of being a solution to the gneiss dome and orogeny problem.

Before a genetic mechanism can be suggested that could account for the association of a convergent plate boundary with the Laramide orogeny, the question must be answered as to whether or not the cumulative displacement on the Shuksan and Rocky Mountain thrust systems represents actual crustal shortening.

Price and Mountjoy (1970, p. 18) hypothesized that the sole of the Rocky Mountain thrusts climbed the westward-dipping upper surface of the Precambrian crystalline basement in response to gravity loading in the vicinity of what is now the Omineca crystalline belt.

The gravity loading resulted from the upwelling of mobile material of the infrastructure, and in their view no crustal shortening is required. However, the Shuswap (Monashee Group), which probably represents both the upwelled material and its wrapping of supracrustal rocks and other supracrustal rocks metamorphosed to like degree in situ, spans only 55–105 km (35–65 mi). Furthermore, the high-grade metamorphic rocks of possible diapiric origin are not continuous elements of the hinterland along their north-south trend west of the Rocky Mountain thrust zone. If crustal shortening were not a factor, the dilatational features in the hinterland that would be required to account for the cumulative offset on the thrusts—in places approaching perhaps 250 km (150 mi)—do not appear to be present. Therefore, we agree with Bally, Gordy, and Stewart (1966, p. 360) that crustal shortening must be invoked to explain those offsets.

The alternative—that the thrust faults represent gravitational gliding from hypothetical highlands located west of the thrust belts within the hinterland—has been advocated by Mudge (1970). The arguments against this hypothesis have been recently summarized by Price (1971), with special reference to the disturbed belt of northwestern Montana, and by Armstrong (1972), with special reference to the Sevier orogenic belt, and are, in our view, compelling.

The required compressional mechanism apparently causes regional crustal contraction and concomitant mobilization and upward injection of infrastructure. Possibly the compressive stresses and high heat flow that caused the thrust faulting and diapirism of gneiss could result from opposing drag of the continental lithosphere by converging convection cells in the underlying asthenosphere. This hypothesis has been offered as an explanation of the bilateral symmetry exhibited by certain paired belts of post-Paleozoic thrust faults in part of the Cordillera by Burchfiel and Davis (1968). According to their interpretation, the magnitude of required crustal shortening in certain segments, such as that across the 40-km-wide (25-mi-wide) span between the east-dipping Shuksan thrust and the vertical or west-dipping Ross Lake fault zone (fig. 7), cannot be accommodated except by swallowing of crustal material at depth. They concluded that crustal shortening of large magnitude is required across the thrust complexes and suggest that convergence of subcrustal convection cells was the responsible orogenic mechanism.

Although we believe that the Ross Lake fault zone is a relatively minor feature and that the contractional axis must lie well to the east, below the Omineca crystalline belt, the convection cell hypothesis does provide a believable thrusting mechanism. In this model, the basement rocks at either side of the bilateral thrust belts—Shuksan and Cordilleran—are dragged toward and under the centrally located allochthon by a frictional couple with the converging convection cells in the mantle below. The crust would be expected to thicken along the axis and to either side of the cells through folding, stacking of thrust sheets, and plastic flow. This thickening is presumably counterbalanced by crustal shortening. The buoyancy of the crust would probably prevent its being swallowed by the mantle at the vortex of the cell.

The model is unsatisfactory as it stands, however, because no triggering force, reason for localization, or physical basis for the convection is provided. These defects can perhaps be overcome by considering the possible ramifications of the ideas articulated by Vine (1966, p. 1411), Palmer (1968, p. 341), and Kistler, Evernden, and Shaw (1971, p. 864–866), whereby, in the Mesozoic, the westward-drifting North American plate overrode an ancestral East Pacific rise locked between underthrusting oceanic plates.

Palmer suggested (1968, p. 344) that the Mesozoic plutonism and volcanism of the Cordillera was attributable to heat emanating from the underthrust rise and that the Cordilleran thrusts formed during "the terminal throes of convection before motion from the overridden crust was stifled***."

Kistler, Evernden, and Shaw (1971) suggested that the loci of magmatic activity in the western United States defined linear belts and that these belts were progressively displaced to the southeast from Jurassic to Cretaceous time (fig. 7). They hypothesized (1971, p. 864–866) that these magmatic belts originated as North America moved northwestward across a linear heat source analogous to an oceanic rise. But Rutland (1973, p. 829) rejected this hypothesis, noting that "the similarity of timing and spatial relationships of plutonism in North America, South America, and Japan surely demands a common explanation***." He concluded that the plate relationships postulated by Kistler, Evernden, and Shaw (1971) could not have ocurred in all these areas.

In our view, what all these areas have in common is an intermittently convergent continental plate–oceanic plate relation during the Mesozoic and Cenozoic. This relation could result in either subduction-zone related plutonism and volcanism of the Andean type (Hamilton, 1969a) or in plutonism and volcanism related to overriding of an oceanic rise. Both types would be related in space in that they occur inland from the convergent plate boundary; they might also appear regionally to be related in time, as periods of strong convergence would be most likely to end in overriding of a rise system.

The oceanic rise system has been considered by some (for example, Dietz and Holden, 1970, p. 4941) as a passive element in the plate tectonic model—simply a fracture along which magma upwells to fill the opening that would otherwise remain while the oceanic crust to either side is rafted or pulled away. For the rise system to exert the profound tectonic and magmatic influence here postulated, it must instead represent an active element—perhaps the crustal expression of a linear zone of high heat flow within the mantle that is not readily extinguished even though underthrust beneath a continental plate.

We speculate that in the study area overriding of the rise system marked the beginning of a transition from arc-type eugeosynclinal deposition and volcanism to an orogenic phase, with the area of magmatism—formerly restricted to the area of subduction-related volcanism immediately inland from the convergent plate boundary—now diffused along and to either side of the axis of the former rise system (fig. 8). Magmas generated during this phase were presumably the products of differential melting of mantle material, modified by assimilation of material from the sialic crust of the overlying continental plate.

Thus as the rise system is overriden production of basaltic magmas gives way to production of intermediate and felsic magmas; the latter are the products of varying degrees of fractional melting of mantle and assimilation of sial. Repeated extraction of this salic differentiate from the mantle and possibly the lower crust above the overridden rise system could leave an increasing accumulation of mafic residue. Possibly the build-up of this dense residue of infusibles could produce a mass that is gravitationally unstable relative to nearby areas within the mantle that lie outside of the zone of fractional melting. Ultimately, at some place along the axis of the overridden rise system, the weight of this more dense residue would overcome the plastic limit or strength of adjacent mantle, would sink, and the mantle to either side would converge catastrophically toward it. The overlying sial would be dragged toward the axis of the sinking mass but not swallowed because of its inherent buoyancy. Once started the cell could extend itself lengthwise along much of the overridden rise system, until it terminated against parts of the rise system so recently overridden that no significant instability existed.

In summary, we speculate that the convergence of oceanic and continental plates at a subduction zone would ultimately result in the overriding of an oceanic rise system by the continental plate, continuing a cycle whose succeeding phases would be dominated in turn by (1) eugeosynclinal sedimentation, (2) orogenesis with plutonism, (3) extreme compression with failure by overthrusting, possibly attended by mobilization and diapiric intrusion of the infrastructure, forming gneiss domes, and (4) the relaxation of compression, with uplift through isostatic rebound. This is basically an orogenic magmatic cycle similar to that advocated by DeSitter (1964, p. 397) and Rutland (1973, p. 828). The reality of the concept of a tectonic cycle has been questioned by Coney (1970). We hypothesize that this cycle is not only rationalized by but is indeed a necessary consequence of the plate tectonic model.

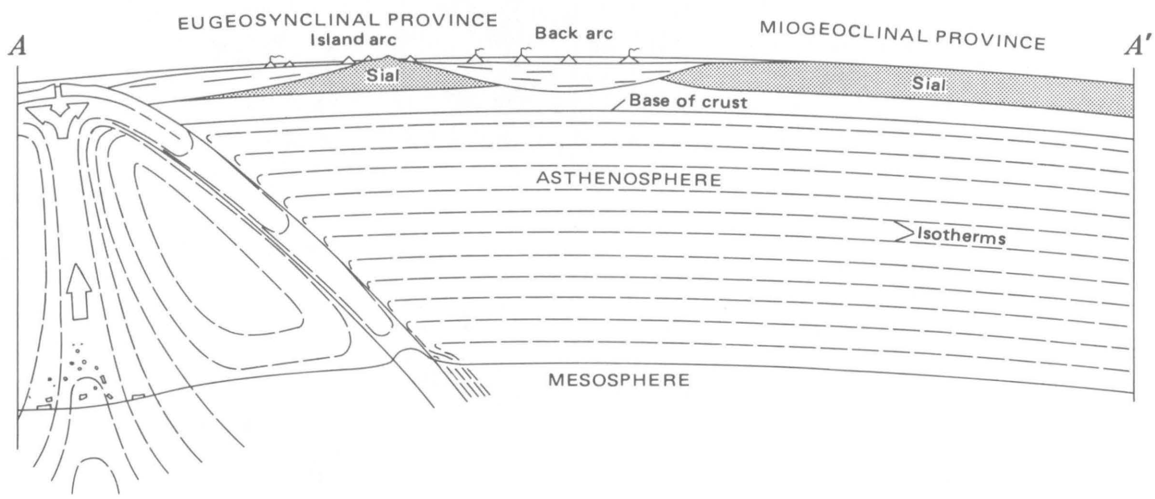
The Omineca crystalline belt may represent part of the relatively rigid center of the Cordilleran-Cascade allochthon, bridging the axis of sinking upper mantle. We speculate that the special features of the Omineca province—gneiss domes and the metamorphic rocks of the Monashee Group—originated through high heat flow and the forcing of hot plastic crustal material of the infrastructure upward along this axis during mid-Cretaceous to Eocene compression. Presumably, the postulated contraction of the continental plate along

FIGURE 8.—Stages in the development of the Omineca crystalline belt (sections drawn along line *A–A′* in fig. 7). *A*, In Permian and Early Triassic time, volcanic and pyroclastic deposits accumulated in island-arc and back-arc basin environments. To the west, an east-dipping subduction zone formed the boundary between oceanic and continental plates. A rise embedded within the oceanic plate was gradually nearing the continental plate. *B*, In Late Triassic and Jurassic time, the thickened eugeosynclinal prisms were invaded by calc-alkalic magmas derived through partial melting of the upper mantle and lower crust within a zone of high heat flow above the overridden rise. The residue (dotted area on figure) remaining in the zone of partial melting became progressively more dense as the hyperfusible part was removed.

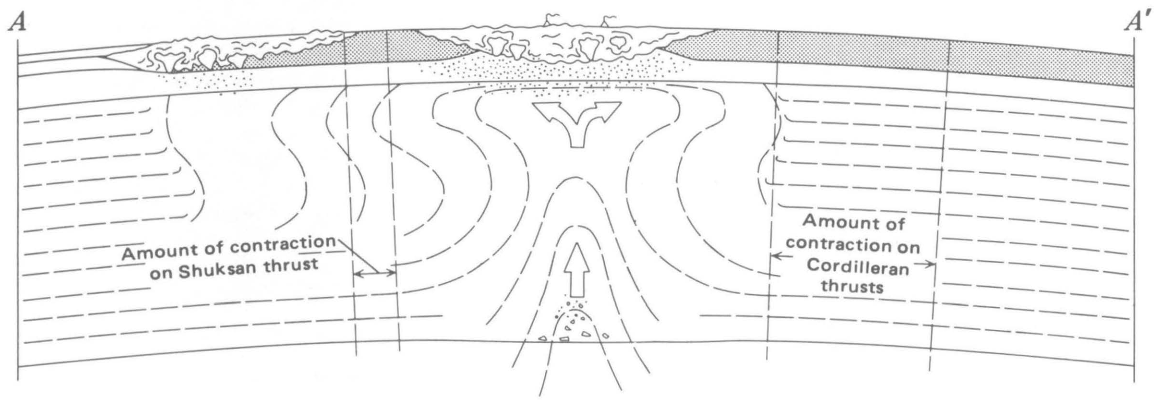
In Late Cretaceous time (*C*), after continued calc-alkalic plutonism through the Early Cretaceous, the dense residue occupying the source area of the magmas catastrophically sank into the asthenosphere, forming a short-lived convection that dragged overlying crust toward the axis of the cell. The crust thickened over the cell, through stacking of thrust sheets and plastic flow. Mobilized elements of the infrastructure penetrated higher levels in the crust, with their culminations forming gneiss domes.

With the demise of the convection cell in latest Cretaceous, the thickened crust isostatically rebounded. Upper crustal levels over the Omineca were rapidly eroded away, exposing elements of the Late Cretaceous infrastructure (the gneiss domes and the Monashee Group) in the Eocene.

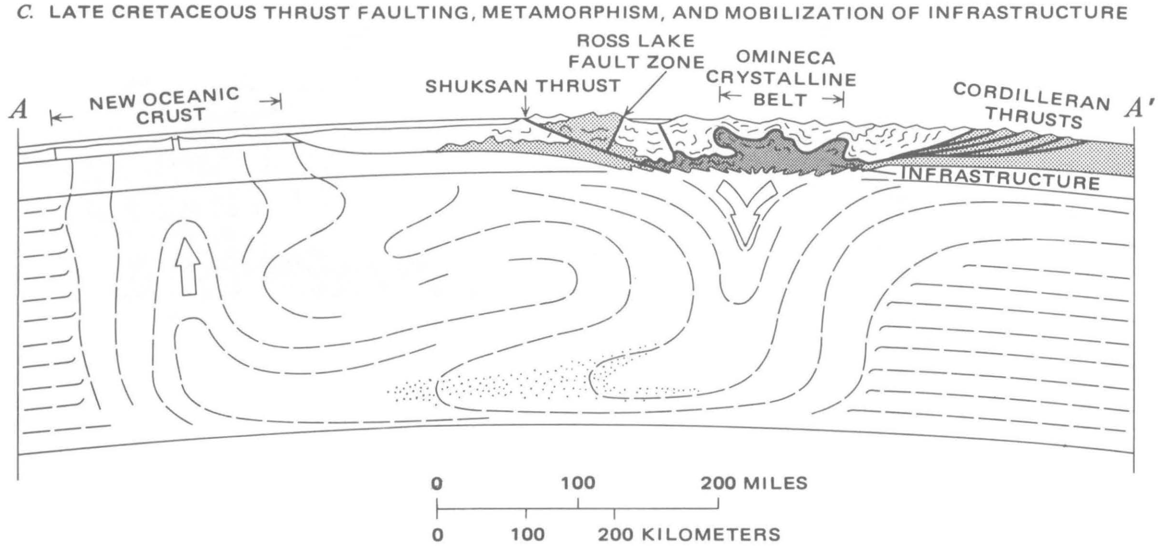
A. PERMIAN AND EARLY TRIASSIC VOLCANISM



B. LATE TRIASSIC AND JURASSIC PLUTONISM AND VOLCANISM



C. LATE CRETACEOUS THRUST FAULTING, METAMORPHISM, AND MOBILIZATION OF INFRASTRUCTURE



the Cordillera must have coincided with a reciprocal addition of new oceanic crust elsewhere, either through accelerated spreading rates at existing oceanic rises or through the formation of new rises.

The convulsive orogenic climax, represented in the study area by the emplacement of the Okanogan gneiss dome, was evidently followed in Late Cretaceous and early Tertiary by uplift and rapid erosion—and consequent cooling of the high-grade metamorphic rocks through the successive blocking temperatures of the dated minerals. This was followed in the Eocene by the postorogenic intrusion of hypabyssal plutons and the extrusion of related volcanic rocks. The Swimptkin Creek and Coyote Creek plutons are probably representatives of the postorogenic plutons. The Eocene deposits fringing the gneiss domes have been warped and locally faulted within the study area. To the east, they form a thick sequence of sedimentary deposits and interstratified volcanic rocks occupying the Republic graben. The north-northeast trend of the graben suggests that the former east-southeast to east-northeast direction of compression that had prevailed in the study area from the Late Triassic (Rinehart and Fox, 1972) became an axis of tension in the Eocene.

The final magmatic event in the region, the extrusion of nonorogenic late Miocene and early Pliocene plateau basalts, introduced a profound change in chemical composition of magma, from the more silicic and more highly differentiated Upper Triassic and post-Triassic igneous rocks.

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