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CORDILLERAN METAMORPHIC CORE COMPLEXES
—From Arizona to Southern Canada

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INTRODUCTION

Cordilleran metamorphic core complexes occur in a sinuous belt that lies west of the Cordilleran fold and thrust belt from Canada to California. It then continues southeastward through the Basin and Range country of Arizona, where it lies athwart the northeast edge of the fold and thrust belt across Arizona, before continuing south into Mexico (Figure 1). This curious geographic distribution, in addition to the relatively recent recognition of the young age of metamorphic fabrics in many areas, has attracted attention to these complexes and led to a variety of hypotheses for their origin.

The Shuswap complex in Canada is the largest and longest recognized metamorphic core complex and is considered the type example (Coney 1980). Only during the past two decades has an awareness of the complexes in the US blossomed. This awareness reached its fullest expression in the recent Geological Society of America memoir devoted to discussion of these metamorphic complexes (Crittenden et al 1980). Because of the richness of that source, this review focuses on ideas and interpretations rather than descriptive details, but some examples must be given to demonstrate both variability and common features.

The investigation of each complex has tended to follow a similar historical pattern. Prior to 1960 most were regarded as exposures of pre-Phanerozoic crystalline basement or granitic intrusive bodies, and thus not structurally active parts of the Mesozoic orogen. In the 1960s, field work and reconnaissance K-Ar geochronometry drew attention to the complexes as sites of Mesozoic and Cenozoic deformation and metamorphism affecting Phanerozoic supracrustal rocks. During the last two decades numerous detailed

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studies of these complexes have revealed a polyepisodic history, with metamorphic protolith ages ranging from early Precambrian to Cenozoic and deformation histories of varying complexity and length.

Coney (1980) reviewed much of the history of discovery and study of these complexes. In Canada the work of Wheeler (1963, 1965, 1966) and Reesor (1965, 1970, Reesor & Moore 1971) brought out the importance of domal structures along the eastern margin of the Shuswap complex. They reviewed evidence for Mesozoic metamorphism and its temporal overlap with emplacement of granitic rocks. Wheeler (1966) used the term Shuswap "metamorphic complex" and in later papers often referred to the area as a "metamorphic core zone." These terms were blended by Coney to "metamorphic core complex"; at the same time there was a shift in emphasis to the complexes further south, especially to those in Nevada and Arizona.

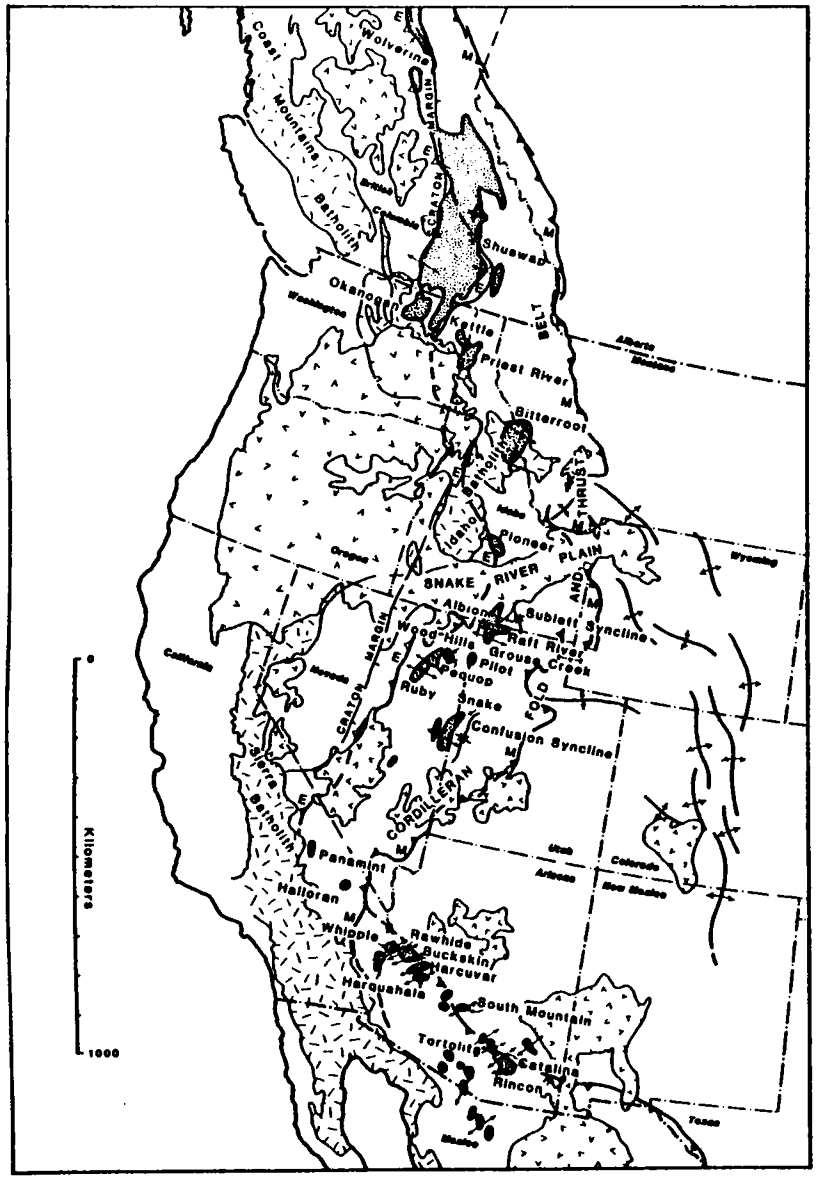

Figure 1 Distribution and tectonic setting of Cordilleran metamorphic core complexes (similar to figures published by Coney 1979, 1980). The complexes are heavily stippled, the trends of regionally consistent late lineations are shown by double headed arrows. Three boundaries of the craton edge are shown. From east to west these are (*a*) the east edge of the miogeosyncline, marked with M, usually a major thrust of the Cordilleran fold and thrust belt; (*b*) the eastern limit of eugeosynclinal rocks, marked with E, usually the limit of material thrust eastward over the western part of the continental margin sediment wedge (this tectonic contact is shown with open thrust symbols); and (*c*) the west limit of older Precambrian basement, labeled "CRATON MARGIN", as indicated by strontium isotope ratios in granitic rocks. The Cordilleran fold and thrust belt, with filled in thrust symbols, diverges from the miogeosyncline (M) in southern California and Arizona. The major Mesozoic batholiths are shown with a scrambled dash pattern and large patches of Cenozoic volcanic rocks with a scattered V pattern. Major basement uplifts of Laramide age are shown by anticline symbols and dejected belts in the hinterland of the fold and thrust belt by syncline symbols.

In the US the work of Peter Misch and his students and co-workers (Misch 1960, Misch & Hazzard 1962, Nelson 1969, Dover 1969, Thorman 1970) showed that Mesozoic metamorphism has affected rocks of the Cordilleran miogeosyncline in Nevada, Utah, and Idaho. These papers put emphasis on the low-angle "decollement" faulting of younger-on-older style, which is regionally associated with the metamorphic exposures.

Early hints at the importance of Cenozoic denudation (Jones 1963) and metamorphism (Mauger et al 1968) in Arizona, and the description by Davis (1975) of gravity gliding in the Rincon Mountains, led to the development of an influential "megaboudin" model for the Arizona core complexes (Davis & Coney 1979).

In a series of papers attempting to relate geochronometry and geologic information for the eastern Great Basin, I introduced the concept of a mobile Mesozoic "infrastructure" in the hinterland of the Sevier Orogenic Belt; pointed out that uplift and cooling of the metamorphic complexes was a Cenozoic phenomena; and suggested that many of the low-angle younger-on-older faults were probably Cenozoic extension and gravity slide structures (Armstrong 1964, Armstrong & Hansen 1966, Armstrong 1972). Controversy arose as various workers attempted to integrate the metamorphic areas into models for Cordilleran structural evolution—with varying degrees of success and agreement (Armstrong 1968b, Price & Mountjoy 1970, Mudge 1970, Campbell 1973, Price 1973, Hose & Daneš 1973, Roberts & Crittenden 1973, Scholten 1973, Fox et al 1977, Brown 1978, Allmendinger & Jordan 1981). Recent reviews of Cordilleran core complexes emphasize Cenozoic deformation and its consequences, as exemplified by the core complexes in Arizona (Davis & Coney 1979, Coney 1979). This emphasis aroused protest from DeWitt (1980), who cited the evidence for Mesozoic development of the core complexes from California to Canada.

Core complexes are controversial because the evidence for timing of deformation is contradictory. In numerous cases structural elements thought to belong to one deformation episode have been shown to be of different age in nearby areas, and, thus, attempts at simple interpretations have failed. In reality most complexes are multiepisodic, with the effects of each deformation episode heterogeneously distributed rather than coincident. An episode of ductile to brittle Cenozoic deformation is common to all core complexes, but its significance declines northward. Geologists working north of 50° latitude in Canada assign it a very minor role.

The orientations of stress axes have locally remained similar in successive episodes because of a constant disposition of zones of weak and rigid lithosphere—large-scale movements are easiest either toward or away from the unremobilized cratonic lithosphere. Major longitudinal strike-slip motion is not important in core complex geology until northern British Columbia and the Yukon are reached. Those areas are not included in this review.

Structural behavior of the rocks in core complexes varies from ductile to brittle as these thermal and strain-rate controlled structural regimes move through the rocks. Price (1972) emphasized the inherent diachronism in the superposition of structures of different style in an evolving orogenic belt. The general sequence is an evolution from soft sediment structures through ductile structures of succeeding higher metamorphic grades; and then into brittle structures as the rocks undergo accumulation, tectonic burial and magma injection, and unroofing. Regional correlation of locally observed strain episodes is fraught with uncertainty.

I have visited many of the 25 complexes listed by Coney (1980); done geochronometry in the Mojave Desert, Great Basin, Idaho-Washington, and Canadian complexes; and spent seven field seasons mapping part of one complex, the Albion Mountains and Middle Mountain in southern Idaho, that shares features with all of them. No doubt these experiences affected my thinking and are evident in the examples cited and concepts developed herein.

UNIFYING FEATURES

All the metamorphic core complexes of the Cordillera display an association of lithologic, structural, and chronometric features: gently dipping foliation, in many places a distinctly cataclastic foliation; well-developed lineation, ductile and/or cataclastic, that is regionally consistent in orientation, and nearly parallel with minor fold axes; intrusive igneous rocks ranging from

Mesozoic to Cenozoic in age that show a variable structural overprint; Cenozoic K-Ar and FT dates for metasedimentary rocks, even when far from intrusive igneous bodies; and evidence of tectonic denudation along low-angle faults formed in a brittle structural regime that has been clearly superimposed on rocks previously deformed in a ductile fashion. Younger-on-older, nonmetamorphic-on-metamorphic, low-angle tectonic contacts are typical. Brecciation is common, and in its extreme development, sheets of breccia, which are most likely the result of gravity slides, are present.

Metamorphic Fabrics

The gently dipping foliation has been a matter of controversy since the disagreement between Daly (1912) and Gilluly (1934) on its origin in the Shuswap complex. Detailed descriptions of rock fabrics now abound and there is general agreement as to the strain involved: flattening perpendicular to foliation and stretching in the direction of the lineation.

Metamorphic fabrics in the core complexes affected by Mesozoic metamorphism and deformation are too varied to receive detailed discussion in this review. The generalizations about gently dipping, commonly cataclastic foliations apply fairly rigorously to Cenozoic structures. Gently dipping foliations are only part of the Mesozoic structural pattern, which also includes upright foliations, spectacular foliation fans, and varied examples of superimposition of structures. The development of flat foliations in an environment of regional compression and nappe movement is not unreasonable. It could be the result of progressive simple shear where large horizontal transport has occurred; or flattening above upwelling material in gneiss domes; or the axial plane foliation of isoclinal recumbent folds.

Notable detailed studies of metamorphic fabrics are those of Jones (1959), Reesor (1965), Hyndman (1968), Ross (1968), Campbell (1970), Fyles (1970), Fyson (1970), Preto (1970), McMillan (1973), Ross & Christie (1979), and Simony et al (1980) in Canada; Compton et al (1977), Compton (1980), Todd (1980), and Miller (1980, 1982) in the Albion-Raft River-Grouse Creek area of Utah and Idaho; Howard (1968, 1980) and Snoke (1980) in the Ruby Mountains of Nevada; Davis et al (1980) in the Whipple Mountains of California; and Banks (1980), G. A. Davis (1980), and Reynolds & Rehrig (1980) in Arizona. The orientation of the prominent Mesozoic and/or Cenozoic late stretching lineation in core complexes is shown diagramatically in Figure 1.

Igneous Rocks

The intrusive igneous rocks associated with core complexes are moderately variable in composition, mostly from biotite-hornblende granodiorite to biotite-muscovite granite, calcalkaline to mildly alkaline, and met- to peraluminous. They range in size from sills and dikes to batholiths. Their initial strontium isotope ratios are relatively radiogenic (Kistler & Peterman 1978, Armstrong et al 1977, Chase et al 1978, Armstrong 1979) and they are typically aluminous in character (Mursky 1972, Swanberg & Blackwell 1973, Best et al 1974, Keith et al 1980, Miller & Bradfish 1980, Anderson & Rowley 1981); these features distinguish these rocks from the granitic rocks typical of Circumpacific batholiths and provide evidence of an origin involving melting or assimilation of crustal rock, rather than direct magma derivation from a subduction zone. At present levels of exposure the abundance of granitic rock in most core complexes is much less than in the coastal batholith belts, and is not greatly different from that in the intermontane zone separating the metamorphic core zone and batholith belt. The major exceptions to this are in the Okanogan (Fox et al 1976, 1977); Kettle (Cheney 1980), and Valhalla (Reesor 1965) domes where large tongues or sheets of Mesozoic (?) granitic gneiss are incorporated into domed metamorphic rock sequences; the Bitterroot lobe of the Idaho batholith (Hyndman 1980), in which the chemical and structural characteristics of the core complexes are superimposed onto the batholith belt in the one zone where it swings far inland from its usual coastal position; and the South Mountain (Reynolds & Rehrig 1980) and Catalina-Rincon-Tortolita (Banks 1980, Keith et al 1980) complexes of Arizona, which include relatively large amounts of Cenozoic granitic rocks.

As a general rule the peraluminous granitic rocks, often with muscovite and garnet as accessories, are younger than the metaluminous hornblende-bearing granitic rocks located nearby. They also do not fit as neatly into the cycles of waxing and waning Cordilleran igneous activity (Armstrong & Suppe 1973, Armstrong et al 1977, Gabrielse & Reesor 1974) as do the metaluminous granitic rocks whose ages define those cycles. Typical examples are the White Creek (Wanless et al 1968, Mursky 1972) and Galena Bay (Read & Wheeler 1976) plutons in Canada, the Bitterroot (Swanberg & Blackwell 1973, Chase et al 1978) and Almo (Armstrong & Hills 1967) plutons in Idaho, the Kern (Best et al 1974) and Ruby Mountains (Kistler & Willden 1969) plutons in Nevada, the Whipple (Anderson & Rowley 1981) and Coyote Mountains (Wright & Haxel 1980) plutons in California, and the Wilderness (Keith et al 1980) pluton in Arizona. Many of the earliest granitic rocks in the regions containing core complexes are distinctly alkaline (Read 1973, Gabrielse & Reesor 1974, Fox et al 1977, Miller 1978).

The ages of granitic complexes in and near the core complexes vary widely as do their structural settings. In general the time span of ductile deformation overlaps the times of pluton emplacement, and the pre-, syn-, and post-kinematic relationships have been uniquely important in establishing the ages of structures.

Cenozoic Isotopic Dates

The Cenozoic K-Ar and FT dates that characterize core complexes came as a glaring contradiction to the scattered evidence for Mesozoic deformation and the once firmly established view that in the Cordillera the Cenozoic was a time of brittle crustal extension rather than regional metamorphism. The early papers by Armstrong & Hansen (1966), Mauger et al (1968), and Gabrielse & Reesor (1964) drew attention to these young dates, and numerous subsequent studies have closely examined and reconfirmed these observations (Ross 1974, Medford 1975, Archibald et al 1977, Mathews 1981, Armstrong 1974, Miller & Engels 1975, Fox et al 1976, Kistler & O'Niel 1975, Martin et al 1980, Dokka & Lingrey 1979, Banks 1980, Reynolds & Rehrig 1980, Keith et al 1980).

Denudation Faults

Arguments for Cenozoic denudation in the eastern Great Basin were marshalled in a paper by Armstrong (1972). This paper reviewed previous observations of denudation structures on a variety of scales and styles, and pointed out that this aspect of the regional synthesis had not received sufficient emphasis. Since then, studies of denudation have flourished. Coney (1974) reexamined the Snake Range "decollement" and Snoke (1980) the Secret Pass area of the Ruby Mountains and both agreed that Cenozoic denudation was the culminating process in the development of these core complexes. Cheney (1980) and Rhodes & Cheney (1981) described faults bounding the Kettle Dome in northern Washington as denudation faults similar to those observed in the southwestern US. The Newport Fault (Miller 1971), which lies on the Priest River complex, is now viewed by many geologists as an early Cenozoic denudation structure (Cheney 1980, Ewing 1981, Reynolds et al 1981, Price et al 1981a, b).

Spectacular examples of Cenozoic denudation in the Mojave-Sonoran desert region of California-Arizona have been described by Shackelford (1980) and Davis et al (1980), and a recent US Geological Survey conference (Howard et al 1981) gave much attention to these and related structures. The core complex memoir gives detailed descriptions of the denudation complexes in Arizona (Rehrig & Reynolds 1980, Reynolds & Rehrig 1980, Banks 1980, G. H. Davis 1980, Keith et al 1980).

In two areas brittle denudation faults show a special relationship to reset K-Ar dates. In the Snake Range (Lee et al 1970, 1980) and in the Whipple Mountains (Martin et al 1980), the dates decrease upward in lower-plate rocks, reaching minimum values of 17.0 and 15.3 Ma, respectively, for samples close to the master detachment-denudation fault.

SIGNIFICANT CONTRASTS

Most confusion concerning core complexes arises because of differences in age of structures and multiplicity of deformation events. These differences are of two major types:

(*a*) From Canada to California the complexes are polygenetic—all contain evidence of Mesozoic metamorphism and deformation—which is related to Cordilleran orogenic development in a setting of plate convergence. All the complexes have also been overprinted by an episode of crustal extension during the Cenozoic. In Arizona most complexes are monogenetic—exclusively the result of Cenozoic extension.

(*b*) The Cenozoic creation or modification of core complexes in an extensional tectonic setting is the consequence of different episodes of crustal extension on the two sides of the Snake River Plain. North of the plain, the extension is an Eocene event. To the south it is Miocene. Episodes of regional extension during the Mesozoic have not been clearly resolved, and potentially are further complexities of the polygenetic core complexes.

The Shuswap, Albion–Raft River–Grouse Creek, and Whipple-Rawhide-Buckskin areas can be used as examples of polygenetic complexes in order to illustrate the timing and nature of Mesozoic core complexes of the Sevier-Laramide hinterland. The Catalina-Rincon-Tortolita complex exemplifies the monogenetic core complexes of Arizona.

Shuswap Complex

Understanding of this large and only partly well-exposed region has come about gradually as detailed mapping and fossil and isotope geochronology have proceeded. Major issues remain in flux but a general outline (Brown et al 1981, Price et al 1981b) is possible. The premetamorphic protoliths that compose the complex include early Precambrian crystalline basement (Chamberlain et al 1979, Brown 1980); a middle Proterozoic rift basin filled with clastic strata (Gabrielse 1972); late Proterozoic through Paleozoic to Triassic sediments, mostly types deposited outboard of the carbonate bank (Read & Wheeler 1976); and allochthonous Paleozoic to Triassic oceanic rocks—including a Devonian volcanic arc assemblage with granitic batholiths (Okulitch 1979)—emplaced tectonically in late Triassic or early Jurassic time (Rees 1981, Parrish 1981). Tectonic imbrication and ductile deformation had affected parts of the protolith assemblage, but no detailed reconstruction of this structural stage is yet possible (Read & Wheeler 1976, Brown 1978).

Igneous activity began anew in late Triassic–early Jurassic time and by the end of the Jurassic, numerous large plutons were emplaced (Gabrielse & Reesor 1974, Pigage 1977). These postdate the main structural development of the complex—an aggregation of Alpine-style nappes with associated metamorphic fabrics (Morrison 1980, Read & Klepacki 1981). The metamorphic culmination in many areas was reached about 180 to 160 Ma ago, a conclusion well documented by batholith geochronometry circumscribing the complex.

Pluton emplacement and movement along major faults continued on into the Cretaceous as the nappe-batholith complex was gradually unroofed (Price et al 1981b, Archibald et al 1977, Brown 1978, Brown et al 1981). At the same time, the entire complex was translated more than 100 kilometers toward cratonic North America while the Rocky Mountain fold and thrust belt developed (Price & Mountjoy 1970). Evidence for Cretaceous ductile deformation is not yet well documented. Doming is a relatively late and modest modification of the nappes that is poorly dated (Read & Klepacki 1981). Some doming may be as old as Jurassic; the final stages were as recent as Eocene.

In the Eocene, after Rocky Mountain thrusting had ceased, an episode of crustal heating and extension further modified the metamorphic complex (Price 1979, Price et al 1981a, Ewing 1981). Igneous activity, extension and denudation structures, and dome accentuation occurred while the crystalline rocks were still at or being raised to, temperatures of 300° to 400°C. This dramatic episode terminated abruptly about 45 Ma ago. Over a large area near coincident with sillimanite-grade rocks, K-Ar and FT dates of minerals were set as isotherms moved rapidly down into the crust. Ductile to cataclastic foliation of Eocene age has been proven in a few Eocene igneous rocks (Ross 1974, Solberg 1976) and may be widespread in the reset K-Ar date terrane. Mylonitization associated with low-angle faults and low-angle faults that juxtapose rocks of contrasting metamorphic grade have been observed on both sides of the complex. However, these features are not as spectacularly developed as in complexes further south and their age is not firmly established (Ross 1973, Brown et al 1981, Murphy 1981, Lane 1981).

Albion–Raft River–Grouse Creek Complex

The chronology of metamorphism and deformation in this area has been resolved as the result of 15 years of study by myself (Armstrong 1968a, 1970, 1976, in preparation, Armstrong & Hills 1967), Compton & Todd (Compton 1980, Compton et al 1977, Compton & Todd 1979, Todd 1980), and Miller (1980, 1982).

The protolith assemblage was early Precambrian crystalline basement with a Precambrian (?) and Paleozoic to Triassic cover sequence that is related to the continental margin miogeosyncline, but thinner and more pelitic; and an allochthonous quartzite-dominated later Precambrian or Paleozoic sedimentary sequence, present only in northern parts of the complex. During the Paleozoic the autochthon may have been a slowly subsiding outer high (Schuepbach & Vail 1980) developed on a late Precambrian divergent continental margin—outboard of the shallow-water carbonate bank. Further offshore in a rise or slope environment lay the depositional environment of the quartzite assemblage.

Regional metamorphism reached a first amphibolite facies culmination in Jurassic time, more than 160 Ma ago. This was followed by a Cretaceous period of imbrication and ductile deformation under conditions of decreasing pressure and temperature. Cretaceous biotite-muscovite granites were emplaced in northwestern parts of the complex.

Profound further modification of the area took place in early to mid-Cenozoic time, accompanied by further emplacement of granitic rock. Near and in the granitic injection complex, temperatures rose to sillimanite grade, and rock strength was dramatically reduced while the area was subject to regional extension. The several basement-cored domes may have risen at this time, but this is not well dated and their growth may be largely Cretaecous. Both interference structure (Armstrong in preparation) and polydiapiric (Miller 1982) explanations for the domes have been advocated. As a result of high temperatures and extreme extension, the nappes were reactivated and thinned while in the Grouse Creek Mountains, near a Cenozoic pluton, the basement became locally remobilized and intruded its sedimentary cover. A strong mylonitic foliation and lineation, with lineation trend nearly parallel to Cretaceous fold axes and lineations, developed in areas deformed at this time. The effects of the Cenozoic episode dominate the structural story in Utah, but only affect the westernmost part of the Albion Mountains. As a result, the distinction of Mesozoic and Cenozoic structures is possible there. Relatively late features are the development of brittle denudation and detachment faults, breccia sheets, and far-traveled gravity slide masses. A remarkably similar overall structural chronology has been worked out in the Ruby Mountains-Wood Hills-Pequop Mountains area (Thorman 1970, Howard 1980, Snoke 1980) and probably also applies to the Snake Range (Misch & Hazzard 1962, Nelson 1969, Coney 1974).

Whipple-Rawhide-Buckskin Mountains

The spectacular denudation structure of the Whipple Mountains and the structural history of the rocks exposed by denudation have been the subject of recent intensive study (Davis et al 1980, Dickey et al 1980, Martin et al 1980, Anderson & Rowley 1981). There, a middle Precambrian basement and scattered remnants of its late Proterozoic and Paleozoic platformal cover have been subject to Jurassic metamorphism and Jurassic and Cretaceous intrusion of granitic rocks. During later Cretaceous time, structurally deeper rocks were intensely mylonitized—with the development of cataclastic foliation and lineation synchronous with the emplacement of granitic sheets parallel to foliation.

Distinctly later, in the Miocene, brittle detachment faulting removed much of a Cenozoic volcanic and sedimentary cover. Slickensides and structures in the disrupted cover indicate that the movement was approximately parallel to the Mesozoic lineation, tempting geologists to find a common genesis for both (Lucchitta et al 1981) in spite of the considerable difference in age.

Catalina-Rincon-Tortolita Complex

The geology of this long-studied area was recently synthesized by G. H. Davis (1980), Banks (1980), and Keith et al (1980). The core complex protolith was middle Precambrian crystalline basement with platform cover of late Precambrian to Mesozoic age. The peripheral effects of latest Cretaceous and early Cenozoic Laramide orogeny (Drewes 1978, Davis 1979) may have been present in the form of detached and deformed sedimentary cover and brittle-fault bounded basement-cored uplifts, but no regional metamorphism had affected the area before the Cenozoic.

Igneous activity started in Laramide time with the emplacement of 75 to 60 Ma old calc-alkaline granodiorite. This was followed by a large laccolithic body of peralkaline granite of Eocene (50 to 44 Ma) age and a final calc-alkaline intrusive of Miocene (29 to 25 Ma) age. Cataclastic flattening and stretching occurred after each of these three intrusive episodes but were largely confined to the time interval from late in the second to late in the third igneous episode. Uplift and rapid cooling, with setting of K-Ar and FT mineral dates, came between 30 and 20 Ma, outlasting the last igneous episode.

During the height of deformation the granites and enclosing country rock were partially converted to mylonitic augen gneiss, with gently dipping foliation and regionally consistent lineation. Stretching also produced ductile normal faults, and flattening created a variety of recumbent folds. High-temperature extension was superseded by brittle denudation, with unmetamorphosed cover moving radially off rising domes along low-angle faults (Davis 1975). The metamorphic cores now stand fringed with chaotic, broken, and folded bits of unmetamorphosed cover that lie in brittle-fault contact with deformed basement, its pre-Cenozoic sediment cover, and intrusive granitic rocks. Davis & Coney (1979) describe the overall structural pattern as one of "megaboundins," with deformation localized at low levels in the Phanerozoic sediment cover where granite spreading was concentrated. This same zone was one of steep metamorphic gradients during early to mid-Cenozoic time.

GEOPHYSICAL EXPRESSION

Metamorphic core complexes do not stand out on geophysical maps of the Cordillera. Seismicity and heat flow reflect late Cenozoic tectonic activity and thus show absolutely no association with metamorphic exposures (Eaton 1980). There is no expression of individual complexes or domes on magnetic anomaly maps, but the complexes all occur in a broad belt of low magnetic relief that is caused by thick sedimentary cover and high crustal temperatures (Berry et al 1971, Mabey et al 1978). The gravity field in areas of the complexes is intermediate (Bouguer anomaly -90 to -160 milligals). Crustal thickness ranges from 20 to 45 km. It is typically about 25 km under Mojave and Arizona complexes; 30 km under Nevada, Utah, and Idaho complexes; and 40 km under complexes in southern Canada (Eaton et al 1978, Smith 1978, Forsyth et al 1974, Berry & Mair 1977, Cumming et al 1979). Late Cenozoic crustal extension can explain the steady southward decrease in crustal thickness. Cady (1980) observed that local gravity highs (10 to 20 milligals) correlated with individual gneiss domes in southern Canada and northern Washington, where Cenozoic modifications of the gravity field are minimal. In the metamorphic core zones of southern Canada and northern Washington, crustal seismic velocities and densities are generally higher than in surrounding areas (Cady 1980, Cumming et al 1979) and no crustal roots are present. The thickening of "lower crustal" rocks in the core zones is consistent with the idea of compressive crustal thickening during Mesozoic orogeny followed by deep erosion and isostatic recovery (Cady 1980).

CRUSTAL EXTENSION

Regional extension is a widely accepted explanation for much of the character of core complexes, particularly for those features acquired during Cenozoic time (Coney 1979, 1980, G. A. Davis 1980). The timing of extension seems to be directly related to concurrent igneous activity: Eocene in Canada and the northwestern US (Lipman et al 1972, Armstrong 1974, Ewing 1981), and Oligocene-Miocene south of the Snake River Plain (Christiansen & Lipman 1972, Snyder et al 1976). Most authors view the extension process as one where shallow rocks undergo brittle fracture and block rotation due to movement along listric normal faults, while ductile necking of warm lower crust occurs simultaneously (Hamilton & Myers 1966, Stewart 1978, Coney 1979, Eaton 1980, G. A. Davis 1980). A counter-proposal (Wernicke 1981a, b) is that low-angle normal faults continue through the crust to root in the mantle. This solves the problem of finding a stretched terrane synchronous in movement and complementary to each low-angle denudation structure, a difficulty that forced some authors to attempt a surficial, gravity-glide explanation for large-scale detachment (Davis et al 1980). A complete description of crustal extension probably incorporates both propositions. The depth to the brittle-ductile transition is clearly variable and under some circumstances may even lie below the crust, but in others it has been within supracrustal strata, as for example in the Catalina-Rincon-Tortolita complex. Gravity gliding cannot be denied for situations where detached rocks have moved across Cenozoic deposits in basins, as in the Rawhide-Buckskin area (Davis et al 1980) or Grouse Creek Mountains (Todd 1980).

The logical structural succession for rocks that have moved from a metamorphic environment to surface conditions is the order observed in core complexes. Ductile recrystallization-flow structures are succeeded by cataclastic flow, then brittle cataclasis, and finally overprinted by local gravity-glide structures, including the development of breccia sheets. This sequence could develop in one protracted extension episode as an overlapping continuum, while rocks from a depth of several kilometers are tectonically denuded and brought to the surface (as in the Catalina-Rincon-Tortolita complex); or it could develop in distinct stages—Mesozoic ductile or cataclastic fabrics overprinted by Cenozoic ductile and brittle faulting (as in the Shuswap, Albion–Raft River–Grouse Creek, and Whipple complexes). Some low-angle detachment surfaces show large-scale corrugations that parallel the extension direction inferred from other structural data (Rehring & Reynolds 1980, Cameron et al 1981).

The most popular plate tectonic explanations for the two Cenozoic episodes of crustal extension are oblique rifting due to a diffuse transform interaction of North America and offshore lithosphere plates (Atwater 1970, Ewing 1981, Price 1979, Price et al 1981a); and back-arc spreading or spreading over a hole in the descending slab due to upwelling asthenosphere (Scholz et al 1971, Dickinson & Snyder 1979, G. A. Davis 1980). These proposals are not mutually exclusive. They are different views of the same process, and they involve an interplay of globally created stress fields, mantle flow patterns, magma generation, and consequent failure of plates as they are weakened by heating.

LOCALIZATION OF THE MESOZOIC METAMORPHIC CORE ZONE

Crustal extension does not explain the restricted geographic location of the metamorphic core complexes. Something more is required, and in the case of the polygenetic complexes of the Sevier-Laramide hinterland, it is almost certainly the presence of the thickened and metamorphosed sialic crust that was generated during Mesozoic orogeny. Superimposition of nonuniform crustal heating and stretching on irregularly thickened and heterogeneous crust could explain the irregular distribution of US core complexes. The alignment of Cenozoic core complexes across Arizona may represent the coincidence of magmatism and heating of sialic crust with regional extension—as in the "megaboudin" model of Davis & Coney (1979).

Dual Thermal Culmination

The formation of Mesozoic metamorphic infrastructural zones in a belt that lies inland from contemporaneous coastal batholith belts requires explanation.

The usual "Cordilleran" or "Andean" model has only one broad magmatic arc—thermal culmination (Burchfiel & Davis 1975, Dickinson 1976). The locus of Mesozoic infrastructural zones is clearly a thermal culmination that is spatially separate from the culmination associated with the batholith belts; the Idaho batholith is the only significant exception. The dual thermal culmination has always been evident in Canada (Monger & Hutchison 1971), and the distribution of the polygenetic core complexes in the US indicates that it is present there as well.

Study of active subduction zones may provide an explanation for dual thermal culminations and thereby explain the Mesozoic thermal pattern of the Cordillera. In an active island arc like Japan (Sugimura & Uyeda 1973) volcanic activity is sharply focused in a narrow belt with a well-defined cutoff toward the trench. This sharp boundary is called the *volcanic front*. In the opposite direction there is an exponential decline over a distance of several hundred kilometers in the production rate of volcanic rocks. This is analogous to the distribution of granitic rocks of any given age in Cordilleran batholith belts. Heat carried into the crust by magma is an explanation for one thermal culmination.

For hundreds of kilometers behind the zone of maximum magma production, a broad zone of high heat flow is present (Sugimura & Uyeda 1973). It is largely coincident with scattered volcanic activity, but the observed heat flux is not correlated with volcanic production rates. A widely accepted explanation for this zone of high heat flow follows from the original suggestions of McKenzie (1969) and Griggs (1972) that the descending slab induces flow of mantle material in the triangular prism of the asthenosphere bounded by floating and sinking lithosphere slabs (Andrews & Sleep 1974, Anderson et al 1976, Toksöz & Bird 1977a, b, Uyeda 1977). Numerical models of Andrews & Sleep (1974) illustrate the consequences of this type of process (Figure 2). As subduction continues from an initial condition of horizontal isotherms, a broad band of the floating lithosphere undergoes heating from below. In the model illustrated, the thermal culmination, which is purely a consequence of flow in the mantle and stretching of the overlying lithosphere, is localized about 300 km from the earth surface point of contact of the plates. This is well behind the expected locus of the magmatic culmination, which would be about 150 km from the same point of contact. Thus there is a possible explanation for two thermal maxima inland from an active subduction zone—one due to magmatic heat transfer and the other due to induced mantle flow. The second culmination could be augmented by magmatic heat transfer, as rising mantle and heated crust undergo partial melting, and by heat generation in sialic basement of the continental margin.

Crustal Shortening

Where the lithosphere plate floating above an active subduction zone is subjected to heating, it will weaken, and may fail, by shortening or elongation. Asthenosphere flow merely provides favorable conditions for lithosphere failure behind a magmatic arc. Integrated global stress patterns dictate whether the failure in a given area will be shortening or elongation. Geologic evidence indicates consistent trench-craton convergence during Mesozoic time along the Pacific Margin of North America (Hamilton 1978); this convergence is accepted here as an observation without explanation.

The detailed structural style of crustal shortening can be exceedingly varied. The heterogeneity of the crust and variable position of the brittle-ductile transition within the crust will both be critical factors in control of structural style, and these will change as deformation proceeds. Figure 3 illustrates three of the many possible patterns that can occur where crystalline basement underlies a stratified cover. If the basement is fairly rigid, large thrust structures may form. A shallow-basement example of this is the Wind River thrust in Wyoming (Smithson et al 1978). Deeply buried examples of large thrust

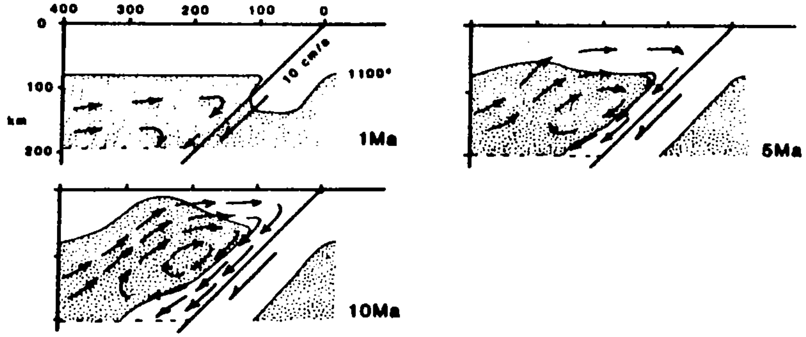

Figure 2 Numerical model of mantle flow and temperature changes induced by a downgoing slab (Andrews & Sleep 1974). The 1100°C isotherm that is shown is approximately the lithosphere-asthenosphere boundary. The scaling is only approximate and will vary as subduction rate, subduction angle, and physical properties of the mantle and lithosphere are changed. Nevertheless, the deduction, based on reasonable choices of physical properties, is that induced flow will occur, and this will change the thermal structure of a broad region above the subducting plate. An exact correspondence with Cordilleran tectonics during the Mesozoic is not expected because the calculation was intended to simulate the formation of marginal basins. The models of Toksoz & Bird (1977a, 1977b) for continental plateaus are similar and equally applicable to the discussion. Numerical models designed to simulate induced flow and thermal evolution of the Cordillera are needed. In these models the radiogenic heat production of the crystalline basement may need to be taken into account to produce the metamorphic core complex thermal culmination.

steps may include the Snake Range and Albion–Raft River–Grouse Creek core complexes. A slightly more ductile basement may fail in thin slices or wedges. This style is observed in the massifs of the Alps (Labhart 1966, 1968, Steck 1968) and is described in the Shuswap complex (Morrison 1980) and Wasatch Mountains of northern Utah (Bruhn & Beck 1981). Basement-cored fold nappes are formed where basement becomes very ductile, and where there is negligible ductility contrast between basement and cover. This is seen in the Pennine zone of the Alps (Milnes 1974) and in the Shuswap complex (Brown et al 1981, Read & Klepacki 1981). In both areas are examples where early basement wedges were overprinted by more ductile structures as metamorphic grade increased (Milnes 1974, Ross 1968, Morrison 1980, 1981).

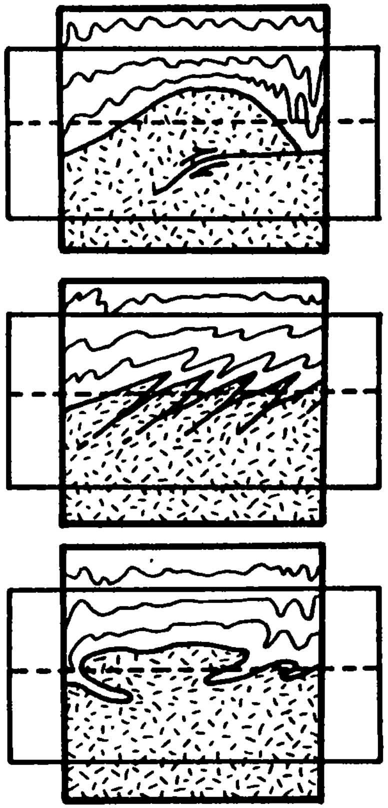

Figure 3 Three styles of crustal shortening, shown very diagramatically. A large thrust involving basement results in considerable structural relief on the basement-cover contact. Imbrication of thin basement wedges and slices or ductile folding together of basement and cover can accomplish similar amounts of shortening but produce less structural relief.

The different structural styles all accomplish crustal shortening, even; though their surface expression may vary dramatically, all presumably played a role during Mesozoic crustal shortening in the Cordillera.

Where thick geosynclinal sediments are involved, the surface expression of crustal shortening need not directly overlie the location of basement shortening. Cordilleran thin-skinned fold and thrust belts are a surficial expression of crustal shortening in basement farther west, including the rocks within or beneath metamorphic core complexes (Royse et al 1975, Price et al 1981b). The area of shortened and thickened crust provides the slope and push to drive supracrustal rocks toward fold and thrust belts (Kehle 1970, Price 1973, Elliott 1976, Chapple 1978).

In the Albion Mountains, telescoping of stratigraphic units appears to have gone on during fold and thrust belt deformation, whereas in the Shuswap complex much of the visible structure predates deformation along the eastern edge of the deformed belt (Brown 1978, Brown et al 1981, Price et al 1981b). Both situations are possible. As the core zone is thrust eastward, climbs thrust ramps, and is exposed by erosion from above, the brittle-ductile transition will move downward, and the zone where deep accommodation of surface shortening occurs will sink deeper into the crust and shift eastward. Earlier episodes of deformation will be frozen into the rocks that are now exposed at the surface, while shortening is still in progress at depth. During later stages of orogenesis the active thrusts will have shifted deeper and eastward, so that the upper levels of the metamorphic core will be passively transported cratonward. This view of upwelling and structural diachronism explains the seeming contradiction that metamorphic core ductile structures predate the later stages of marginal fold and thrust belt deformation.

The 100 to 200 km of crustal shortening observed in supracrustal rocks is compatible with the amount of sialic basement aggregated in Cordilleran metamorphic core zones—if two factors are taken into account (Brown 1978, Price et al 1981b). The first is that at the end of geosyncline subsidence, the sialic basement in polygenetic core complex areas was probably thin, attenuated by faulting and ductile stretching, and partially replaced by sima, as is now the situation along divergent margins. The second factor is Cenozoic extension, which may have removed any thickened crustal roots that were present at the end of the Mesozoic. The present exposed area of polygenetic core complex rocks in any regional cross section is reasonably well correlated with geologic estimates for the amount of shortening in the corresponding part of the fold and thrust belt.

Uplifted and Dejected Belts

Where a thick suprastructure is still present in the hinterland of the Cordilleran fold and thrust belt, there may be ductile folding or multiple thin-slice basement wedging at depth, with little surface expression. In contrast, large thrust steps are expressed in a particular pattern that is observed in the Snake Range and Albion–Raft River–Grouse Creek areas. This pattern has been called an uplifted infrastructure-dejected belt couple (Armstrong 1978). A geometric clue to a large thrust step is provided by the large local structural relief. In both cases the arched ranges, exposing metamorphic infrastructure, are paired with exceptionally deep structural depressions, the dejected belts, that lie immediately to the east. In the case of the Snake Range the depression is the Confusion Range synclinorium (Hose 1977); in that of the Albion–Raft River–Grouse Creek area, it is the Sublett synclinorium (Armstrong et al 1978). Both depressions contain rocks as young as Triassic and are the only places where rocks that young are preserved between the metamorphic core and the fold and thrust belt. The inferred deep structure from the Snake Range to the fold and thrust belt is shown in Figure 4. A similar geometric logic has been used in interpreting deep structure in the metamorphic core of the Canadian Rocky Mountains (Simony et al 1981, Price & Fermor 1981, Price et al 1981b), following the rules for balanced sections discussed by Dahlstrom (1970).

CONCLUSION

There are a variety of processes and events responsible for Cordilleran metamorphic core complexes. The polygenetic complexes originated as metamorphic infrastructure in the hinterland of the Cordilleran fold and thrust belt as a result of lithosphere heating above a long-lived subduction zone. Their origin was also partly a consequence of mid-Mesozoic imbrication of oceanic terranes and the craton margin. Crustal shortening and thickening in the metamorphic core zone produced a large volume of deformed rock and were responsible for bringing rocks from depths of 10 to 20 km to surface exposure. Isostatic response to crustal thickening and ramping up large thrust steps were parts of the uplift process. The irregular exposure of US core complexes is explained by the variable structural style of crustal shortening and different depths of erosion through the suprastructure. At the end of the Mesozoic the metamorphic core rocks that we now see at the surface were within a few kilometers of the surface and were probably still warmer, more buoyant, and more ductile than surrounding rocks, which had not undergone such dramatic excursions into ductile structural regimes.

Cenozoic volcanism and upwelling of hot asthenosphere again led to increased lithosphere temperatures and structural failure, this time extensional. The Mesozoic infrastructure was remobilized, and accidents of block faulting and tectonic denudation, following irregularly distributed deep-seated crustal stretching, led to exposure of the polygenetic core complexes as we see them

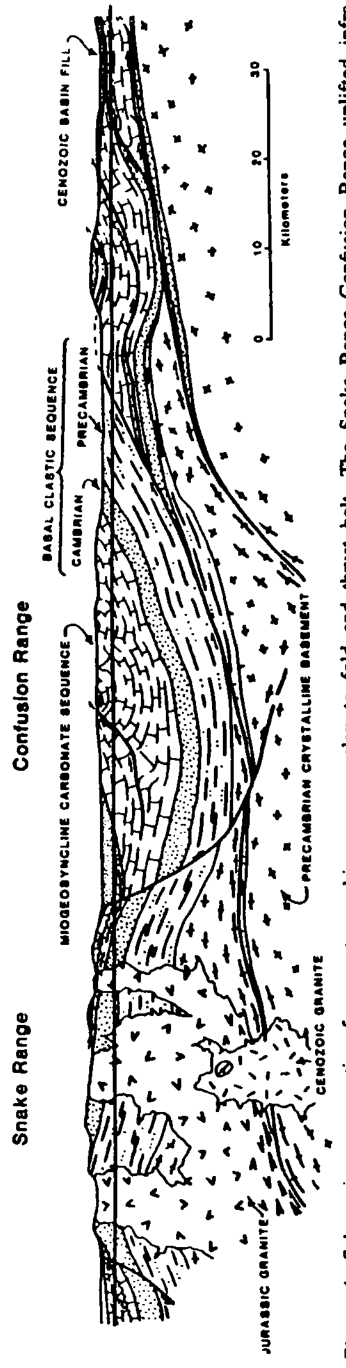

Figure 4 Schematic cross section from metamorphic core complex to fold and thrust belt. The Snake Range–Confusion Range uplifted infrastructure–dejected belt couple are shown as an example of structural relief produced by a thrust step involving crystalline basement. This section is similar to one drawn by Coney (1979, 1980), based on an earlier shallow section published by Armstrong (1972).

today. Gravitational glide structures were formed as a consequence of topographic relief on uplifted blocks. The same back-arc weakening of lithosphere above a descending plate that operated during Mesozoic time may have had a role in localizing the Cenozoic core complexes.

In Arizona during mid-Cenozoic time, magmatism and lithosphere failure encroached cratonward, even beyond the Mesozoic orogenic front—something that did not occur farther north in the Cordillera. As a consequence of this stepping forward of back-arc thermal effects, a special type of core complex—the monogenetic "megaboudin"—was created. The continuous zone of Cenozoic lithosphere failure, with its varied geological consequences, links the two core complex types.

A core complex can be described in a few words: it is an exposure of rocks that were once ductile lower crust, on which shallow brittle extensional features have been superimposed. In these special areas we have an opportunity to view the myriad effects and complexities of lithosphere failure. No simple single prescription of events and processes can explain all the complexes. Each has its own unique protolith and deformation history and provides a new geological puzzle.

There are lessons to be learned in the study of Cordilleran core complexes that are applicable to the study of metamorphic core zones in other Pacific-type orogenic belts. In addition, our understanding of the process of crustal extension, such as has occurred whenever divergent continental margins have been formed on the site of waning orogenic belts, is enriched.

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MID-OCEAN RIDGES:
Fine Scale Tectonic, Volcanic and Hydrothermal Processes Within the Plate Boundary Zone

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INTRODUCTION

A first order model of spreading centers as idealized linear boundaries of crustal and lithospheric generation provides only a gross understanding of global scale plate kinematics. As we attempt to understand the complexity of crustal and lithospheric structure of two thirds of the earth's surface, it is becoming increasingly necessary to study the tectonic, volcanic, and hydrothermal processes within the spreading center plate boundary zone. All oceanic crust bears the imprint of these processes. This review focuses on a few selected topics concerning the fine scale tectonics and geophysics of the active axial zone of mid-ocean ridges with reference to associated volcanic and hydrothermal processes. It draws heavily on recent studies that use deeply towed instrument packages, multi-beam bathymetric mapping, ocean bottom instruments, and ALVIN (e.g. the Famous, AMAR, RISE, and Galapagos expeditions).

We begin with a review of the large-scale structure of spreading centers. We then take a close look at the axial neovolcanic zone and progress away from the axis through the active tectonic zones. Next we consider the characteristics of the axial magma chamber and associated hydrothermal actvity, as well as the generation of magnetic anomaly "stripes" and their implications for crustal generation. One of our findings is that the initial two-dimensional model of volcanic and tectonic zones must be expanded upon to allow for variations

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