Geophysical Interpretation of the Gneiss Terrane of Northern Washington and Southern British Columbia, and Its Implications for Uranium Exploration

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By JOHN W. CADY *and* KENNETH F. FOX, JR.

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Gravity highs suggest that gneiss domes of the Omineca crystalline belt are the surface expression of a zone of dense infrastructure

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By JOHN W. CADY and KENNETH F. FOX, JR.

ABSTRACT

The Omineca crystalline belt of northeastern Washington and southern British Columbia has a regional Bouguer gravity high, and individual gneiss domes within the terrane are marked by local gravity highs. Models of crustal structure that satisfy the limited available seismic-refraction data and explain the gravity high over the gneiss terrane permit the hypothesis that the core metamorphic complexes are the surface expression of a zone of dense infrastructure that makes up the upper 20 km (kilometers) of the crust within the crystalline belt.

The Omineca crystalline belt is characterized regionally by low aeromagnetic relief. The gneiss domes and biotite- and biotite-muscovite granites are generally marked by low magnetic relief, whereas hornblende-biotite granites often cause magnetic highs. Exceptional magnetic highs mark zones of magnetic rock within the biotite- and biotite-muscovite granites and the gneiss domes; these areas are worthy of study, both to determine the origin and disposition of the magnetite and to explore the possible existence of uraniferous magnetite deposits.

INTRODUCTION

This study began as an attempt to use gravity and aeromagnetic maps to help relate regional geology to the geology of the Midnite uranium mine in northeastern Washington. A preliminary gravity map of northeastern Washington (Cady and Meyer, 1976) showed that the Midnite mine, although it occurs in a local gravity low, lies at the south end of a belt of gravity highs which are in part associated with gneisses of the Omineca crystalline belt (fig. 1). In order to better evaluate this relationship, the study area was enlarged to include four 1°×2° quadrangles in British Columbia in addition to four quadrangles previously studied (Cady and Meyer, 1976) in northeastern Washington and northern Idaho. Models of crustal structure were made that cross the Omineca crystalline belt, the Columbia Plateau, and, peripherally, the northern Cascade Range in Washington and the Purcell fold belt in Idaho. In expanding the study area, we have neglected many of the details of geology and geophysics that might bear on uranium exploration. In northeastern Washington, proprietary aeromagnetic data are probably better than public data.

GEOLOGIC SETTING

The principal study area, outlined in the center of figure 1, overlaps the contact between two depositional provinces (Yates and others, 1966), the miogeoclinal province on the east and the eugeosynclinal province on the west (pl. 1). Plate 1 shows the major rock types of the principal study area, and is overprinted with gravity and magnetic maps. The miogeoclinal province is floored chiefly by Precambrian marine clastic deposits of cratonic derivation, including the Proterozoic Belt Supergroup (Purcell Group in Canada), the overlying Windemere Group, and the upper Belt equivalent, the Deer Trail Group. These rocks are overlain by lower Paleozoic shelf deposits, including both clastic and carbonate strata. Rocks of volcanic origin form a minor proportion of the westernmost elements of both the Precambrian and lower Paleozoic parts of this succession. The Windemere rocks form a narrow, westward-thickening miogeoclinal wedge trending northwest from northeastern Washington to southwestern Yukon Territory. This belt is believed to mark the western edge of the craton in late Precambrian time (Stewart, 1972; Gabrielse, 1972; Harrison and others, 1974, p. 2).

The eugeosynclinal province apparently consists of an aggregation of discrete tectonic terranes. Within the area encompassed by plate 1, these were collectively referred to as the "eastern assemblages" by Beck and others (1980). Individual tectonic terranes within this part of the province have not as yet been outlined in their entirety. Indeed, the existence of some is only suspected.

The province is floored at various places by lower Paleozoic, upper Paleozoic, or lower Mesozoic volcaniclastic and volcanic deposits. These include strata of Ordovician age assigned to the Covada Group by J. R. Snook (oral commun., 1980), the Permian Anarchist Group, the Pennsylvanian(?) Mount Roberts Formation, and the Permian or Triassic(?) Kobau Formation, and other formations (Fox and others, 1977, fig. 4). Except for the northwestern corner of the study area, these rocks are folded and faulted but, within a given tectonic

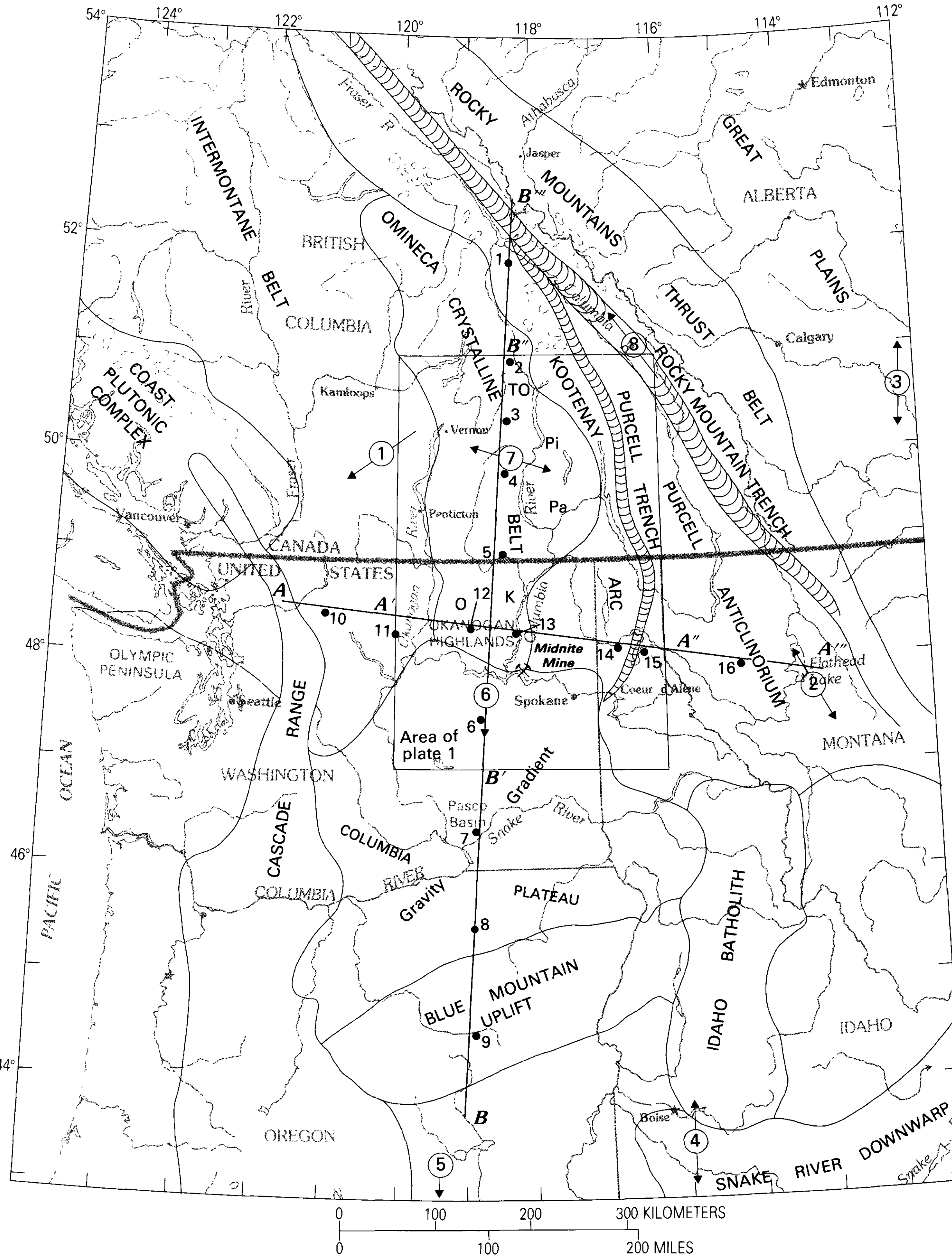


FIGURE 1.—Index map of the northwestern United States and southwestern Canada, showing the principal study area and surrounding region. *A–A'''* and *B–B'''*, locations of profiles along which models were prepared. Circled numbers indicate the location of seismic-refraction sections shown in figure 3, and the arrows indicate the direction of the refraction lines (two arrows show sections are reversed). Locations of gneiss domes: TO, Thor-Odin; Pi, Pinnacles; Pa, Passmore; O, Okanogan; K, Kettle. Numbered dots near the profile lines show sites of selected gravity values on table 1.

terrane, are stratally intact sequences that are in places unconformably overlain by rocks of Late Triassic and Jurassic age (Read and Okulitch, 1977; Little, 1978).

The Paleozoic rocks chiefly consist of volcanic wacke, siltstone, argillite, limestone, greenstone, bedded chert, chert breccia, and chert conglomerate. The Upper Triassic and Jurassic rocks are predominantly greenstone and weakly metamorphosed ribbon chert and pyroclastic rocks, though limestone, dolomite, and epiclastic rocks are locally abundant (Rinehart and Fox, 1976, p. 9).

Chronologically equivalent (in part) rocks immediately northwest of the study area, and perhaps extending into its northwestern corner in Canada, are part of the Cache Creek Group. These rocks were previously believed to be the northwestern counterpart of the Anarchist Group, Mount Roberts Formation, and Covada Group of the study area (Fox and others, 1977), but recent descriptions and interpretations indicate that the Cache Creek Group in its type area near Kamloops, British Columbia, is stratally disrupted melange (Monger, 1977, p. 1846; Travers, 1978), containing lithologic elements of Early Mississippian to Late Triassic age, and is locally metamorphosed within the blueschist facies (Monger, 1977, p. 1844). These features suggest that the Cache Creek rocks may mark the position of a formerly convergent plate boundary. Davis and others (1978, p. 3) placed the extension of that boundary in Washington near long 119°30′ W. along the north-trending axis of the Okanogan River valley, and they suggest that this was the western boundary of the North American craton in latest Paleozoic time. However, comparable geologic units of late Paleozoic and early Mesozoic age are present on both sides of this feature (Rinehart and Fox, 1972; Fox, 1970, 1978). If a suture is present in this region, it is probably at about long 120° W. and thus separates two very dissimilar tectonic terranes, the Methow graben (Barksdale, 1975) and the Okanogan highlands.

Rocks of both the eugeosynclinal and miogeoclinal provinces are metamorphosed to varying degrees, are cut by, or grade to, granitic rocks, and are patchily overlain by Tertiary (chiefly Eocene) volcanic, volcaniclastic, and epiclastic deposits.

In Washington, the contact between strata of the eugeosynclinal and miogeoclinal provinces appears to be tectonic (Yates and others, 1966, p. 54); depositional overlap has not been established, though the presence of volcanic rocks in the western part of the miogeoclinal succession was considered evidence of a transitional boundary between the two provinces by Yates (1970, p. 29; 1973). In British Columbia, the contact was drawn at the base of the eugeosynclinal Milford Group of late Paleozoic age by Fox and others (1977, fig. 3), and from there was extended southward along the Kootenay arc (Yates, 1973). Throughout much of its extent, this contact lies within highly deformed or highly metamorphosed strata. However, the Milford Group was reported by Cairnes (1934, p. 36–38) to overlie the lower Paleozoic Lardeau Group (here provisionally considered part of the miogeoclinal succession, though it contains subordinate members of volcanic origin) concordantly but unconformably. Davis and others (1978, p. 4) stated that farther north in British Columbia, the Milford Group appears to lie in normal stratigraphic succession on the Proterozoic to Devonian miogeocline. Hence, they included most of what is here referred to as the eugeosynclinal province with the North American craton as it existed in latest Paleozoic time.

In Washington, the easternmost part of the eugeosynclinal province includes rocks of the Ordovician Covada Group according to J. R. Snook (oral commun., 1980) and the Upper Permian and Lower Triassic rocks of Kelly Hill (Kuenzi, 1965), and the Pennsylvanian(?) Mount Roberts Formation. The age of the Covada Group immediately to the south is Ordovician (J. R. Snook, oral commun., 1980). On the basis of poorly preserved plant fossils, it had previously been very provisionally considered Pennsylvanian(?) by White (*in* Bancroft, 1914, p. 14). The Covada Group consists of marine turbidites, argillite, and intercalated greenstone. The coarser grained detrital rocks contain substantial admixtures of quartz, plagioclase, and potassium feldspar of probable granitic provenance (U.S. Geological Survey, 1974b). Though the source of this detritus is not known, the miogeocline immediately to the east seems unlikely. Granitic rocks that could have been exposed to erosion during the Paleozoic are unknown within this part of the miogeoclinal province. Hence, an allochthonous origin of the Covada tectonic terrane must be considered. If allochthonous, the suture must correspond to the boundary between miogeoclinal and eugeosynclinal provinces. The suture could be the northern extension of that between comparable eugeosynclinal and miogeoclinal rocks of western Idaho, as proposed by Hamilton (1976) and Jones and others (1977).

To summarize, the opinions of recent workers in the area seem to be converging toward the theory that the western side of the North American craton was truncated, and several, perhaps many, allochthonous blocks were sutured to it in latest Paleozoic or early to middle Mesozoic time. However, opinion is divided on whether the easternmost suture (1) corresponds to the eugeosynclinal-miogeoclinal provincial boundary, or (2) lies farther to the west within the eugeosynclinal province near long 120° W.

Our principal study area also embraces part of three orogenic belts superimposed on the depositional provinces discussed above. These belts include the Intermontane belt, Omineca crystalline belt, and Purcell anticlinorium (fig. 1). The continuation of these orogenic belts to the south of the study area is concealed beneath plateau basalt of the Columbia River Basalt Group of Miocene age.

The Omineca belt is distinguished from adjacent belts by the presence of extensive areas of high-grade gneiss and schist possessing a homogenous penetrative metamorphic fabric having two chief elements: (1) planar, moderately dipping to horizontal foliation, and (2) uniformly west-northwest-, west-, or east-northeast-trending lineation. These rocks include those of the Monashee Group of the Shuswap terrane of Jones (1959) and the Grand Forks Group of Preto (1965) in British Columbia, and the Tonasket Gneiss (Snook, 1965; Fox and others, 1976, p. 1219), and the metamorphic rocks of Tenas Mary Creek (Parker and Calkins, 1964, p. 5). These metamorphic rocks include orthogneiss, paragneiss, and migmatitic gneiss, and typically are sillimanite grade.

However, the Omineca, as defined above, is distinguished by structural fabric and not by metamorphic grade. This distinction is important, for kyanite- and sillimanite-grade metamorphic rocks without the distinctive structural fabric of the Monashee Group of the Shuswap terrane also are locally present within the Intermontane belt and the Purcell fold belt. Though these belts are of variable grade and age of metamorphism, they are typically only weakly metamorphosed. In the Loomis quadrangle, within the Intermontane belt, Rinehart and Fox (1972, p. 23–24) found that the Ellemeham Formation, a metamorphosed, predominantly volcanic and volcaniclastic formation of Triassic or Jurassic age, has in its upper member a sedimentary breccia containing clasts of previously metamorphosed rocks of older (Permian or younger) formations. Thus, at least two episodes of metamorphism, separated by an intervening period of uplift, planation, deposition, and burial, must have occurred.

The gneiss domes in the study area are part of a discontinuous belt of gneiss domes extending at least from central British Columbia southward to Mexico along the axial zone of the Cordillera (Fox and others, 1977, p. 18; Davis and Coney, 1979, p. 121). The origin of these gneiss domes is controversial, though it is now clear that not all the domes developed at the same time or in the same way. The Snake River Plain may mark a significant discontinuity separating chiefly upper Mesozoic domes to the north from middle to upper Tertiary domes to the south (Crittenden and others, 1978). From the many views expressed on the origin of the domes, two central themes emerge: (1) the gneiss domes originated through mobilization and diapiric flow of elements of the infrastructure and are analogous to the gneiss domes of the Caledonides of Greenland elegantly described by Haller (1971) (Price and Mountjoy, 1970; Fox and Rinehart, 1971; Campbell, 1973; Brown, 1978); (2) the gneiss domes are upwarped and deeply eroded bulges that expose a decollement of regional extent between suprastructure and infrastructure (abscherungzone of Armstrong and Hansen, 1966; Cheney, 1979).

The plutonic rocks within the study area are chiefly of Late Triassic to Eocene age (Gabrielse and Reesor, 1964; Fox and others, 1977; Miller and Engels, 1975), and range from quartz diorite to granite, and from monzonite to nepheline syenite in composition. In addition, small alpine-type ultramafic plutons are widespread throughout the eugeosynclinal province (McKechnie, 1965).

Most of the granitic rocks appear to fall within two groups, according to the identity of their micaceous, mafic, and accessory constituents. In one, the micaceous and mafic minerals are chiefly biotite and hornblende; in the other, they are biotite or biotite and muscovite. These groups correspond roughly to the I- and S-types, respectively, of Chappell and White (1974). According to Chappell and White, the I-type is derived through partial melting of igneous source material, whereas the S-type is derived through partial melting of sedimentary source material. Chemical differences between the two types are chiefly a reflection of the influence of the sedimentary cycle on the chemical composition of detritus.

In other words, the granitic rocks reflect, in their composition, the degree of sedimentary fractionation of their source. As stated by White and Chappel (1977), "the weathering process releases Na and Ca which are removed in solution and concentrated in sea water, evaporites and shelf limestones, and gives rise to clays which absorb K during sedimentation and diagenesis." This process results in peraluminous sedimentary rocks, in which the ratio of sodium to potassium is low; subsequent ultrametamorphism of these rocks produces S-type granitic rocks having these same chemical characteristics.

Within the study area, the Precambrian to Paleozoic sedimentary bedrock of the miogeoclinal province consists of thick sequences of argillite, siltite, quartzite, limestone, and the like, rocks which are the end product of advanced sedimentary fractionation. Hence, granitic rocks derived through partial melting of this material should be S-type.

In contrast, the rocks of the eugeosynclinal province consist of volcanic and volcaniclastic strata that show little sedimentary fractionation; hence, these rocks retain the chemical ratios of their primary igneous parent. Granitic rocks derived from partial melting of this material should be I-type.

Chappel and White's (1974) theory thus rationalizes the greater abundance of S-type rocks in the miogeoclinal province compared to the eugeosynclinal province. Those I-type granitic rocks that are present in the miogeoclinal province must have originated through partial melting of igneous parent material underlying the thick sedimentary wedge now exposed at the surface. Conversely, those S-type granitic rocks that are present in the eugeosynclinal province must have originated through partial melting of chemically differentiated sedimentary rocks (or their igneous and metamorphic derivatives) underlying the thick eugeosynclinal prism now exposed at the surface. This is an important deduction, for it implies that the eugeosynclinal rocks overlie, in places, at least, a sialic basement, and not former oceanic crust.

This conclusion is reinforced by Armstrong's (1979) study of Sr^{87}/Sr^{86} ratios in igneous rocks in British Columbia and northern Washington. Armstrong found that the zone of transition from ratios of 0.704 or less on the west to ratios of 0.706 or higher on the east follows the western boundary of the Omineca crystalline belt (as it is shown on fig. 1) south in southern British Columbia, then swings sharply westward to enclose the high-grade core of the Cascade Range, then eastward along the northern edge of the Columbia Plateau. According to Armstrong, high initial ratios occur across that part of Washington included by us in the Omineca crystalline belt. Igneous rocks having these higher ratios probably formed at least in part through melting of an old differentiated crust, implying that the eugeosynclinal province is at least in part underlain by a sialic basement of pre-late Paleozoic age.

In the principal study area, the biotite-hornblende rocks locally contain pyroxene as an additional constituent and are typically quartz diorite, granodiorite, or quartz monzonite. The biotite- or biotite-muscovite granitic rocks contain accessory garnet and allanite; some contain monazite. They are commonly gneissose, and include trondjhemite, granodiorite, and quartz monzonite. The biotite- and biotite-muscovite granitic rocks of northeastern Washington and adjacent parts of Idaho (Miller and Engels, 1975) are part of this group. Granitic gneiss included within the metamorphic core complexes (pl. 1) is, with local exceptions, compositionally similar to the biotite- and biotite-muscovite granitic rocks.

The leucocratic, aegirine-, riebeckite-, or aegirine-augite-bearing granites do not fit neatly into either group, but because they do contain amphibole or pyroxene or both, they were placed within the hornblende-biotite group of plate 1. The Kuskanax batholith in British Columbia (Read, 1973, p. 8) is an example.

The alkalic igneous rocks, including shonkinite, monzonite, pyroxenite, syenite, nepheline syenite, and trachyte form a diffuse east-trending belt across the eugeosynclinal province at the 49th parallel (Fox, 1973). These rocks show a range in age comparable to that of the other plutonic rocks, that is, Late Triassic to Eocene, and include volcanic as well as plutonic representatives. Their unusual composition may reflect some unusual compositional feature of the lower crust below this belt (Fox, 1977).

The volcanic, volcaniclastic, and epiclastic deposits of Tertiary age include two groups, the older and more extensive of which is Eocene, and the younger of which is mostly Miocene (Mathews and Rouse, 1963; Rouse and Mathews, 1961; Mathews, 1964; Hills and Baadsgaard, 1967; and Pearson and Obradovich, 1977).

At their base, the Eocene rocks typically consist of quartzo-feldspathic sediments eroded from nearby sources and deposited on a profound angular unconformity beveled across older metamorphic and plutonic rocks. The sediments are overlain by pyroclastic deposits, crystal lithic tuffs, that in turn are overlain by flows of quartz latite and rhyodacite; locally, alkalic rocks such as trachyte and trachyandesite form the base of the volcanic succession. The sediments and overlying flows are cut by hypabyssal intrusive equivalents of the flows.

In Canada, the eruption of these volcanic rocks was accompanied by block foundering, north-south block faulting, formation of large cauldron subsidence features, and emplacement of north-trending dike swarms (Souther, 1970). In Washington, particularly thick sequences of the Eocene volcanic rocks accumulated in north-northeast-trending volcanic-tectonic depressions such as the Republic graben (Muessig, 1967, p. 95–96) and the Toroda Creek graben (Pearson and Obradovich, 1977, pl. 1).

The volcanic rocks are part of a broad south-southeast-trending belt of Eocene extrusive rocks that extends into eastern Idaho and southwestern Montana. This zone has been referred to as the "Challis arc" (Vance, 1979), and its origin ascribed to melting processes at an Eocene subduction zone. According to this hypothesis, the subduction zone originated at a convergent plate boundary, which at that time was located to the southwest, near what is now central Oregon (Simpson and Cox, 1977). However, the apparent association of these volcanic rocks with tensional features (grabens), coupled with the alkalic character of some of the flows near the base of the sequence, indicate that the lavas originated through rifting and not through arc volcanism.

The Miocene rocks chiefly consist of the plateau basalts of the Columbia River Basalt Group. Judging from the distribution of erosional outliers of the basalt, flows of this unit inundated at least the southern edge of the map area (pl. 1). The basal contact is apparently a surface of low relief beveled across all Eocene and older rocks. This surface now slopes gently to the south.

In the northeastern corner of the study area is a structurally controlled physiographic feature of subcontinental scale, the Rocky Mountain trench. The trench is a linear furrow 3 to 16 km wide at its alluviated base, and extends southeastward from about lat 60° N. to about lat 47° N., a distance of about 1,600 km. At lat 52° N., the Rocky Mountain trench intersects the northern end of a less conspicuous, though well-defined south-trending furrow known as the Purcell trench. The Purcell trench extends southward to about lat 48° N.

The Rocky Mountain trench is apparently the physiographic expression of a linear zone of block-fault valleys and faults of diverse type (Leech, 1966). The faults are chiefly Paleocene or younger and include reverse, normal, and thrust faults, but despite the linearity of the trench, strike-slip offset of Tertiary age or younger has not been positively established (Leech, 1966). Leech (1964) suggested that the trench marks a deep, through-going structure such as a fracture zone, whose pattern is diffused and complicated in breaking through the near-surface rocks. The trench may also mark the western edge of the crystalline basement of the Precambrian Shield and the western edge of the craton, as it existed in Precambrian time (Leech, 1966). This boundary could form a rheomorphic discontinuity in the lower crust that has been printed up through younger cover rocks. This hypothesis finds some corroboration in the general coincidence of the trench and the strip of Windemere rocks that fringed the craton in late Precambrian time (Gabrielse, 1972). However, at the intersection of the Rocky Mountain trench, the Purcell trench, and the Kootenay arc, the belt of Windemere rocks diverges from near coincidence with the Rocky Mountain trench and extends along the Kootenay arc to the southwest.

CENOZOIC PLATE-TECTONIC SETTING

The present plate-tectonic geometry of the Pacific Northwest is now fairly well known through studies of Vine and Wilson (1965), Vine (1966), and Atwater (1970). The model now generally accepted is as follows: the North American, Pacific, and Juan de Fuca plates meet at a triple junction west of Vancouver Island (Atwater, 1970). The Queen Charlotte fault forms the boundary between the Pacific and North American plates, an ill-defined subduction zone forms the boundary between the Juan de Fuca and North American plates, and the Juan de Fuca ridge forms the boundary between the Pacific and Juan de Fuca plates. For at least the past several million years or so, the Pacific plate has been moving right laterally past the North American plate at about 5½ cm/yr (centimeters per year) (Minster and others, 1974), and the Juan de Fuca and Pacific plates have been moving apart as new oceanic crust is accreted to each at the Juan de Fuca ridge. Concurrent with this spreading, the Juan de Fuca plate has underthrust the North American plate, interposing a slab of oceanic crust of unknown thickness and extent at an unknown depth below part of the North American plate, and quite possibly, below all or part of the study area.

The northern edge of this subducted slab must extend generally eastward from the triple junction, and, hence, might form an important subcrustal discontinuity at depth somewhere in southern British Columbia or perhaps northern Washington. This possibility was recognized by Hill (1972, p. 1647) and stressed by Stacey (1973, p. 626), though not formally treated by either of these researchers in their gravity- and seismic-based crustal models. Stacey implied that the northern edge of this subducted slab might coincide with lat 51° N., based on the previously noted velocity transition of P_n waves from 7.8 km/s (kilometers per second) south and 8.1 km/s north of lat 51° N. suggested by White and others (1968, Model A), and the sharp increase in seismicity south of this transition.

GEOPHYSICAL SETTING

Geophysical studies of the Canadian cordillera prior to 1971 were reviewed by Berry and others (1971). Updated studies were published by Cumming and others (1979) and Monger and Price (1979). Drawing from these reviews and original sources, we briefly summarize the geophysical data that bear on the problem of explaining gravity highs over the Omineca crystalline belt.

GRAVITY STUDIES

Figure 2 is a regional simple Bouguer gravity map of the cordillera of southern Canada and northern United States. Among the more prominent features of the map are a low (minimum less than -200 mgal or milligal) over the Rocky Mountain trench, a high over the Columbia Plateau (maximum more than -50 mgal), a 140 mgal gradient zone between negative Bouguer anomalies inland and zero to -20 mgal Bouguer anomalies along the Pacific Ocean shore, and a low (minimum less than -230 mgal) over the Idaho batholith. An important lesser feature is an undulating gravity high (range of -120 to -130 mgal) over the Omineca crystalline belt and part of the Intermontane belt that lies

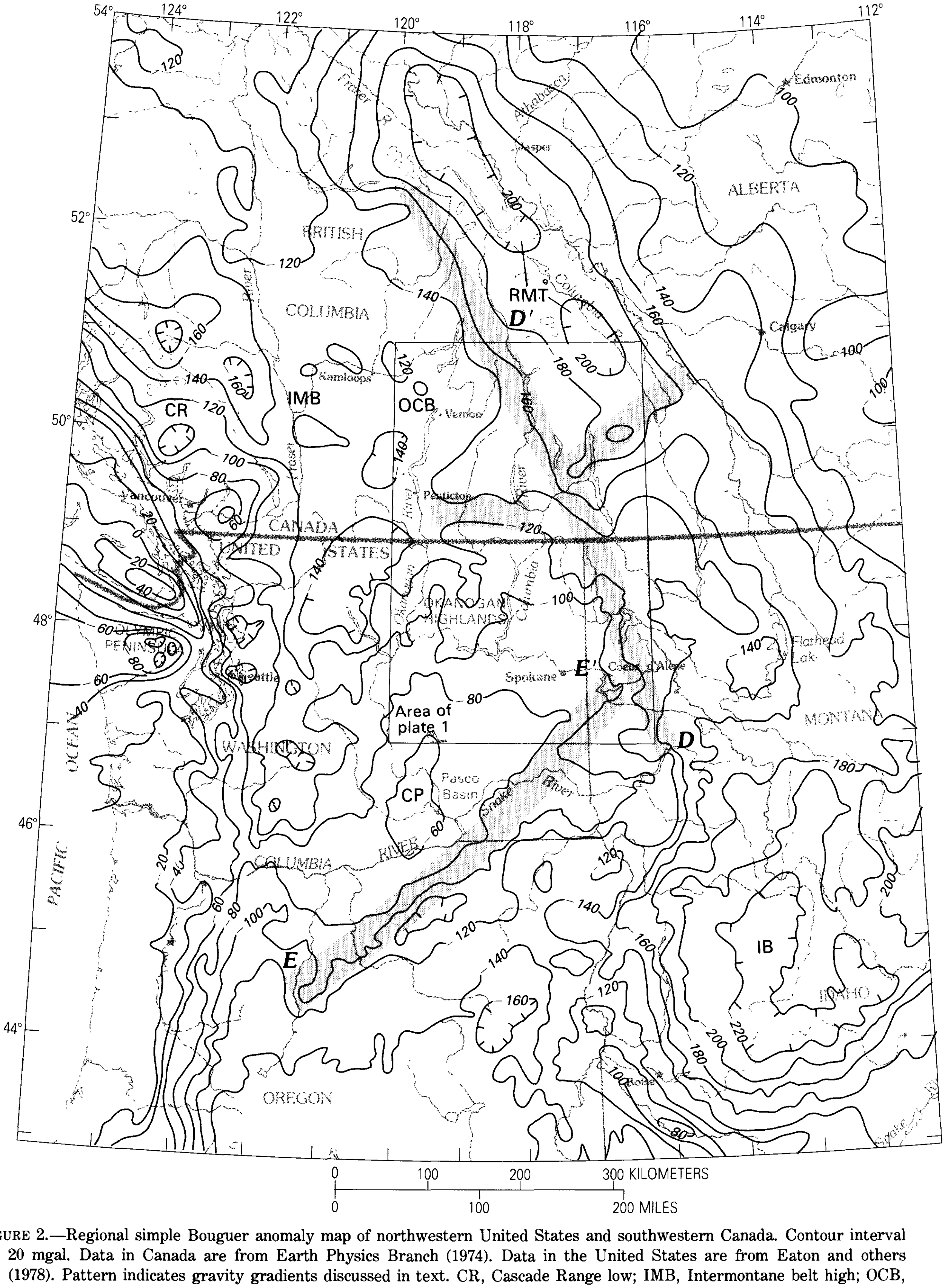


FIGURE 2.—Regional simple Bouguer anomaly map of northwestern United States and southwestern Canada. Contour interval 20 mgal. Data in Canada are from Earth Physics Branch (1974). Data in the United States are from Eaton and others (1978). Pattern indicates gravity gradients discussed in text. CR, Cascade Range low; IMB, Intermontane belt high; OCB, Omineca crystalline belt high; RMT, Rocky Mountain trench low; CP, Columbia Plateau high; IB, Idaho Batholith low.

between a low over the Cascade Range and the low over the Rocky Mountain trench.

Stacey (1973) used gravity data and seismic and isostatic constraints to prepare crustal structure models across the southern Canadian cordillera. He intentionally reduced the influence of anomalies related to near-surface geology by averaging the Bouguer values over areas of 30′ latitude by 1° longitude. Consequently, his preferred models, which allow for variations in average crustal density, average mantle density, and depth to the crust-mantle boundary, cannot account for the local anomalies discussed in this paper. Specifically, in his models A and B (Stacey, 1973, fig. 6), the gravity gradient west of the Rocky Mountain trench is explained by relief in the crust-mantle boundary, at a depth of 25 to 45 km. However, we show that this gravity gradient is too steep to be caused by a boundary deeper than 10 to 15 km.

Stacey (1973, p. 619–620) briefly discussed the relationship between gravity and surface geology, observing that Bouguer gravity values are 20 to 30 mgal more negative over granitic rocks than over intervening areas, and reporting average densities of plutonic rocks (2.62 g/cm^3), 0.03 and 0.05 g/cm^3 (grams per cubic centimeter) less than for volcanic and metamorphic rocks, respectively.

SEISMIC REFRACTION STUDIES

Berry and Forsyth (1975) synthesized refraction studies in the Canadian cordillera and concluded that there is a difference in the velocity of the upper mantle between the north and the south. The boundary appears to lie between lat 51° and 52° N.; P-wave velocity in the upper mantle to the north is 8.0 km/s, whereas that to the south is 7.83 km/s (Berry and Forsyth, 1975, p. 196–197). They suggested (their figs. 13b and 14) that the crust in the southern cordillera has an average velocity of 6.4 km/s, and that it gradually thickens from 25 km below the coast plutonic complex to 37 km under the Omineca crystalline belt. In the absence of an adequate number of stations in the Omineca crystalline belt, however, the interpretation there is based largely on extrapolation.

Cumming and others (1979) established a partially reversed refraction line that crossed the Omineca crystalline belt about lat 50° N. They determined that the dominant crustal feature along their profile is an undulating M-discontinuity that dips gently to the east from a depth of about 30 km beneath the Intermontane belt to about 40 km beneath the Purcell fold belt. A 5-km-high bump in the undulating M-discontinuity underlies the Omineca crystalline belt. They found that the upper crust in the Omineca crystalline belt has a velocity of 6.2 km/s, 0.15 km/s faster than in the Intermontane belt. In contrast to previous studies in the Canadian cordillera, they found it necessary to include in their model a lower crustal layer 12 km thick with a velocity of 6.9 km/s. These results are important, for they help to explain the gravity high observed over the Omineca belt.

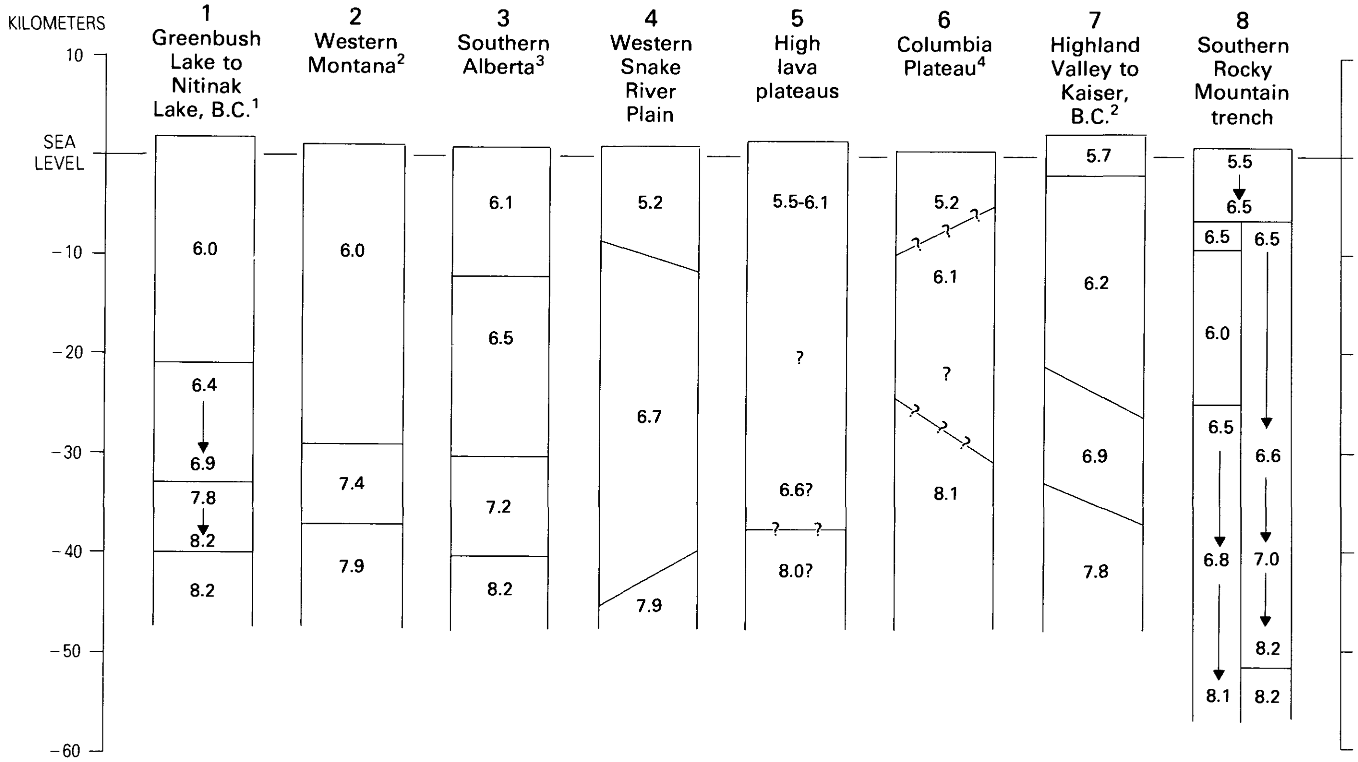
From an unreversed seismic-refraction profile along the Rocky Mountain trench, Bennett and others (1975) and Spence and others (1977) determined that the crust between lat 50° N. and lat 53° N. is between 50 and 60 km thick. Refractors having apparent P-wave velocities of 6.5 to 6.6 km/s and 8.22 km/s were interpreted to be the surface of the Precambrian basement and the M-discontinuity, respectively.

Hill's (1972, 1978) unreversed profile, roughly coincident with the 118°30′ meridian, extends southward across the Omineca crystalline belt and Columbia Plateau from a shotpoint near lat 51° N. In his original model, Hill (1972) assumed that the upper mantle had a velocity of 7.9 km/s. Time-term analysis of quarry blasts in the Pasco basin (at the confluence of the Snake and Columbia Rivers) (J. P. Eaton, oral commun. 1976) showed the upper mantle there to have a velocity of 8.0 to 8.2 km/s, so, in his revised model, Hill (1978) used a velocity of 8.1 km/s in the Pasco basin, and projected this velocity further north. Because the later reversed profile of Cumming and others (1979) showed upper mantle velocities of 7.8 to 8.0 km/s near the northern end of Hill's profile, the lower velocities may be more accurate. Hill (1972, 1978) inferred crustal layers of 6.0 and 6.6 km/s velocity in the Omineca crystalline belt, less than the 6.2 and 6.9 km/s velocity observed by Cumming and others (1979). The difference may arise in part from a southward dip of refractors along Hill's unreversed profile. Another important factor is that Hill's profile is along the Republic graben, a structure that probably contains lower velocity material than most of the Omineca crystalline belt (see the section on "Major features of the gravity map").

Crustal sections from these and other refraction studies are presented in figure 3. Boundaries for these sections are plotted as bar symbols in the gravity models of figures 5 and 6.

OTHER GEOPHYSICAL STUDIES

Several lines of evidence indicate that the Canadian cordillera is geophysically anomalous compared to the Canadian Shield to the east. In surface-wave studies, Wickins (1971, 1977) observed low upper mantle velocities in the entire southern Canadian cordillera, and locally found evidence for a low velocity zone in the lower crust. Heat-flow measurements in Canada (Jessop and Judge, 1971; Judge, 1977) confirmed Blackwell's (1969) suggestion that the "Cordilleran



[1]Simplified from a continuous velocity-depth curve.
[2]Reversed.
[3]Deep-reflection seismology.
[4]Profile for section 6 from Omineca crystalline belt into Columbia Plateau. Section shown is for Columbia Plateau. Northern part of profile (Hill, 1978, fig. 7) was influenced by the 6.0–km/s upper crustal velocity found in section 1 to the north. A higher upper crustal velocity (6.2 km/s) is shown in section 7 to the north. Hence, the velocities labeled 6* in figures 5–7, taken from Hill (1978, fig. 7) should not be considered as definitive values for crustal velocity in the southern Omineca crystalline belt. For further discussion, see section entitled "Intermediate density crustal layers."

FIGURE 3.—Vertical crustal sections showing compressional-wave velocity in km/s, determined for locations shown on figure 1. Arrows show increasing velocity with depth. Sloping lines show sloping seismic layering. Seismic boundaries and velocities queried where uncertain. Sources of data: 1, Berry and Forsyth, (1975); 2, Steinhart and Meyer (1961, p. 341); 3, Kanasewich and others (1969); 4–6, Hill (1978); 7, Cumming and others (1979); 8, Bennett and others (1975) on left and Spence and others (1977) on right.

Thermal Anomaly Zone" continues into Canada west of the Rocky Mountain trench.

Caner (1970) observed that long wavelength aeromagnetic anomalies are attenuated west of the Rocky Mountain trench, and he suggested that the cause might be "upwelling of the Curie point isotherm, breakup of large structures by tectonic and metamorphic mechanisms, and/or a more silicic crust." Haines and others (1971) found that there is an aeromagnetic quiet zone, lacking significant anomalies of any wavelength, associated with the Omineca crystalline belt. A shallow Curie-point isotherm or breakup of large structures cannot explain the absence of short-wavelength anomalies. A highly silicic crust would be unlikely to cause the gravity highs associated with the Omineca crystalline belt. The rocks of the Omineca crystalline belt may be, on the average, deficient in magnetite, possibly as a result of their metamorphic history.

Geomagnetic depth sounding and magnetotelluric studies in the Canadian cordillera (Caner and others, 1971; Dragert and Clarke, 1977) showed strong attenuation of the Z-component of magnetic variations in the frequency range 10^{-4} to 10^{-3} Hz (hertz). Models of conductivity structure that satisfy these measurements require a low resistivity (5 ohm-meter) layer having a well-defined top at a depth of 10 to 15 km, and a poorly determined bottom at a depth of 25 to 55 km. The eastern boundary of the conductive layer is well defined. It coincides with the Rocky Mountain trench north of lat 51°N., but swings west and follows the Kootenay arc further south (fig. 1).

Caner (1970) argued that the high heat flow, the aeromagnetic quiet zone, the moderately conductive upper mantle, and the low seismic velocities in the upper mantle can be explained by a "hot upper mantle and/or lower crust." High temperature alone, however, cannot explain the conductive zone 10 to 15 km deep in the crust, so Caner (1970) called upon the hydrated lower crust theory proposed by Hyndman and

Hyndman (1968). As explained below, the inferred hydrated zone coincides with a zone of intermediate density ($\rho=2.84$ to 2.86 g/cm^3) in our gravity model.

EAST-WEST GEOPHYSICAL FEATURES

Although the first-order geophysical features follow the approximate north-south trends of the cordillera, several second-order east-west features may be significant. Kanasewich and others (1969) found seismic evidence for an 11-km-deep Precambrian rift in Alberta between lat 50° N. and lat 51° N. Using sparse gravity and aeromagnetic data, they extrapolated the rift southwestward across the Rocky Mountains and into the Purcell anticlinorium between lat 49° N. and lat 50° N.

A zone of east-west gravity gradients (fig. 2), decreasing to the north, occurs just north of lat 49° N., near long 118° W. Another zone of gradients runs to the northeast from near lat 50° N., long 117° W. These gradient zones lie near the westward extrapolation of the Precambrian rift mentioned above.

From seismic-refraction studies, Berry and Forsyth (1975, p. 196–197) concluded that there is a difference in upper mantle velocity between the northern and southern parts of the cordillera. The boundary appears to be between lat 51° and 52° N.; mantle velocity to the north is 8.04 km/s, whereas that to the south is 7.83 km/s. This boundary lies north of our principal study area.

An east-west conductivity boundary cuts across the main north-south conductivity structure near lat 49°15′ N., long 117° W. (Lajoie and Caner, 1970; Caner and others, 1971). They hypothesized that the boundary was caused by left-lateral offset of the main north-south conductor by a fault associated with the Alberta rift system reported by Kanasewich and others (1969). Dragert and Clarke (1977) suggested that the east-west conductivity anomaly is caused, not by strike-slip offset of the north-south conductor, but by a conductor associated with, or terminated by, the hypothetical Alberta rift system.

Cady (1980a) presented a preliminary version of the gravity models described below and explained the gneiss domes as the surface expression of a zone of deep infrastructure. He suggested that the dominant mechanism for the emplacement of gneiss domes was isostatic uplift and erosion of crust thickened by regional compression. From the coincidence of dense rocks in the gravity model with an electrically conductive layer detected by geomagnetic depth sounding, he inferred that the same rock type, possibly amphibolite, was responsible for both the gravity and electrical anomalies.

GRAVITY STUDIES

SOURCES OF GRAVITY DATA

Bouguer gravity data (pl. 1*A*) in the United States are from Cady and Meyer (1976), and in Canada are from Stacey and others (1973). The Canadian data, as originally published, were referenced to the old Potsdam gravity datum. To make them compatible with the United States data, the Canadian data were converted to the new Potsdam datum and the new Geodetic Reference System of 1967 using Formula B of the International Association of Geodesy (1971, p. 74). For plate 1*A*, an approximate correction of -5 mgal, accurate to ±0.4 mgal, was made. North of lat 49° N., the original contours, relabeled -115, -125, *** -205, are from the Canadian map (Stacey and others, 1973). The contours labeled -120, -130, *** -200 were interpolated and hence are less precise.

Gravity data used in the gravity profiles outside the principal study area shown in plate 1*A* are from Eaton and others (1978) in the United States and from Stacey and others (1973) in Canada. The profiles are not as precise outside the principal study area, as they are taken from small-scale maps.

MAJOR FEATURES OF THE GRAVITY MAP (PLATE 1*A*)

The gravity map (pl. 1*A*) shows the northern flank of a broad positive Bouguer gravity anomaly centered in the interior of the Columbia Plateau province near lat 46°30′ N. and long 119°30′ W. (Bonini and others, 1974). The rising southerly gradient of Bouguer gravity that dominates the southern and southwestern parts of the map area coincides with decreasing topographic elevations towards the center of the Columbia Plateau. Seismic data and gravity modeling (discussed below) show that the high is caused primarily by shallowing of the mantle under the plateau. Decreasing Bouguer anomaly values toward the northeastern corner of the map area indicate thickened crust beneath high mountains.

Superimposed upon the regional trend are smaller features, often marked by steep gradients. Bouguer anomaly highs or noses labeled A, B, C, D, E, and F occur at the Thor-Odin, Pinnacles, Valhalla, Passmore, Okanogan, and Kettle gneiss domes, respectively. These gravity highs do not correlate with topographic lows, as would be the case if the crust were of constant density and the Bouguer anomaly relief resulted only from isostatic compensation of topography. Particularly notable are the Valhalla dome, where the closed gravity contour occurs over elevated terrain, and the Thor-Odin dome, where the gravity values increase in the direction of increasing topographic elevation.

A more quantitative estimate of the relationship between elevation and Bouguer anomalies is given in table 1, for selected gravity stations shown on figure 1. A least-squares analysis using the 16 stations yielded a line having the equation

$$\text{Bouguer anomaly} = -46.83 - 0.0769 \times \text{elevation in meters.}$$

Twenty kilometers was chosen as the radius of the circle over which elevations were averaged because it yielded a better inverse correlation of Bouguer anomaly with elevation than either 15-, 30-, or 40-km radii. Presumably, 20 km represents the lateral distance over which most isostatic compensation occurs.

If the straight line having slope 0.0769 mgal/m (milligals per meter) is taken as representing the normal relationship between Bouguer anomaly and elevation for the region, positive residuals from the line indicate local mass excesses, and negative residuals indicate local mass deficits. The Omineca crystalline belt is characterized by positive residuals. An abrupt transition from +15 mgal west of the Purcell trench to -14 mgal east of the Purcell trench (sites 14 and 15 in fig. 1 and table 1) probably marks the eastern boundary of the Omineca crystalline belt in the subsurface.

The Bouguer gravity positive anomaly over the interior of the Okanogan gneiss dome probably is caused by the density contrast between its core and the partially enveloping granitic gneiss and structureless granitic rock. The dry-bulk density of 55 specimens of the Tonasket Gneiss, forming the west-central part of the Okanogan gneiss dome, ranges from 2.61 to 3.05 g/cm^3, and averages 2.75 g/cm^3. The rocks sampled include layered quartz dioritic gneiss, amphibolite, and interlayered and cross-cutting leucosomes. The densities of an additional 80 samples from areas of the Okanogan gneiss dome north and east of the west-central part range from 2.59 g/cm^3 to 2.98 g/cm^3, and average 2.67 g/cm^3. The densities of 123 samples of the lineated granitic gneiss and structureless granitic rock bordering the Tonasket Gneiss within the Okanogan gneiss dome range from 2.60 g/cm^3 to 2.70 g/cm^3, and average 2.65 g/cm^3.

TABLE 1.—*Relationship between average elevation and Bouguer anomalies for sites in the study area in northern Washington and southern British Columbia*

Site (fig. 1)	Bouguer anomaly (mgal)	Elevation[1] (m)	Residual[2] (mgal)
1	-195	1430	-39
2	-168	1348	-18
3	-135	1301	+11
4	-155	1442	+ 2
5	-112	824	- 3
6	- 85	530	+ 2
7	- 72	290	- 4
8	-113	674	-15
9	-143	1481	+17
10	-163	1534	+ 1
11	-110	1129	+23
12	-120	1091	+10
13	- 94	708	+ 6
14	- 94	827	+15
15	-130	910	-14
16	-140	1318	+ 7

[1]Averaged in a 20-km-radius circle centered on the site.
[2]Between Bouguer anomaly at each site and line resulting from least squares-analysis.

A north-south-trending gravity high, labeled F, is associated with a high-grade metamorphic terrane referred to as the Kettle dome (Cheney, 1976), but this anomaly culminates to the south over low-metamorphic-grade eugeosynclinal rocks. The low-grade rocks, which are assigned to part of the Ordovician Covada Group by J. R. Snook (oral commun., 1980) were mapped and described by Becraft (1966). In view of their volcanic provenance, the rocks of the Covada Group may have an average density significantly greater than 2.67 g/cm^3. The density of 15 samples of the Covada Group taken from widely scattered outcrops along the Columbia River ranges from 2.66 to 2.86 g/cm^3, and averages 2.73 g/cm^3. The positive anomaly over them in this area may reflect this above-average density, or it may be caused by the subsurface extension of a high-density phase of the nearby core complex.

Between the Okanogan and the Kettle domes lies the Republic graben, which is marked by a gravity low wider than the graben itself. The graben appears to be an Eocene depression filled and locally rimmed with coeval volcanics (Sanpoil Volcanics and Klondike Mountain Formation) and porous, weakly cemented volcaniclastic rocks (O'Brien Creek Formation). The negative anomaly there is in part due to the low average density of the Eocene rocks.

Based on a detailed gravity survey of this area, Soule (1975) concluded that the gravity low was caused by the low average density of two superimposed masses: (1) a lens-shaped body of silica-rich igneous rock wider than the graben itself; and (2) the volcanic and volcaniclastic fill of the graben. The gravity anomalies indicate vertical displacement of from 2 to 3.5 km on the bounding faults, with density contrasts of 0.08 to 0.12 g/cm^3 between the volcanic rocks within the graben and the country rock.

Other local gravity highs may indicate doming of high-density rocks in both exposed and covered parts of the layered gneiss terrane. Of these, highs G and H occur over layered gneiss. Highs I and J occur over eugeosynclinal deposits in close proximity to layered gneiss. High J may indicate a buried gneiss dome along the trend of the Valhalla and Passmore domes.

Other gravity highs occur within the miogeoclinal belt. Highs K and L roughly coincide with a terrane of high-grade gneiss and schist north-northeast of Spokane, Wash., and are thought to be either metamorphosed parts of the Belt Supergroup or possibly pre-Belt rocks (Anderson, 1940, p. 20; Weissenborn and Weis, 1976). These high-grade metamorphic rocks resemble rocks of the metamorphic core complexes in their gravity expression, and, judging from descriptions by the authors cited above, in lithology, structure, and metamorphic grade as well.

Finally, an east-northeast-trending high, M, at lat 48° N., long 118° W., crosses the boundary separating the eugeosynclinal and miogeoclinal zones. It is accompanied by a parallel gravity low, N, within which the Midnite mine is located. Possibly, the high marks a buried gneiss dome and the low marks a partially buried granitic pluton.

One of the conspicuous features of the gravity map is the zone of steep gradients, labeled *D–D′* that trends north-northwest from the southeast corner of the map area. In this area, the zone lies near, but does not coincide with, the southern termination of the Purcell trench. Further north, the zone bifurcates. The more pronounced limb flanks the core metamorphic complexes at Thor-Odin, Pinnacles, Valhalla, and Passmore domes, thus coinciding with the eastern boundary of the Omineca crystalline belt. The other, more subdued limb (defined by the -180 mgal contour), coincides approximately with the Purcell trench in Canada and the northern part of the boundary between eugeosynclinal and miogeoclinal provinces.

The southern part of this boundary along the Columbia River coincides with a faint alinement of weak gravity gradients as far south as about lat 48°15′ N. The overall pattern is probably X-shaped, the left diagonal weakening and merging with the background along its strike to the southwest.

In the next section we show that the source of the gravity-gradient zone *D–D′* must reside in the upper crust at a maximum depth of 12 km. The association of the high-gradient zone with the eastern boundary of the core metamorphic complexes in Canada, and with the probable correlatives of the core metamorphic complexes in Idaho, suggests that the zone marks the eastern side of a belt of relatively dense infrastructure, of which the core metamorphic complexes are the surface expression. The presence of the gravity high in the Kootenay arc in Washington and Idaho suggests that high-grade metamorphic rocks of the infrastructure are present between the mapped eastern boundary of the Omineca crystalline belt and the Purcell trench.

The very weak zone of gravity gradients along the boundary between eugeosynclinal and migeoclinal provinces probably indicates small differences in density and thickness between the country rocks of these respective terranes. The weakness of this zone implies that there is no substantial change in density of basement rocks or lower crust, coincident with the boundary between the eugeosynclinal and miogeoclinal provinces.

The northwest-trending zone of gravity gradients is broken in two places by conspicuous southwest-trending gravity lows, O and P. Low O occurs over an extensive area of granitic rocks and some high topography. The low is probably caused by deep-seated, low-density plutonic rocks. Low P also occurs over granitic rocks and suggests, especially near its northeastern end, a deep-seated pluton.

The east-trending gravity high, Q, bounded to the north by a prominent gradient zone, was mentioned previously as one of several, possibly related, east-trending geophysical anomalies. If the source of these anomalies were in the upper mantle, a geologic structure that might explain them is the northern edge of the subducted part of the Juan de Fuca plate. However, this hypothesis conflicts with Riddihough's (1978) model of the subducted slab below western Washington, a model that satisfies the constraints of both gravity and seismic-velocity data. Riddihough assumed that the slab slopes shallowly eastward below the Olympic Peninsula and then bends downward beneath Puget Sound, and that the top of the slab lies at depths about 100 km beneath the Cascade Range. If the slab does not bend upward east of the Cascade Range, it would lie at depths in excess of 200 km below the study area, much too deep to generate the sharply defined east-trending gravity anomaly, Q. The north flank of anomaly Q has a steep gradient zone (10 to 20 mgals in a width of 6 to 12 km) that indicates that the density contrasts responsible for the anomaly are present in the upper crust.

High Q is puzzling because it cuts across the geologic grain of the region, extending across the Omineca crystalline belt and into the Purcell fold belt. The only lithologic unit exposed at the surface that correlates with the gravity high is the eastern part of a diffuse east-trending belt of Upper Triassic to Eocene alkalic igneous rocks described by Fox (1973). Fox (1977) speculated that the alkalic rocks might mark the location of a subcrustal structural or compositional feature. Conceivably, this feature could cause the observed anomaly.

The prominent northwest-trending high, R, in the northwest corner of the map area coincides with a belt of argillite, quartzite, andesitic lava, tuff, and limestone correlated with the Cache Creek Group by Jones (1959). The belt is flanked on the north by granitic rocks, over which the gravity low, S, is centered. On the south, the belt is flanked chiefly by Eocene volcanic rocks overlying granitic rocks, which coincide with a broad gravity low. Hence, the anomalies R and S are probably caused by density contrasts within the exposed bedrock.

The Bouguer gravity map of the United States (Woollard and Joesting, 1964) shows local gravity highs over other core metamorphic complexes identfied by Fox and others (1977)—parts of the Raft River area in Utah and Idaho, the Snake Range in Nevada, and the Rincon Mountains in Arizona. The same highs have been eliminated in the more recent Bouguer anomaly map of Eaton and others (1978), possibly due to the smoothing effect of machine contouring. It would be interesting if gravity anomalies could be used to trace a high-density infrastructure across the country, but this would be very difficult in the western cordillera of the United States, where the gravity field has been so strongly affected by Cenozoic tectonics.

DETAILED GRAVITY MODEL ACROSS PURCELL TRENCH

In the previous section, we identified gravity-gradient zone *D–D′* on plate 1*A* as the eastern boundary of high-density infrastructure that penetrates the Earth's surface as gneiss domes. Here we defend the interpretation that the gravity gradient is caused by density variations in the upper crust, not in the lower crust or upper mantle.

Closely spaced (less than 2 km) gravity data are available across the Purcell trench along the central part of line *C–C′* in plate 1 (D. L. Peterson, unpub. data, reported by Cady and Meyer, 1976). Gravity gradients were calculated from the simple models shown on figure 4 to determine the maximum possible depth to the source of gradient *D–D′*. Because the gravity gradient is continuous over a long distance perpendicular to the profile, the main features of the gravity gradient were calculated using a two-dimensional model having infinite extent, or "strike length," perpendicular to the profile. Density contrasts were determined by least-squares analysis to yield the best possible fit to the observed profile for the geometry used in each model.

Profile A was calculated using a 5-km-thick, two-dimensional slab (model A) having its top at a depth of 20 km and a density contrast of 0.24 g/cm^3. The gradient of profile A is much gentler than that of the observed gravity field. The slab is too deep.

Profile B was calculated for a 5-km-thick, two-dimensional slab (model B) having its top at a depth of 10 km and a density contrast of 0.23 g/cm^3. The fit of the profile to the central part of the observed gradient zone is excellent. In other tests, it was found that 12 km was the greatest depth at which a slab of any density contrast could reproduce the observed gradient. Calculations using this two-dimensional slab failed to reproduce the peak west of the trench and the trough east of the trench. These features can be seen to be three-dimensional, having a limited strike length, on the gravity map (pl. 1*A*) and cannot be reproduced by a two-dimensional slab.

A "2½-dimensional" gravity model (Cady, 1980b) (model C, fig. 4) was constructed using realistic source-body strike lengths determined from the strike length of anomalies on the map. In model C, the area west of the Purcell trench is underlain by a nearly two-dimensional (150-km strike length) slab about 5 km deep and 10 km thick. A 2½-dimensional (strike-length 40 km) bulge on the slab underlies the local gravity high west of the trench. Both the slab and the bulge have a positive density contrast of 0.10 g/cm^3. The local gravity low east of the trench is caused by a buried pluton, shown in the model by a 2½-dimensional body having a density contrast of -0.09 g/cm^3 and a strike length of 50 km.

The models shown in figure 4 show that the source of gravity gradient *D–D′* must lie in the upper crust. The models do not, however, take into account seismic constraints or show the expected variation of density with depth in the crust, as do the regional models described in the following section.

REGIONAL CRUSTAL MODELS

Gravity highs occur over high-grade metamorphic rocks of the Omineca crystalline belt. These highs do not simply correlate with topographic lows, so they must result, at least in part, from density variations in the crust, and not simply from the isostatic compensation of topography. We have prepared three two-dimensional crustal models that can account for the observed gravity highs. The models draw upon the results of the 2½-dimensional modeling described above, but as their purpose is to elucidate the regional structure, no attempt was made to match every detail of the observed gravity profile.

The models are not unique because an infinite set of hypothetical models can account for any given gravity anomaly. The models are partially constrained, however, by the published seismic-refraction profiles shown on figure 3. The locations of these refraction profiles are plotted in figures 1, 5, and 6. In the models, we assume that velocity and density are related, according

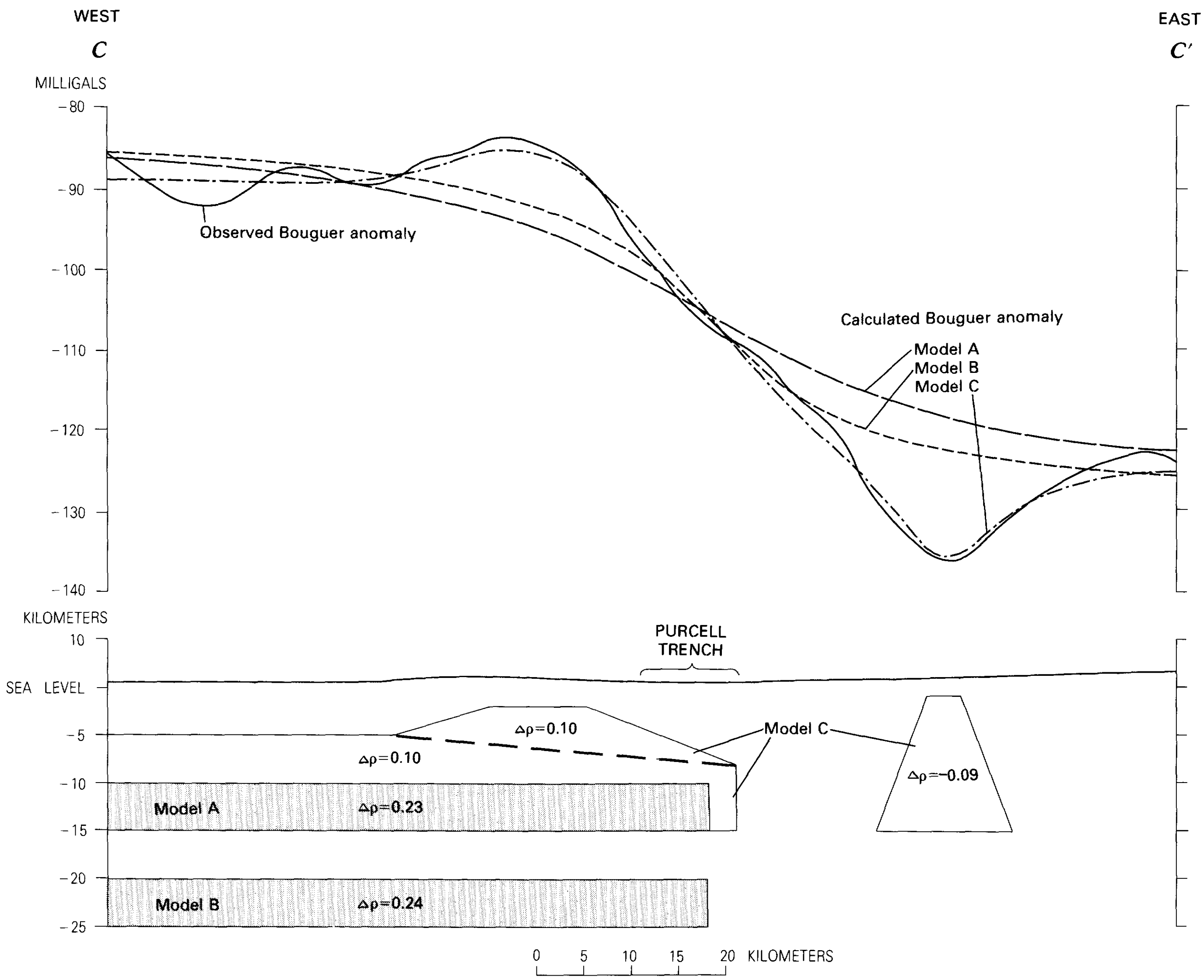


FIGURE 4.—Detailed gravity profiles and models across the Purcell trench along line *C–C′* (pl. 1*A*). Models A and B are two-dimensional. Model C is 2½-dimensional; strike length is 150 km below dashed line, 40 km above dashed line, and 50 km for isolated buried pluton to the east. Density contrast, Δρ, in g/cm^3. Arbitrary constants have been subtracted from the calculated gravity to allow comparison with the observed Bouguer anomaly.

to the empirical curve of Bateman and Eaton (1967). We assume that density increases downward in the crust, and, following Prodehl (1970), we assume that the lower crust contains rocks of intermediate density ($\rho = 2.8$–3.0 g/cm^3), even where these intermediate-density rocks have not been detected by raypath analysis of seismic-refraction profiles. The purpose of extending the models so far to the west, south, and east is to provide a regional setting for modeling the Omineca crystalline belt. The gravity and topographic data are more precise near the center of the model than near the ends, because they were taken from larger scale maps with closer contour intervals. The topography along each profile was determined from topographic maps by averaging values within a zone 10 km wide perpendicular to the profile and 5 km wide parallel to the profile. To minimize errors in calculated gravity at the ends of the profiles, the structure at the end of each profile was extended as near-infinite horizontal layers. Earth curvature has been ignored. The models are infinite perpendicular to the plane of the profile (two-dimensional). The actual geology is not precisely two-dimensional, especially on the north-south line (profile *B′–B″*, pl. 1*A*), but most of the features included on the model have strike lengths wider than their depth of burial. Calculations from a sample two-dimensional model were compared with those from a 2½-dimensional model, and no significant differences were noticed.

EAST-WEST GRAVITY MODEL

Model A (fig. 5) is a simplified, purely hypothetical model, extending from the Cascade Range in Washington across the Omineca crystalline belt to the Purcell fold belt in Montana, along section A–A'''. It tests the hypothesis that gravity variations along the profile could result entirely from mantle relief predicted by Airy isostatic compensation of topography. In the model, relief in the boundary between the upper mantle, having a density of 3.21 g/cm^3, and the lower crust, having a density of 2.90 g/cm^3, can explain all but the steepest gradients in the observed gravity profile. The relief in the upper mantle must, however, be 10 to 20 km to produce the observed anomalies, amounts much greater than the relief predicted by isostasy. For example, the surface elevation just west of the Purcell trench (at lat 48°13′ N., long 116°43′ W.), averaged over a 20-km radius, is about 800 m, whereas the elevation east of the Purcell trench (lat 48°09′ N., long 115°52′ W.) is about 1300 m. To isostatically compensate for the difference of 0.5 km, assuming a surface density of 2.72 g/cm^3 and a density contrast at the base of the crust of 3.21–2.90 = 0.31 g/cm^3, would require mantle relief given by the equation

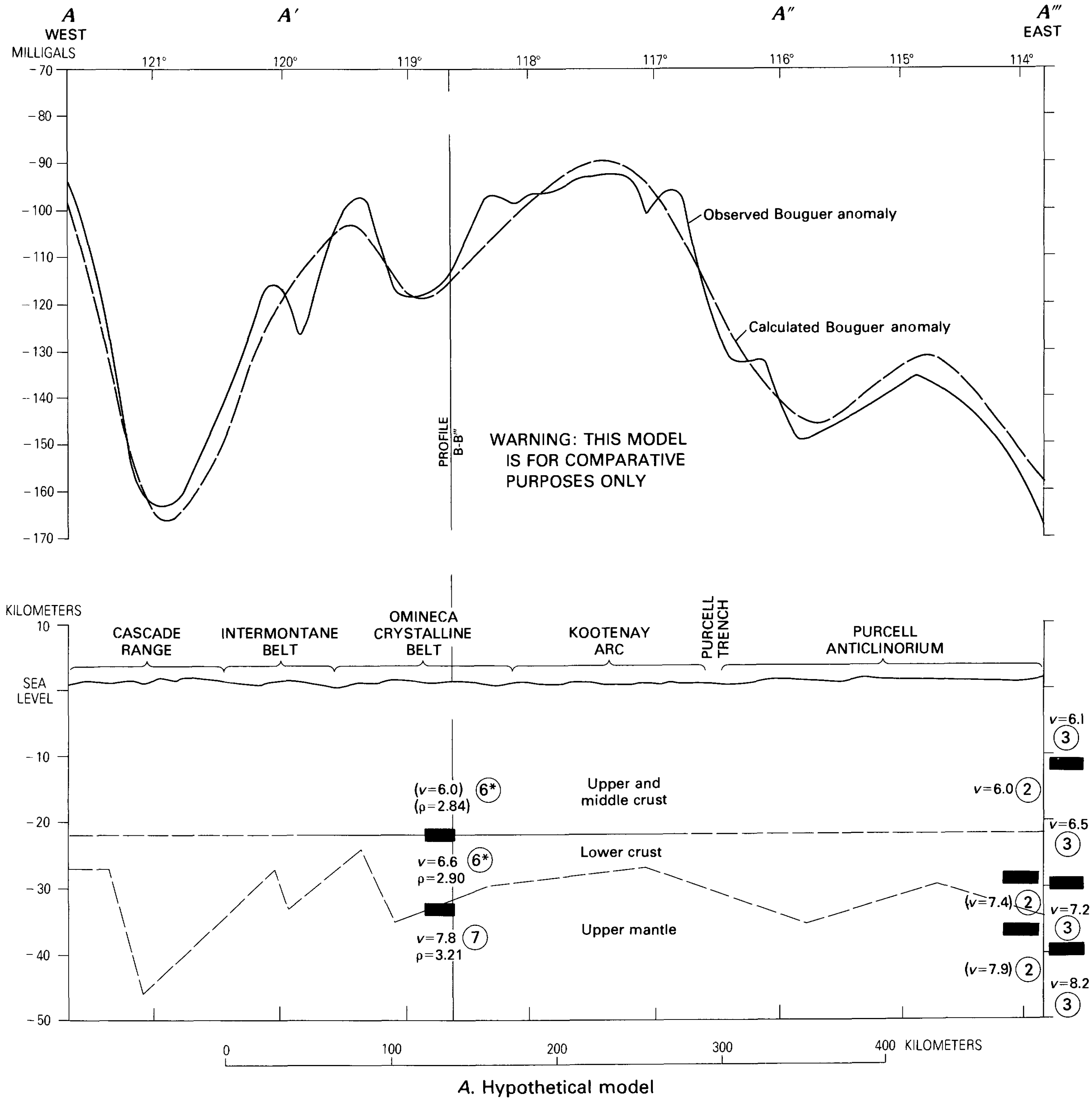
$$\text{relief} = 0.5\ \text{km} \times 2.72\ \text{g/cm}^3/0.31\ \text{g/cm}^3 = 4.39\ \text{km}.$$

The mantle relief required in the model to cause the observed gravity gradient is 9 km, or almost twice the relief predicted from isostasy. Hence, density variations within the crust are required to explain the observed anomalies. Because they are closer to the observer, smaller, isostatically stable mass differences in the crust can cause the same anomalies as isostatically unbalanced relief in the crust-mantle boundary.

Model B (fig. 5) is geologically more realistic, in that it shows lateral density variation within the crust. It agrees with Hill's (1972) unreversed north-to-south seismic-refraction interpretation at their intersection, with a single exception. We show an intermediate crustal layer, having a density of 2.8 g/cm^3, and it is discussed below in the section on "North-south gravity model." The dense lower crust ($\rho = 3.08$ g/cm^3) of Kanasewitch and others (1969) and Steinhart and Meyer (1961, chap. 9) has been grouped with the 2.90 g/cm^3 lower crustal layer at the east edge of the model. The model does not include the low-velocity upper crust ($v = 5.95$ km/s) and low-velocity upper mantle ($v = 7.94$ km/s) of Steinhart and Meyer, which have not been confirmed by later studies (Hill, 1978). A low-velocity upper mantle ($v = 7.8$ km/s, $\rho = 3.21$ g/cm^3) under the Omineca crystalline belt is required to satisfy the seismic observations of Cumming and others (1979), made at lat 50° N. The location of the transition to normal-velocity crust has not been determined seismically, but from other geophysical data (in the section on "Other geophysical studies") we assume that anomalous mantle underlies the entire Omineca crystalline belt and terminates to the east under the Purcell trench. Mantle relief is approximately that predicted by Airy isostatic compensation of topography. The upper boundary of the 2.9 g/cm^3 layer lies at a depth of 22 km in the center of the model, as estimated by Hill (1972). This boundary, in the model, is depressed under the Cascade Range, on the assumption that low-density plutons extend to considerable depth. The lower crust arches a few kilometers higher in the east, under the gravity high over the Purcell anticlinorium, somewhat as indicated by Wynn and others (1977). The uppermost crust was assumed to have a density of 2.72 g/cm^3. This assumption is supported in the east by density measurements of Belt rocks by Harrison and others (1972). For simplicity, the same density was assumed in the west, where samples (discussed above) revealed average densities of 2.65, 2.67, 2.73, and 2.75 g/cm^3 for various kinds of granitic and metamorphic rocks.

By taking the above-mentioned densities as fixed, and by allowing the boundaries to vary within the limits set by seismic-refraction data, we hypothesized that a layer of high-grade metamorphic rocks, having densities intermediate between 2.72 and 2.90 g/cm^3, could cause the gravity highs observed over the Omineca crystalline belt. To create the model, we made repeated adjustments to the boundary between the hypothesized high-grade metamorphic zone and the overlying rocks having densities of 2.72 g/cm^3. In each trial, the density of the high-grade zone was determined by least-squares analysis. When an acceptable fit to the observed data was found, the least-squares routine yielded a density of 2.84 g/cm^3 for the high-grade metamorphic zone. This density is higher than the measured average density (2.75 g/cm^3) of the Tonasket Gneiss, but is well within its density range (2.61 to 3.05 g/cm^3).

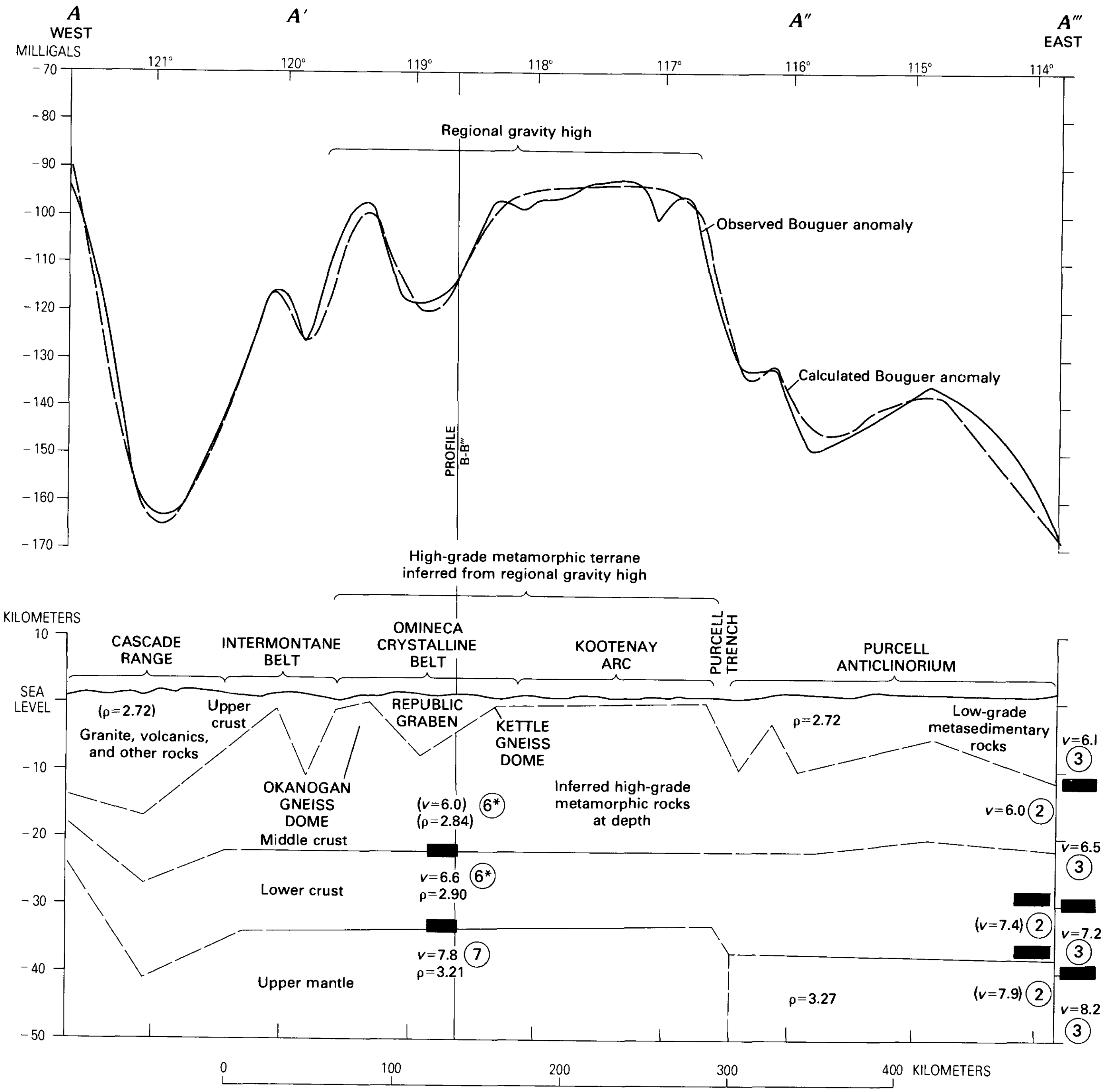
The modeling results and the observed correlation of local gravity highs with gneiss domes suggest that the regional gravity high over the Omineca crystalline belt is caused by intermediate density, high-grade metamorphic rocks that are warped upward and reach the surface in the cores of the gneiss domes. This model indicates that the cores of gneiss domes should be denser than the surrounding rocks. This concept contradicts the common observation that the cores of gneiss domes contain felsic, easily melted granite and migmatite that were diapirically emplaced because of their low density. Our gravity model is supported by our limited sampling of the Okanogan gneiss dome, but published density data are unavailable for other gneiss domes of the Omineca crystalline belt. The gravity data clearly show that dense rocks are present in or below the gneiss



A. Hypothetical model

FIGURE 5.—Gravity profiles and hypothetical model (*A*) and preferred model (*B*) of crustal structure along line *A–A‴* on figure 1. Solid lines in the gravity model indicate boundaries that are corroborated by seismic evidence; all other boundaries are dashed. Solid rectangles also show seismic boundaries; circled numbers are the seismic-refraction studies cited in figure 3. v, compressional-wave velocity in km/s; values in parentheses were ignored in the present study. ρ, density in g/cm^3, estimated from seismic velocities using an empirical velocity-density relation (Bateman and Eaton, 1967) or from surface density measurements (for example, 2.72 in Purcell anticlinorium in *B*); values in parentheses satisfy the gravity data but are not supported by seismic evidence in the vicinity of the section. The velocities labeled 6* should not be considered as definitive values for crustal velocity in the southern Omineca crystalline belt (fig. 3).

domes within the upper 5 to 10 km of the crust. If subsequent sampling should show that the dense rocks are not exposed at the surface, then dense rock must systematically underlie the gneiss domes.

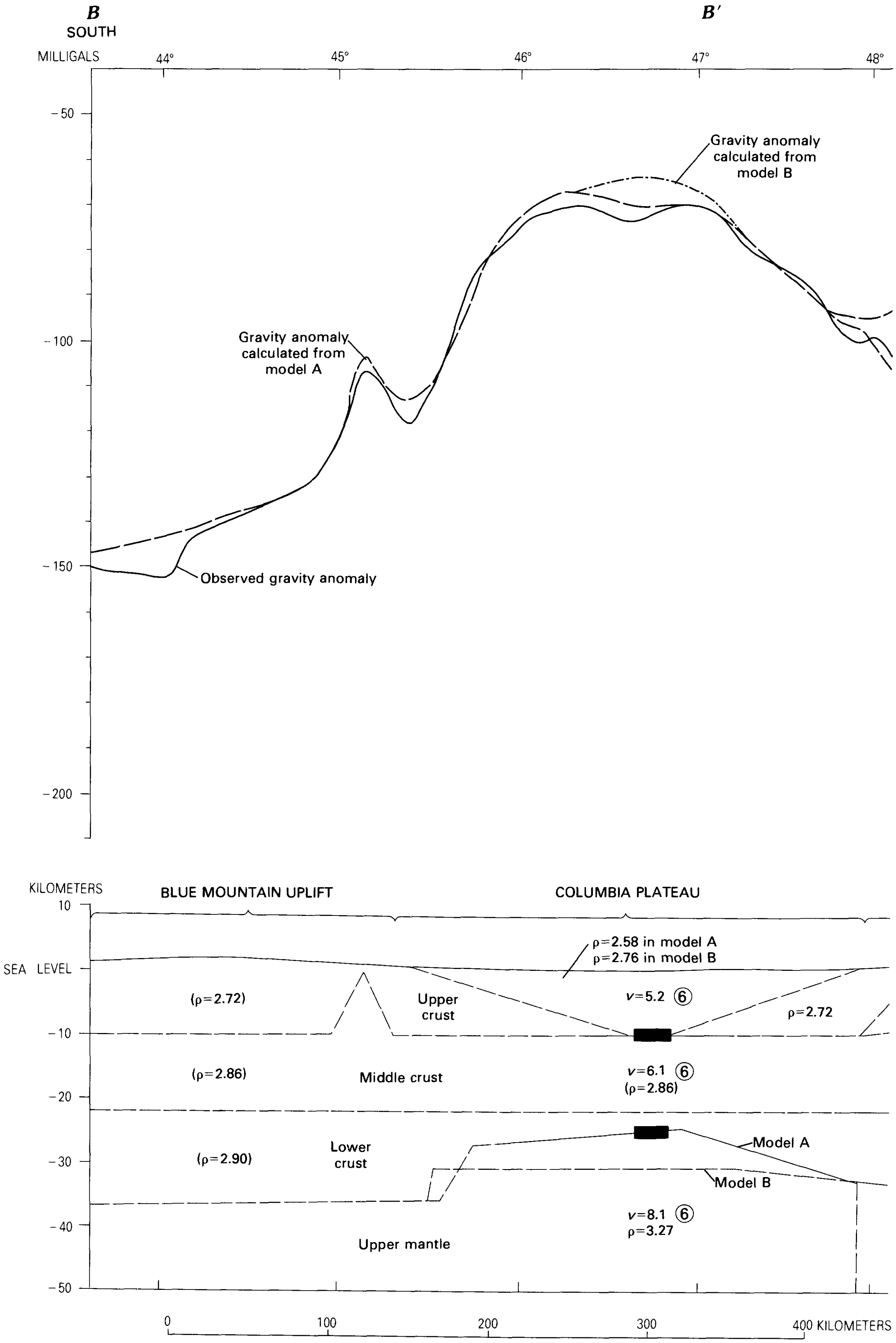


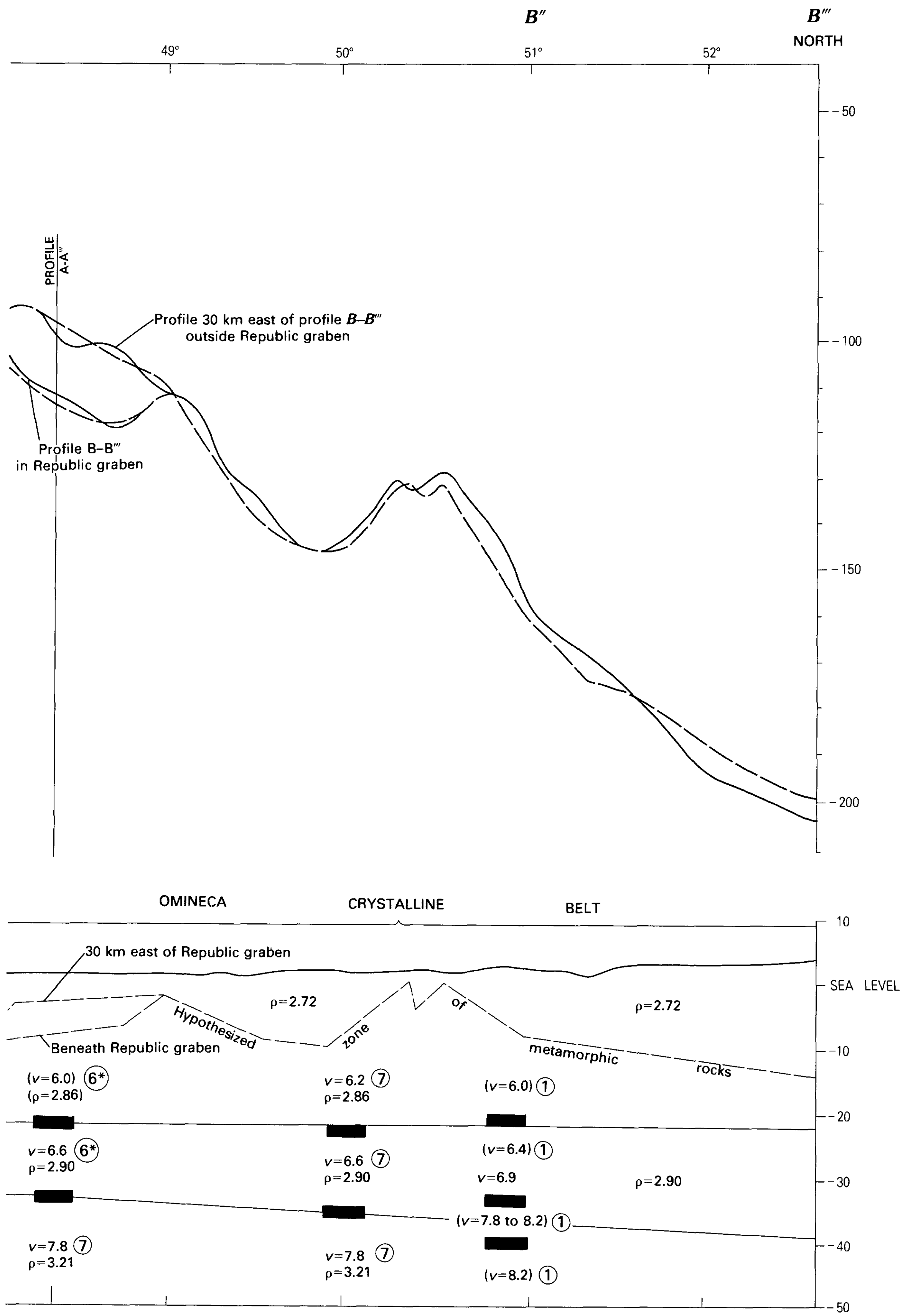
B. Preferred model

NORTH-SOUTH GRAVITY MODEL

Model A (fig. 6) was constructed for north-south profile *B–B‴* that runs close to long 118° W. from the high lava plateaus of Oregon across the Blue Mountain uplift, Columbia Plateau, and Omineca crystalline belt to the vicinity of Revelstoke, B. C. The model is compatible with, but does not necessarily corroborate, a

FIGURE 6 (following page).—Gravity profiles and models of crustal structure along line *B–B‴* in figure 1. Solid lines in the gravity model indicate boundaries that are corroborated by seismic evidence; all other boundaries are dashed. Solid rectangles also show seismic boundaries; circled numbers are the seismic-refraction studies cited in figure 3. *v*, compressional-wave velocity in km/s; values in parentheses were ignored in the present study. ρ, density in g/cm³, estimated from seismic velocities using an empirical velocity-density relation (Bateman and Eaton, 1967); values in parentheses satisfy the gravity data but are not supported by seismic evidence in the vicinity of the section.


B
SOUTH
B′
MILLIGALS
44°
45°
46°
47°
48°
-50
-100
-150
-200
Gravity anomaly calculated from model B
Gravity anomaly calculated from model A
Observed gravity anomaly
KILOMETERS
10
SEA LEVEL
-10
-20
-30
-40
-50
BLUE MOUNTAIN UPLIFT
COLUMBIA PLATEAU
ρ=2.58 in model A
ρ=2.76 in model B
(ρ=2.72)
Upper crust
v=5.2 6
ρ=2.72
(ρ=2.86)
Middle crust
v=6.1 6
(ρ=2.86)
(ρ=2.90)
Lower crust
Model A
Model B
v=8.1 6
ρ=3.27
Upper mantle
0
100
200
300
400 KILOMETERS


B″
B‴
NORTH
49°
50°
51°
52°
-50
-100
-150
-200
PROFILE A-A‴
Profile 30 km east of profile B–B‴ outside Republic graben
Profile B-B‴ in Republic graben
OMINECA
CRYSTALLINE
BELT
10
SEA LEVEL
-10
-20
-30
-40
-50
30 km east of Republic graben
Beneath Republic graben
Hypothesized zone of metamorphic rocks
ρ=2.72
ρ=2.72
(v=6.0) 6*
(ρ=2.86)
v=6.2 7
ρ=2.86
(v=6.0) 1
v=6.6 6*
ρ=2.90
v=6.6 7
ρ=2.90
(v=6.4) 1
v=6.9
ρ=2.90
(v=7.8 to 8.2) 1
v=7.8 7
ρ=3.21
v=7.8 7
ρ=3.21
(v=8.2) 1

seismic interpretation by Hill (1978) that included data from an unreversed north-to-south seismic-refraction profile (Hill, 1972) and time-term studies of crustal structure from a seismograph array in the Pasco basin (J. P. Eaton, oral commun., 1976). The boundary between low-velocity upper mantle, observed by Cumming and others (1979) in the Omineca crystalline belt, and normal upper mantle, observed by J. P. Eaton in the Pasco basin, is arbitrarily drawn at the boundary between the Omineca crystalline belt and the Columbia plateau. Presumably, the boundary follows the south side of the Omineca crystalline belt to the Purcell trench, where it turns north. The model approximately agrees with east-west model B at their mutual intersection, and similarly uses density variations within the crust to explain gravity highs over the Omineca crystalline belt.

Although there is a large Bouguer gravity high over the Columbia plateau (really a basin), the high is lower in amplitude than would be predicted from isostatic reasons, solely because of the low elevation of the plateau. Hence, there is a mass deficiency associated with the plateau. According to J. P. Eaton (oral commun., 1976), compressional-wave velocities in the Pasco basin (central part of the Columbia plateau) are only 5.2 km/s to a depth of 10 km. Bateman and Eaton's (1967) velocity-density curve predicted a density of 2.58 g/cm^3—low for basalt, but possible if the basalt is highly vesicular or contains abundant joints and interflow zones.

In the north-south model A (fig. 6), a wedge-shaped zone having a density of 2.58 g/cm^3 and a maximum thickness of 10 km represents the basalt filling of the Pasco basin. A corresponding bulge in the mantle, having a minimum depth of 25 km, explains the gravity high over the Pasco basin. The gravity gradient near lat 45°40′ N. is explained entirely by sharp relief in the mantle, because the low density predicted by seismic results for the Columbia River Basalt Group precludes the source of this gradient from being in the upper crust. A regional gravity map (Eaton and others, 1978) shows that this gradient zone (*E–E′* in fig. 2) has a length of about 300 km, southwest to northeast. The zone may mark a major tectonic transition between shallow mantle in the northwest and deeper mantle in the southeast. The intracrustal boundary shown by Hill (1972) at a depth of 22 km in the Omineca crystalline belt has been extended all the way across the section, because, in the absence of evidence to the contrary, density probably increases gradually with depth in the crust.

In model A, as in model B (fig. 5), the densities discussed above were taken as fixed. Deep boundaries were permitted to vary within the range allowed by seismic-refraction data, the top of a hypothetical intermediate crustal layer was adjusted to improve the fit with the observed gravity data, and the density of this intermediate crustal layer was determined by least squares. This density was 2.86 g/cm^3—not significantly different from the 2.84 g/cm^3 determined on the north-south profile.

The north-south profile, chosen to follow Hill's (1972) seismic-refraction line, follows the east side of the Republic graben, which is filled with low- density sedimentary and volcanic rocks. The gravity along this line, shown by a solid line on figure 6, can be seen to be 10 to 20 mgal lower than it is along a parallel line 30 km to the east, between lat 48° N. and lat 49° N. A variation of the top of the 2.86 g/cm^3 layer was calculated to match the gravity along the alternate line, and is shown in the model in figure 6 by the line labeled "east of Republic graben."

J. P. Eaton's (oral commun., 1976) seismic-array studies are not the only determinations of seismic velocity and, by inference, density in the Pasco basin. Seismic-delay time and gamma-gamma density logs were made in a well 3250 m deep in the Rattlesnake Hills, Wash., at the edge of the Pasco basin (Raymond and Tillson, 1968). Seismic velocities of about 6.5 km/s and densities of about 2.76 g/cm^3 were found in most of the coherent basalt flows. These values plot on Bateman and Eaton's (1967) empirical velocity-density curve, so they can be considered reliable, at least on the small scale of borehole measurements. Lower values were found in the less coherent interflow zones.

Model B was made to test the effect of using 2.76 g/cm^3 instead of 2.58 for the Columbia River Basalt Group. The only difference from model A incurred by this change was that the top of the mantle was lowered from -25 km to -31 km, under the Pasco basin. The density of the intermediate crustal layer, determined by least squares, was also approximately 2.86 g/cm^3.

INTERMEDIATE-DENSITY CRUSTAL LAYERS

How much confidence can we place in the existence of an intermediate-density (2.84 to 2.86 g/cm^3) crustal layer? Following is a summary of evidence, both for and against, the layer.

There is nothing sacred about the density of 2.84 to 2.86 g/cm^3 determined by means of the gravity modeling program. An empirically determined constant was subtracted from the gravity effect of the prismatic model to allow comparison between relative changes in the calculated gravity and relative changes in the observed negative Bouguer anomaly. Hence, information about the total gravity effect of any layer is lacking. The calculated density is simply that required to cause the variations in gravity associated with the high-grade metamorphic belt, assuming that relief in the top of the layer is about 20 km, as shown in the models. From restored stratigraphic cross sections and metamorphic-grade considerations, Campbell (1973) argued that the crystalline belt was uplifted about 11 km relative to the Rocky Mountains. If the density of the anomaly-producing rocks were less than 2.84 to 2.86 g/cm^3, then the relief in the causative layer would have to be greater than that estimated by Campbell.

Seismic studies are ambivalent about the presence of an intermediate-density crustal layer. Because Hill's (1972) refraction profile was unreversed, he could not determine absolute velocities in the presence of possibly dipping layers. He observed an apparent velocity of 6.2 km/s for the P crustal phase, but assumed an upper crustal velocity of only 6.0 km/s, based upon data obtained in British Columbia west of the Omineca crystalline belt (White and others, 1968). More recent studies (Cumming and others, 1979) show upper crustal velocities of 6.2 km/s in the Omineca crystalline belt near lat 50° N. Furthermore, Hill's profile may not be representative of the Omineca crystalline belt, because north of lat 48° N. it is in the Republic graben.

Further south, in the Pasco basin, neither Hill (1972) nor Eaton (quoted in Hill, 1978) detected an intermediate layer. The intermediate layer is extended to the south edge of the gravity model for simplicity, but no claim is made for its presence there.

Cumming and others (1979) observed a P-wave velocity of 6.2 km/s in the upper crust of the Omineca crystalline belt at lat 50° N. The Bateman and Eaton (1967, fig. 5) empirical velocity-density curve predicted a density of only 2.76 g/cm^3 for a velocity of 6.2 km/s. Some rock types, however, have a higher density than this curve predicts. For example, the quartz diorite gneiss shown on the Bateman-Eaton graph would have a P-wave velocity of 6.2 km/s and a density of 2.86 g/cm^3 at a pressure of 3 kilobars (3×10^8 newtons per square meter). Hence, the density of 2.84 to 2.86 g/cm^3 calculated for the gravity models lies within the range of density expected from seismic observations further north.

The final, and perhaps most important, evidence for an intermediate-density crustal layer is the association of a regional gravity high with the Omineca crystalline belt, and local gravity highs with individual gneiss domes. Qualitatively, these data suggest a merging at depth of the dense cores of gneiss domes into a continuous dense crustal layer.

AEROMAGNETIC STUDIES

MAJOR FEATURES OF THE AEROMAGNETIC MAP (PLATE 1*B*)

Plate 1*B* is a generalized aeromagnetic map showing locations of selected magnetic highs. These highs were taken from a number of original surveys, each of which had its own specifications, that is, direction, line spacing, elevation, contour interval, and datum. Thus, the identification of the features shown on plate 1*B* is subjective and qualitative. The reader is cautioned to use the original sources, which are identified on an index figure on plate 1*B*, for detailed interpretation of any particular area.

Aeromagnetic highs occur over some plutonic rocks, particularly those bearing hornblende and biotite, over the Eocene volcanic rocks, and over a small fraction of the high-grade metamorphic rocks. The short-wavelength magnetic highs in the Republic graben, labeled A, are an example of the anomalies caused by Eocene volcanic rocks.

Most of the magnetic highs over plutonic rocks occur over the hornblende-biotite-bearing granitic rocks. Examples of these anomalies are labeled B. Many areas of hornblende-biotite rock are unaccompanied by magnetic highs, however. Several magnetic highs, labeled C, occur in areas of biotite-muscovite and biotite-bearing granitic rocks. The magnetic signature is a useful but imperfect guide to plutonic rock type.

More often than not, the high-grade metamorphic rocks are unaccompanied by magnetic highs. The Thor-Odin, Passmore, and Kettle domes are magnetically featureless. There are, however, several notable exceptions.

Magnetic-high D, accompanied by a local gravity low, occurs in metamorphic rocks west of the Thor-Odin dome. The geologic map of Jones (1959) shows a north-south fault close to the axis of the magnetic high, but no lithologic change is shown that could account for the geophysical anomalies.

High E occurs over granodiorite gneiss and layered sedimentary gneiss on the north flank of the Valhalla gneiss dome (Reesor, 1965). The magnetic high cuts across lithologic boundaries and continues to the northeast into an area of biotite or biotite-muscovite granitic rocks (labeled C).

High F occurs over an area of layered gneiss and both biotite-muscovite and biotite-bearing granitic rocks mapped by Little (1960). Little's geologic map shows nothing that would explain the sharp northern and southern boundaries of the magnetic highs.

High G occurs over the central part of the Okanogan gneiss dome, close to the axis of the gravity high. The geologic map of Snook (1965) shows nothing that can explain the magnetic anomaly.

High H occurs over high-grade metamorphic rocks north of the Kettle dome. The cause of the magnetic high is unknown.

The magnetic expression of rocks is primarily a function of the abundance of the accessory mineral, magnetite, and may vary according to original composition, regional metamorphic grade, and contact metamorphism. Although we see in the study area a predominance of magnetic highs over hornblende-biotite plutonic rocks and featureless magnetic patterns over layered gneiss and biotite and biotite-muscovite plutonic rocks, the exceptions are significant. Often the boundaries of coherent magnetic highs cut across mapped rock units. Rarely do the magnetic patterns, at least as they are plotted on the scale of plate 1*B*, have shapes that suggest that they arise from the structure of gneiss domes, contact-metamorphic aureoles around plutons, or zonations within plutons. (Two exceptions are ring-shaped highs B* and C* east of the Purcell trench between lat 49° and 50° N., which wrap around plutons.) A possible explanation for the chaotic magnetic pattern is that it results primarily from differences in original composition, whereas subsequent plutonism, metamorphism, and deformation have complicated the pattern without imprinting a new coherent pattern.

In the section on "Other Geophysical Studies," we cited regional aeromagnetic data (Caner, 1970; Haines and others, 1971) that showed that there is an aeromagnetic quiet zone associated with the Omineca crystalline belt. The regional extent of our data is insufficient to permit correlation with the regional aeromagnetic data cited above, except that the absence of long-wavelength anomalies is consistent with the existence of the magnetic quiet zone and the hypothesis that it is related to the cordilleran thermal anomaly.

GEOPHYSICS AND URANIUM EXPLORATION

Geophysics is most useful as an exploration tool if there is a sound geological model that can predict specific geophysical anomalies for which to look. Nash (1979) discussed five environments favorable to uranium occurrence in northeast Washington, but his descriptions are not geophysically explicit. Furthermore, the regional geophysical data discussed in this report are not detailed enough to locate most mineral deposits. Acknowledging these limitations, we briefly discuss the published geophysical data and how it might bear on uranium exploration.

MAGNETIC ANOMALIES AND BIOTITE AND BIOTITE-MUSCOVITE GRANITES

Nash (1979) cited the importance of biotite and biotite-muscovite granites as a source of leachable uranium because these granites contain abundant uranium (mean 8.8 ppm or parts per million), and the uranium occurs in labile phases, such as biotite and magnetite. We showed above that biotite and biotite-muscovite granites can generally be distinguished from biotite-hornblende granite by the presence of magnetic lows. Hence, the primary use of aeromagnetic data would be to use magnetic lows to identify uranium-bearing granite.

A secondary use of aeromagnetic data, however, would be to identify concentrations of uranium-rich magnetite. Nash (1979, p. 25) cited analyses of muscovite-bearing porphyritic quartz monzonite that contains as much as 180 ppm uranium. The aeromagnetic map shows that most of the muscovite-bearing granitic rocks are indicated by low magnetic relief and, hence, have a low magnetite content. The magnetic highs labeled C on plate 1*B*, however, indicate magnetite concentrations in muscovite-bearing granitic rocks. If the magnetite responsible for these magnetic highs is uraniferous, it could be a source of leachable uranium (Nash, 1979, p. 19). More accurate aeromagnetic maps might be useful in locating smaller, possibly economic, concentrations of radioactive magnetite.

GEOPHYSICAL SETTING OF THE MIDNITE MINE

The Midnite mine was cited by Nash (1979, p. 33) as an example of a contact metasomatic deposit. The metasomatic model, combined with geophysical data, can be used to predict the occurrence of similar deposits along the trend containing the Midnite mine. However, the published geophysical data do not have the resolution needed to identify anomalies associated with most ore deposits.

Uranium at the Midnite mine (found in autunite, pitchblende, and coffinite) is produced from a tabular zone in a roof pendant of metamorphosed Precambrian pelitic and calcareous rocks adjacent to its basal contact with a Cretaeous(?) porphyritic quartz-monzonite pluton (Nash and Lehrman, 1975). The Precambrian rocks are part of the Proterozoic Y Togo Formation (Campbell and Loofbourow, 1957; Becraft and Weis, 1963, p. 6). According to Nash (1979, p. 14), the quartz monzonite has an average uranium content of 14.7 ppm, and, because the uranium resides in the labile phases magnetite and biotite, it could be the source of the uranium in the deposits. For comparison, the uranium content of 78 samples from 32 granitic bodies in north-central and northeastern Washington ranges from 0.44 to 12 ppm and averages 2.83 ppm (Munroe and others, 1975).

As is true for the region (pl. 1*A*), gravity lows occur over most granitic rocks in the vicinity of the Midnite mine (fig. 7), and gravity highs occur over adjacent metasedimentary rock (for example, G1 and G2 in fig. 7). High G1 occurs over eugeosynclinal rocks assigned to the Ordovician Covada Group by J. R. Snook (oral commun., 1980). The magnetic-anomaly pattern over the low-grade metamorphic rocks shows only low relief, though the corresponding gravity-anomaly pattern is complex.

An irregular, elongate gravity high, G2, trends eastward, crossing the boundary between the eugeosynclinal and miogeoclinal provinces, and it also crosses the boundary between the Paleozoic rocks of the miogeoclinal province and the Precambrian rocks. The western lobe of the high is probably caused by the density contrast between rocks of the Covada Group and hornblende- and biotite-bearing granitic rocks to the south. This high reaches its maximum amplitude over Precambrian metasedimentary rocks and greenstone north of the Togo Formation. Much of the greenstone is metabasalt (Campbell and Raup, 1964). It may be, at least in part, the source of the gravity high over the Precambrian rocks.

The southwestern part of a conspicuous, northeast-trending gravity low (GL1) is associated with porphyritic quartz monzonite (assigned to the Loon Lake Granite by Becraft and Weiss, 1963) east and south of the Midnite mine. A detailed gravity profile across the area in the immediate vicinity of the Midnite mine showed Bouguer anomaly values 2 to 4 mgal lower over the quartz monzonite than over the Togo Formation. Limited sampling around the Midnite mine shows an average density of 2.64 g/cm^3 for the quartz monzonite and 2.72 g/cm^3 for the Togo Formation. The extent and orientation of the gravity low suggests that the porphyritic quartz monzonite is part of a large northeast-trending body of relatively low-density plutonic rock that underlies the area of GL1.

The gradient between the northeast-trending high and low anomaly pair (G2 and GL1) probably marks the boundary between quartz monzonite to the southeast and metamorphic rock of the Togo Formation to the northwest. The Midnite mine lies close to this boundary. If this boundary played any role in the formation of the uranium deposits, then other places along the gravity gradient would be likely targets for exploration.

Because the Midnite mine occurs at a roof pendant of the Togo Formation in the quartz monzonite (Nash and Lehrman, 1975), other roof pendants are possible exploration targets. Several small magnetic highs occur southeast, east, and northeast of the Midnite mine, which could indicate foundered roof pendants. The high centered 8.5 km southeast of the mine (M1) occurs over a contact between porphyritic quartz monzonite and the Cretaceous Loon Lake Granite. Wildcat drilling in the area of this high (J. T. Nash, oral commun., 1975) revealed metasedimentary rock, perhaps a foundered roof pendant, at a depth of 300 m. No uranium was found.

About 10 km northwest of the Midnite mine is a magnetic high (M2, fig. 7) coincident with gravity low GL2. Much of the anomaly area is covered by Tertiary rocks, but outcrops of the Cretaceous Loon Lake Granite are abundant. The Tertiary rocks include sedimentary, pyroclastic, and volcanic rocks of the Gerome Andesite of Becraft and Weis, 1963, (p. 33–38). These are the northerly prong of a narrow north-northwest-trending belt (pl. 1*B*). The gravity anomaly is centered over the northerly sector of this belt of volcanic rocks, and the associated north-northwest-trending maximum of the magnetic high coincides closely with their mapped distribution in this area. These anomalies are superimposed on an oval, northeast-trending magnetic anomaly (M3, fig. 7), in part of lower magnitude. The M3 anomaly may indicate the outline of a granodioritic pluton of the hornblende-biotite suite.

The peculiar pattern of the peaks of the magnetic highs (M2 and M4), which wrap around just inside the north and east borders of the inferred pluton (defined by high M3) and cut across its center, might be explained by zonation within the pluton.

URANIUM DEPOSITS AND METAMORPHIC ROCKS

Nash (1979) cited several areas of northeastern Washington where there is a potential for Rossing-type uranium deposits. Uranium in such deposits is concentrated in anatectic granite and migmatite derived from reworked sialic crust. High uranium concentrations are expected "in the more deeply eroded portions of the orogen where basement rocks, high-rank metasedimentary rocks, and anatectic granites are closely intermingled" (Rogers and others, 1978, p. 1547). We have inferred that local gravity highs indicate upwarped high-metamorphic-grade domes within the metamorphic terrane, where we would expect to find the most deeply eroded section. The Ferry Mountains, cited by Nash (1979) as a Rossing-type terrane containing uranium-rich rocks, coincide with the Kettle Dome and its associated gravity high shown in plate 1. Other gravity highs indicate deeply eroded high-grade metamorphic rocks that may also have uranium potential.

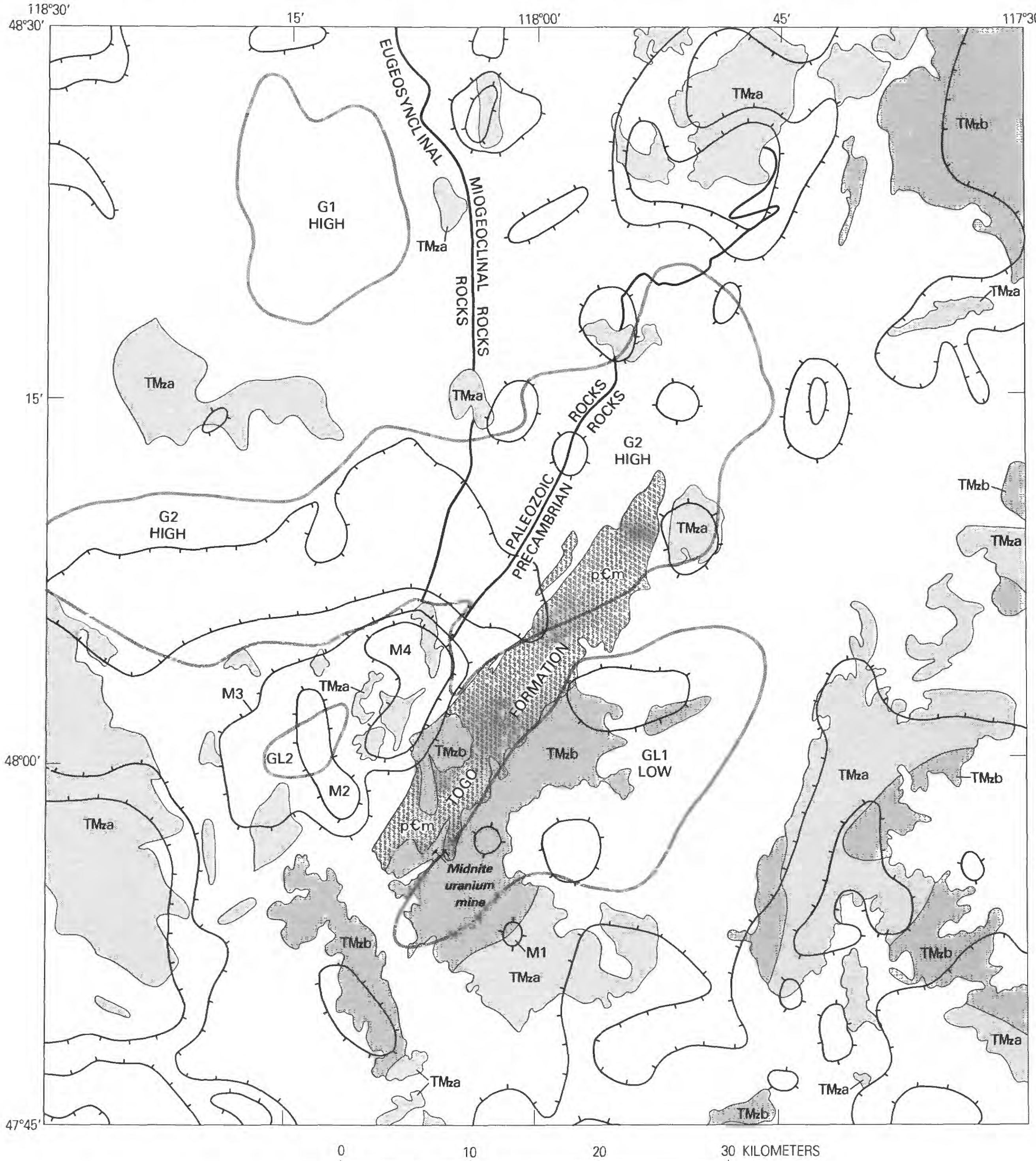


FIGURE 7.—Sketch map of the Midnite uranium mine area showing geophysical and geological setting and selected rock units. Letters and number (G1, M2) indicate gravity and magnetic anomalies discussed in test. Geology simplified from Miller and Yates (1976), Griggs (1973), Becraft and Weiss (1963), and Fox and others (1977). Aeromagnetic contours simplified from U.S. Geological Survey (1973, 1974a). Gravity contours simplified from Cady and Meyer (1976).

Rocks at or near the periphery of the gneiss domes may also be favorable sites for uranium concentration. The forceful juxtaposition of the gneiss domes and country rocks may have mechanically fractured the country rocks near the domes, creating fracture porosity and permeability. The weakly metamorphosed or un-

EXPLANATION

TERTIARY AND MESOZOIC:

TMza — Biotite-hornblende and pyroxene-amphibole-bearing granitic rocks

TMzb — Biotite-muscovite and biotite-bearing granitic rocks

PRECAMBRIAN:

p€m — Metamorphosed pelitic and calcareous rocks (part of miogeoclinal deposits, pl. 1)

CONTACT

SELECTED GRAVITY CONTOUR—G1 and G2 are highs discussed in text, and GL1 and GL2 are lows discussed in text

SELECTED MAGNETIC CONTOUR—Inward-directed hachures indicate lows, outward-directed hachures indicate highs. M1, M2, M3, and M4 are selected magnetic anomalies discussed in text

All other rocks types are unpatterned, including Tertiary basalt, Tertiary sedimentary and volcanic rocks, eugeosynclinal deposits, miogeoclinal deposits other than Togo Formation, and gneisses of the metamorphic core complexes. For more detail, see plate 1.

metamorphosed country rock, containing abundant hydrous minerals (chlorite, muscovite, and amphibole) or connate water, was a potential source of water for hydrothermal systems. The gneiss dome or core metamorphic complex was a source of labile uranium and primary water, and its heat at the time of emplacement could have produced a hydrothermal system. Finally, the steep thermal gradient at the margin of the dome may have facilitated precipitation of uranium from solution.

Similar views have been expressed by Cheney (1980). According to Cheney, "most pegmatitic deposits in the metamorphic rocks are too small or low grade to be commercial Rossing-type deposits. However, the uranium from such pegmatites may have been reconcentrated in Tertiary strata or in the cataclastic zones peripheral to the core complexes. Those in cataclastic zones may have formed in a manner similar to the unconformity-vein deposits of Saskatchewan or in zones of hydrothermal alteration. Thus, the margins of the core complexes deserve to be intensively prospected."

Regional gravity maps are a tool for locating the margins of core complexes, especially where the geology is very complex or the margins are covered by overlying rocks. The margins occur in the gradient zones flanking the gravity highs over the gneiss domes. In the case of completely buried gneiss domes, where mineralization may occur in Tertiary strata directly above the dome, the center of the gravity high is the preferred target.

CONCLUSIONS

The primary observation of this study is that the Omineca crystalline belt is accompanied by a regional gravity high, and that individual gneiss domes within the belt cause local gravity highs. The steepness of the gravity gradient along the east side of the regional gravity high shows that the source of the high must lie, at least in part, in the upper part of the crust. Regional gravity models, constrained by the available seismic-refraction data, show that the gravity highs can be explained by 10-km-high bulges in an intermediate crustal layer having a density of 2.84 to 2.86 g/cm^3. The geologic interpretation of the gravity models is that the core metamorphic complexes are the surface expression and perhaps exposures of the uppermost part of the dense infrastructure that makes up most of the upper 20 km of the crust within the crystalline belt.

Aeromagnetic data are much less useful than gravity data for interpreting crustal structure. The Omineca crystalline belt lies in a magnetic quiet zone, probably related to the cordilleran geothermal anomaly. The magnetic anomalies that do exist form a chaotic pattern that provides little regional information. Magnetic anomalies, however, provide an imperfect guide to plutonic rock type. Hornblende-biotite granites are often accompanied by magnetic highs, and biotite and biotite-muscovite granites often occur in magnetic lows.

Several geophysical methods may be useful in uranium exploration. Aeromagnetic lows can be used to distinguish biotite and biotite-muscovite granites, and the rare magnetic highs over biotite and biotite-muscovite granites can be used to locate concentrations of magnetite, some of which have been shown to contain uranium. Detailed aeromagnetic data may be useful for locating contact-metasomatic uranium deposits, but the published data are not adequate for this purpose. Finally, gravity highs show bulges of the high-grade metamorphic terrane that could be a favorable location for Rossing-type uranium deposits, and gradients surrounding these highs show the margins of metamorphic core complexes, sites for possible concentration of uranium derived from the core complexes.

The northwest-trending, steep gravity gradient marking the eastern side of the Omineca crystalline belt may be a site of uranium concentration. We have proposed above that the southeastern part of this zone marks the concealed eastern side of a bulge of dense infrastructure, represented and perhaps exposed to the west in the gneiss domes of the Omineca belt. If so, the geologic setting of rocks above and along the southeastern part of this zone may be favorable for uranium concentration.

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