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Abstract

The deep crustal rocks exposed in the Ruby-East Humboldt metamorphic core complex, northeastern Nevada, provide a guide for reconstructing Eocene crustal structure ~50 km to the west near the Carlin trend of gold deposits. The deep crustal rocks, in the footwall of a west-dipping normal-sense shear system, may have underlain the Piñon and Adobe Ranges about 50 km to the west before Tertiary extension, close to or under part of the Carlin trend. Eocene lakes formed on the hanging wall of the fault system during an early phase of extension and may have been linked to a fluid reservoir for hydrothermal circulation. The magnitude and timing of Paleogene extension remain indistinct, but dikes and tilt axes in the upper crust indicate that spreading was east-west to northwest-southeast, perpendicular to a Paleozoic and Mesozoic orogen that the spreading overprinted. High geothermal gradients associated with Eocene or older crustal thinning may have contributed to hydrothermal circulation in the upper crust. Late Eocene eruptions, upper crustal dike intrusion, and gold mineralization approximately coincided temporally with deep intrusion of Eocene sills of granite and quartz diorite and shallow intrusion of the Harrison Pass pluton into the core-complex rocks.

Stacked Mesozoic nappes of metamorphosed Paleozoic and Precambrian rocks in the core complex lay at least 13 to 20 km deep in Eocene time, on the basis of geobarometry studies. In the northern part of the complex, the presently exposed rocks had been even deeper in the late Mesozoic, to >30 km depths, before losing part of their cover by Eocene time. Nappes in the core plunge northward beneath the originally thicker Mesozoic tectonic cover in the north part of the core complex. Mesozoic nappes and tectonic wedging likely occupied the thickened midlevel crustal section between the deep crustal core-complex intrusions and nappes and the overlying upper crust. These structures, as well as the subsequent large-displacement Eocene extensional faulting and flow in the deep crust, would be expected to blur the expression of any regional structural roots that could correlate with mineral belts. Structural mismatch of the mineralized upper crust and the tectonically complex middle crust suggests that the Carlin trend relates not to subjacent deeply penetrating root structures but to favorable upper crustal host rocks aligned within a relatively coherent regional block of upper crust.

Introduction

The complex history of Cenozoic extensional faulting in the Basin and Range province has dismembered older structures, but in doing so has exposed different parts of preexisting overthickened crustal sections. Metamorphic core complexes (Fig. 1), because they juxtapose upper crustal rocks against deeper rocks unroofed by extensional processes, offer opportunities to investigate and integrate structure and processes of the deep and shallow crust. Of special interest are the conditions below areas such as the Elko-Carlin area of northeastern Nevada, with its oil and gas accumulations (Garside et al., 1988; Flanigan et al., 1990) and Carlin-type gold deposits (Ilchik and Barton, 1997; Hofstra and Cline, 2000). The nearby Ruby-East Humboldt metamorphic core complex affords a view into crustal levels that were at one time as deep as ≥30 km and may shed light on the structural foundation below the Elko-Carlin area at the time of Carlin-type gold mineralization.

Carlin-type gold deposits are epigenetic, disseminated, auriferous pyrite deposits, typically hosted in calcareous sedimentary rocks (Hofstra and Cline, 2000). Along the northwest-striking Carlin gold trend (Fig. 2), the deposits formed at crustal depths of about 1 to 5 km (Kuehn and Rose, 1995; Henry and Ressel, 2000; Hofstra and Cline, 2000). Mineral deposits at several sites have been dated or tied chronologically to intrusions within a few million years of 40 Ma, late Eocene (Seedorff, 1991; Henry and Boden, 1998; Hofstra et al., 1999, 2000; Chakurian et al., 2000; Henry and Ressel, 2000; Ressel et al., 2000; Arehart et al., 2003).

The deep crustal rocks in the Ruby-East Humboldt core complex have been proposed to have structurally underlain upper crustal rocks that now lie tens of kilometers to the west before late Cenozoic crustal extension (Jansma and Speed, 1990; Howard, 1992; Newman and Witter, 1992). In order to...
explore deep crustal conditions and structure in late Eocene time as part of a crustal framework for Carlin-type mineralization, this paper evaluates such a reconstruction and compares tectonic and magmatic styles at 40 ± 5 Ma at deep and shallow crustal levels. The reconstruction offered here interprets the Eocene middle crust as tectonically shuffled as a result of contractile crustal thickening in the Mesozoic followed by partial thinning before, during, and since the episode of Carlin-type gold mineralization. This structural complexity implies that no simple middle crustal structure or rock body formed or forms a root for the Carlin mineral trend.

However, Eocene crustal processes in the region included magmatic intrusion, normal faulting, a geothermal gradient enhanced by crustal thickening, and ponding of surface water in active structural basins. These processes may have helped to drive Eocene hydrothermal systems and to trigger Carlin-type mineralization in favorable host rocks. A favorable host facies may occupy a relatively intact upper crustal block detached from its original crustal roots.

This paper begins with a brief review of the pre-Eocene framework of the region and then attempts to integrate an Eocene regional picture. Appraisal of the former depth of the deep crustal rocks and their relative displacement points to a possible restoration of deep crustal and shallow crustal structures. An integrated geologic cross section presented for the upper ~20 km of the Eocene crust raises the question of pre-extension Mesozoic structures that helped bury the metamorphic rocks. Scrutiny of the upper crust's regional Eocene setting and evidence for deep crustal Eocene magmatic and tectonic processes leads to a brief discussion of implications that the deep crustal evidence may have for processes affecting the upper crust, including Carlin-type gold mineralization.

Pre-Eocene Tectonic Framework of Northeastern Nevada

Northeastern Nevada (Fig. 1) is broken into many fault blocks of the Basin and Range province but can be divided for this paper into three tectonic domains (Figs. 2, 3). Armstrong
Mylonitization caused by normal-sense shearing roofed the metamorphic rocks during Tertiary time (Snoke et al., 2000). Extensional tectonic processes unroofed the metamorphic core complex, allochthons of western-facies Paleozoic rocks near the top of the upper crustal section in the Elko-Carlin domain, and deeper inferred thrust duplications of miogeoclinal and basement rocks concealed under the Elko-Carlin domain. Older features include abundant Mesozoic granite (crosses) under the metamorphic core complex, allochthons of western-facies Paleozoic rocks near the top of the upper crustal section in the Elko-Carlin domain, and deeper inferred thrust duplications of miogeoclinal and basement rocks concealed under the Elko-Carlin domain. CR = Cortez Range; PR = Piñon Range; RM = Ruby Mountains; SM = Spruce Mountain and southern Pequop Mountains.

The Elko-Carlin domain forms a western region where pre-Cenozoic thrust sheets of deep-water, western-facies Paleozoic strata overrode the miogeoclinal rocks. Lead and strontium isotopic ratios indicative of source regions for plutonic rocks signal that concealed North American basement thins westward beneath this domain (Tosdal et al., 2000). This domain hosts the Carlin-type gold deposits of the Carlin trend and the Independence district (Fig. 2). The thrust faults and associated middle to upper Paleozoic orogenic and overlap strata resulted from a series of shortening events along the western margin of North America, including the Antler (Devonian-Mississippian), Sonoma (Permian-Triassic), Elko (Jurassic), and Sevier (Cretaceous and early Tertiary) orogenicies. By the time of final marine deposition in the Triassic, the shortening events apparently thickened the crust sufficiently that the entire region subsequently emerged as a persistent highland. All younger strata in northeastern Nevada are non-marine. Mesozoic thrust faults then reshuffled Paleozoic thrust sheets, complicating the dating of specific thrust movements and determination of the relative importance of Paleozoic and Mesozoic thrusting events (Coats and Riva, 1983; Jansma and Speed, 1990; Ketner, 1998; Taylor et al., 2000).
Backthrusting in the Elko-Carlin domain is suggested by Paleozoic(?) northward-overturned folds in the Bull Run Mountains (Ehman, 1985), Cretaceous(?) northwest-directed thrusting proposed near Carlin (Jansma and Speed, 1990), and mid-Mesozoic westward-overturned folds and thrusts in the Piñon and Cortez Ranges (Smith and Ketner, 1977; interpreted as part Eocene by Ketner and Alpha, 1992). In the core complex domain, the premetamorphic (pre-Cretaceous[?]) west-directed Ogilvie thrust in the Ruby Mountains (Fig. 4; Howard et al., 1979), Mesozoic northwestward-overturned folds in the Wood Hills (Thorman, 1970), and a reverse fault in the Pequop Mountains (Camilleri and Chamberlain, 1997) all suggest backthrusting.

Mesozoic magmatism in northeastern Nevada resulted in numerous Jurassic and less abundant Early Cretaceous intrusions, a 2.7-km thickness of Upper Jurassic volcanic rocks in the Cortez Range near the west side of the area of Figures 2 and 3, and Late Cretaceous granites that are voluminous in the deep crustal metamorphic complex but scarce at upper crustal levels (Muffler, 1964; Coats, 1987; Elison, 1995; Lee and Barnes, 1997; Snoke et al., 1997; McGrew et al., 2000; Mortensen et al., 2000).

**Eocene Position of Deep Crustal Rocks Exposed in the Core Complex**

*Paleodepth*

Geochronology and geobarometric studies of mineral assemblages in the Ruby-East Humboldt metamorphic core complex indicate Mesozoic structural burial to depths two to three times the stratigraphic thickness, metamorphism, and Cenozoic tectonic exhumation (Hodges et al., 1992; McGrew and Snee, 1994; Camilleri and Chamberlain, 1997; McGrew et al., 2000; Table 1). The history of burial and unroofing are central to reconstructing the Eocene crustal section. A west-dipping Tertiary shear zone, exposed as a mylonitic carapace (0.03–2 km thick) forming the western side of the Ruby Mountains and East Humboldt Range (Fig. 4), overprinted
the metamorphic rocks during extensional exhumation (Snoke and Lush, 1984; Wright and Snoke, 1993; Mueller and Snoke, 1993a).

Paleodepths of the metamorphic complex compiled from mineral-assemblage geobarometry studies (Table 1) show that maximum calculated pressures and depths of burial increase northward in the complex (Fig. 5). Pressures up to 9 kbars corresponding to depths >30 km, calculated for prograde metamorphic mineral assemblages in Neoproterozoic miogeoclinal strata in the East Humboldt Range and Clover Hill, are two to three times the load of stratigraphic burial (McGrew et al., 2000). The amount of tectonic burial and subsequent denudation exceeds that of most other Cordilleran core metamorphic complexes. The overload is thought to apply to times of maximum crustal thickening in the Late Cretaceous or earlier (Hodges et al., 1992; Camilleri and Chamberlain, 1997; McGrew et al., 2000). Geobarometric pressure determinations, and inferred tectonic burial of Cambrian and Neoproterozoic strata, decrease both southward in the Ruby Mountains and eastward to the Wood Hills. Indicated maximum metamorphic temperatures decrease eastward to the Pequop Mountains and southward to Spruce Mountain, where lower Paleozoic strata contain only lower greenschist facies effects, and higher strata are subgreenschist facies (Thorman, 1970; Harris et al., 1980; Camilleri and Chamberlain, 1997).

Pressures determined for mineral assemblages formed during Tertiary mylonitization range from 3.1 to 5 kbars (Table 1; Table 1. Mineral Geobarometry in the Ruby Mountains Metamorphic Complex

<table>
<thead>
<tr>
<th>Locality and rock</th>
<th>Km NNE of A (Figs. 2, 4, 5, 6)</th>
<th>P (kbars)</th>
<th>Depth equivalent (km)</th>
<th>T (°C)</th>
<th>Event</th>
<th>Estimated age (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Ruby Mountains-East Humboldt Range</strong></td>
<td></td>
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<td></td>
<td></td>
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<tr>
<td>Harrison Pass pluton</td>
<td>25</td>
<td>~3</td>
<td>~10–12</td>
<td>~250–500</td>
<td>Intrusion</td>
<td>36</td>
<td>Barnes et al., 2001</td>
</tr>
<tr>
<td>Dawley Canyon area</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Neoproterozoic(? pelite</td>
<td>30–32</td>
<td>(2.6)–3.7</td>
<td>(&gt;10)–14</td>
<td>400–570</td>
<td>1st metamorphism (M1)</td>
<td>~153</td>
<td>Hudec, 1992</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.5–4.7</td>
<td>13–18</td>
<td>540–660</td>
<td>2nd metamorphism (M2)</td>
<td>~153</td>
<td>Hudec, 1992</td>
</tr>
<tr>
<td>Mahew Creek area</td>
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<tr>
<td>Neoproterozoic pelite</td>
<td>42</td>
<td>4–4.5</td>
<td>15–17</td>
<td>500–550</td>
<td>2nd metamorphism (M2)</td>
<td>~85</td>
<td>Jones, 1999</td>
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<tr>
<td>Ruby Mountains</td>
<td></td>
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<tr>
<td>30–85</td>
<td>3.4–6.5</td>
<td>12–24</td>
<td>365–700</td>
<td>Metamorphism</td>
<td>~75</td>
<td>Hudec, 1992</td>
<td></td>
</tr>
<tr>
<td>Upper Lamoille Canyon Granitoids</td>
<td>57</td>
<td>5.5–6.5</td>
<td>21–25</td>
<td>~300</td>
<td>Intrusion</td>
<td>29</td>
<td>Snoke et al., 1999</td>
</tr>
<tr>
<td><strong>Northern Ruby Mountains and southern East Humboldt Range</strong></td>
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<td></td>
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<tr>
<td>Unmylonitized schist</td>
<td>78–92</td>
<td>5.9–6.7</td>
<td>22–25</td>
<td>680–780</td>
<td>Metamorphism</td>
<td>~75</td>
<td>Hodges et al., 1992</td>
</tr>
<tr>
<td></td>
<td>3.6–4.3</td>
<td>13–16</td>
<td>550–600</td>
<td>Mylonitization nearby</td>
<td>40–20</td>
<td>Hodges et al., 1992</td>
<td></td>
</tr>
<tr>
<td><strong>Southern East Humboldt Range</strong></td>
<td></td>
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<td></td>
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<td></td>
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<tr>
<td>Mylonite</td>
<td>95</td>
<td>3.1–3.7</td>
<td>13–14</td>
<td>600</td>
<td>Mylonitization</td>
<td>40–20</td>
<td>Hurlow et al., 1991</td>
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<tr>
<td><strong>Eastern East Humboldt Range</strong></td>
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<td></td>
</tr>
<tr>
<td>Mafic rock</td>
<td>110</td>
<td>9.5</td>
<td>36</td>
<td>800</td>
<td>Metamorphism</td>
<td>&gt;85</td>
<td>McGrew et al., 2000</td>
</tr>
<tr>
<td>Pelite (deep level)</td>
<td>104–110</td>
<td>5.0–9.4</td>
<td>18–33</td>
<td>650–750</td>
<td>Retrogression and new metamorphism</td>
<td>85–30</td>
<td>McGrew et al., 2000</td>
</tr>
<tr>
<td>Pelite (mylonitic)</td>
<td>94</td>
<td>5 ± 1.4</td>
<td>19 ± 5</td>
<td>630 ± 100</td>
<td>Mylonitization</td>
<td>~40–20</td>
<td>McGrew et al., 2000</td>
</tr>
<tr>
<td>Quartz diorite sill (preliminary pressure estimate)</td>
<td>102–110</td>
<td>4.5–5.5</td>
<td>17–21</td>
<td>~500</td>
<td>Intrusion</td>
<td>40</td>
<td>McGrew and Snee, 1994</td>
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<td><strong>East of East Humboldt Range</strong></td>
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<tr>
<td>Clover Hill</td>
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<tr>
<td></td>
<td>115</td>
<td>4.2–5.8</td>
<td>16–22</td>
<td>580–620</td>
<td>Overprint</td>
<td>&lt;45</td>
<td>Hodges et al., 1992</td>
</tr>
<tr>
<td>Wood Hills</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cambrian pelite</td>
<td>110</td>
<td>5.5–6.4</td>
<td>20–24</td>
<td>530–650</td>
<td>Retrogression</td>
<td>&lt;115</td>
<td>Hodges et al., 1992</td>
</tr>
</tbody>
</table>

1Estimated pressure and ambient temperature determined from metamorphic mineral assemblage compositions in pelitic and quartzofeldspathic rocks except where noted.
2Miarolitic cavities present; authors estimated pressure as lower than indicated (3.6 ± 0.3 to 5.4 ± 0.6 kbar) by Al-in-hornblende barometry.
3Temperature of subgreenschist- to amphibolite-facies host rocks (Burton, 1997).
4Hudec (1992) and Jones (1999) differed in their interpretations of the age of their M2 metamorphic event.
5Pressure determined by Al-in-hornblende technique for granitoids; temperature shown is for host rocks.
7Temperature-conditions determined from mafic metamorphic mineral assemblage.
Hurlow et al., 1991; Hodges et al., 1992; McGrew et al., 2000). A preliminary aluminum-in-hornblende pressure estimate of 4.5 to 5.5 kbars was reported for a late Eocene sill in the East Humboldt Range (McGrew and Snee, 1994), and similar estimates were reported for late Oligocene (29 Ma) intrusions in the Ruby Mountains (Snoke et al., 1999). Intrusion of the circa 36 Ma Harrison Pass pluton at the south end of the metamorphic core was shallower, estimated by Barnes et al. (2001) to be at pressures of about 3 kbar.

**Timing of exhumation**

Decompression and exhumation of the core complex included an early stage dated uncertainly as before late Eocene (?) or older and a better-dated late Oligocene-Miocene stage (McGrew and Snee, 1994). Metamorphic rocks in the East Humboldt Range show a clockwise pressure-temperature path with decompression indicating that they may have experienced as much as 10 km of unroofing in the first stage (McGrew and others, 2000). This unroofing may have occurred by middle Eocene time or earlier, on the basis of complex hornblende $^{40}$Ar/$^{39}$Ar age spectra that suggest a range of early Tertiary cooling ages (Dallmeyer et al., 1986; McGrew and Snee, 1994). Thermochronologic studies of these and other metamorphic core-complex rocks in and near northeastern Nevada have attributed early Tertiary ages to Eocene cooling by extensional unroofing (Dallmeyer et al., 1986; McGrew and Snee, 1994; Lee, 1985; McGrew et al., 2000; Wells et al., 2000; Wells, 2001). Apparent hornblende ages in the East Humboldt Range young downward, from ~60 to ~30 Ma within an exposed altitude interval of ~0.5 km (McGrew and Snee, 1994). This observation could be interpreted to suggest that the first stage of major exhumation was older and that this Paleocene to Oligocene interval reflects very slow (~0.02 mm/yr) unroofing. The timing of early unroofing of the East Humboldt Range therefore remains uncertain.

The burial depths at various times (summarized in Fig. 5) suggest that the north part of the core complex lost one third to one half of its load in the first stage of exhumation (McGrew et al., 2000). Rocks in the central part of the core complex lack kyanite and were never buried so deeply as in the north, and yet magmatic hornblende geobarometry suggests they were still at apparent depths of 21 to 25 km in late Oligocene time (Snoke et al., 1999). This suggests that first-stage exhumation primarily involved the overthickened northern area, in accord with the idea that extensional collapse in the Basin and Range province was focused in the most overthickened areas (Coney and Harms, 1984; Spencer and Reynolds, 1990). The climactic second stage of exhumation affected the entire Ruby-East Humboldt core complex and peaked in the Miocene, when rapid denudation is documented by quenched mineral ages and clasts that were shed into sedimentary basins (Smith and Ketner, 1976b; Dokka et al., 1986). Other core complexes in the Great Basin also show evidence of Miocene quenching attributed to a phase of rapid extensional denudation (Dumitru et al., 2000).

**West-rooted extensional shear zone**

Structures listed in Table 2 all point to westward rooting of the fault zone and major west-directed tectonic unroofing of the core complex. These features include westward dip of the mylonitic shear zone, top-to-west shear indicators such as mylonitic stretching lineation and foliation and shear (S-C) fabrics as well as disharmonic folds, westward younging of
shallower and cooler levels as the core complex was being as the shear zone and deeper infrastructure rose updip to sequence of shoaling depths of extensional shearing, preserved lonites and the low-angle faults record a plastic-to-brittle se-

Ma (Smith and Howard, 1977; Snoke, 1980, 1992). The my-

younger above older rocks and displace rocks as young as 13

crustal section, forming a series of klippen that place mostly

Disharmonic folds

The disharmonic Soldier Creek nappe and most outcrop-scale disharmonic folds in the mylonitic shear zone show top-to-west drag directly down the azimuth of the stretching lineation (Howard, 1966, 1968, 1980)

Cooling ages

Biotite K-Ar and apatite and zircon fission-track cooling ages in the infrastructure and mylonitic carapace range from about 35 to 9 Ma in the Ruby Mountains-East Humboldt Range and consistently young westward, indicating an eastward-tilted section, or progressive quenching beneath a west-moving upper plate (Kistler et al., 1981; Blackwell et al., 1985; Reese, 1986; Hudec, 1990; McGrew and Snee, 1994)

Tiled Tertiary beds

Tertiary strata flanking the core complex and in several ranges on either side dip dominantly eastward into a series of west-dipping normal faults (Mueller, 1964; Hope, 1972; Smith and Howard, 1977; Stewart and Carlson, 1977; Smith and Ketner, 1975; Snoke, 1986; 1992; Fink and van de Kamp, 1992; Hazlett et al., 1992; Schall, 1992); hanging-wall Eocene strata on the west flank of the Ruby Mountains dip more steeply than upper Miocene strata (Smith and Howard, 1977), suggesting that tilting progressed before and after the late Miocene

Fault striae

Fault striae at the base of extensional klippen trend 310° near Harrison Pass (Burton, 1997); semi-ductile “hot slickenlines” below klippen on the west side of the Ruby Mountains 25 km to the north trend an average 270°; slickenside striae there are more variable in orientation; Smith and Howard’s (1977) earlier interpreted the klippen as thrust faulted, but the faults superpose nonmylonitic, low-grade, lower Paleozoic strata onto higher-grade mylonites and thus attenuate structural sec-

Normal faults at north flank of core complex

Mueller (1993), Mueller and Snoke (1993a), and Mueller et al. (1999) concluded there are three systems of low-angle normal faults at the north end of the core complex: an early Tertiary one and a Neogene-Quaternary one both down to west, and an intermediate-age one (mapped only north of the core complex) down to east

Dip of west-flanking normal faults

Mapped west- or northwest-dipping normal faults showing movement into the late Quaternary bound the west flank of the Ruby-East Humboldt Range (Sharp, 1939; Mueller and Snoke, 1993a; Howard, 2000); seismic reflection profiles across the range-front fault system suggest it dips about 20° NW off the Ruby Mountains and East Humboldt Range below a half graben of southeast-dipping strata (Effimoff and Pinezich, 1981; Reese, 1986; Smith et al., 1989; Hazlett et al., 1992; Saturanga and Johnson, 2000)

Table 2. Evidence That West-Rooted Extensional Shear and Faulting Unroofed the Core Complex

<table>
<thead>
<tr>
<th>Feature</th>
<th>Evidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mylonite-zone dip</td>
<td>The mylonitic shear-zone carapace dips and roots dominantly westward or northwestward except where locally domed (Snoke, 1980, 1992; Howard, 1987, 2000; Mueller and Snoke, 1993a)</td>
</tr>
<tr>
<td>Mylonite fabric</td>
<td>The stretching lineation in the mylonites, indicating shear azimuth, strikes consistently to the west-northwest (275°–290° azimuth) throughout the northern Ruby Mountains and East Humboldt Range (Howard, 1980; Snoke and Lush, 1984); structural indicators in the mylonites such as S-C intersections indicate dominantly top-to-west shear, only locally top-to-the-east (Snoke and Lush, 1984)</td>
</tr>
<tr>
<td>Disharmonic folds</td>
<td>The disharmonic Soldier Creek nappe and most outcrop-scale disharmonic folds in the mylonitic shear zone show top-to-west drag directly down the azimuth of the stretching lineation (Howard, 1966, 1968, 1980)</td>
</tr>
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</tr>
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| Fault striae | Fault striae at the base of extensional klippen trend 310° near Harrison Pass (Burton, 1997); semi-ductile “hot slickenlines” below klippen on the west side of the Ruby Mountains 25 km to the north trend an average 270°; slickenside striae there are more variable in orientation; Smith and Howard’s (1977) earlier interpreted the klippen as thrust faulted, but the faults superpose nonmylonitic, low-grade, lower Paleozoic strata onto higher-grade mylonites and thus attenuate structural sec-
| Normal faults at north flank of core complex | Mueller (1993), Mueller and Snoke (1993a), and Mueller et al. (1999) concluded there are three systems of low-angle normal faults at the north end of the core complex: an early Tertiary one and a Neogene-Quaternary one both down to west, and an intermediate-age one (mapped only north of the core complex) down to east |
| Dip of west-flanking normal faults | Mapped west- or northwest-dipping normal faults showing movement into the late Quaternary bound the west flank of the Ruby-East Humboldt Range (Sharp, 1939; Mueller and Snoke, 1993a; Howard, 2000); seismic reflection profiles across the range-front fault system suggest it dips about 20° NW off the Ruby Mountains and East Humboldt Range below a half graben of southeast-dipping strata (Effimoff and Pinezich, 1981; Reese, 1986; Smith et al., 1989; Hazlett et al., 1992; Saturanga and Johnson, 2000) |

cooling ages in the footwall, eastward-tilted Tertiary strata in allochthonous fault striae, and mapped and seismically im-

aged west-rooted faults. Cooling ages of 20 to 25 Ma that record climatic quenching of the core complex rocks (Dokka et al., 1986) imply that an Eocene reconstruction of the Ruby-

East Humboldt Range should restore much of the unroofing.

In the mylonitic shear zone, rock units and lithologic layering thin markedly (Howard, 1980; Snoke, 1980; Snoke and Lush, 1984). Mylonitization, although it possibly began in Eocene or earlier time, affected 29 Ma intrusive rocks and was finished by 25 to 20 Ma as determined from cooling ages. It thus operated between 29 and 20 Ma as determined from cooling ages. Below I speculate that an extensional transfer fault occurs at this boundary, which could bear on restoring the amount of separation on the extensional fault system.

The exposed core complex plunges northward beneath cover rocks in the vicinity of Wells, Nevada, near a zone of east-striking faults (Mueller and Snoke, 1993a; Camilleri and Chamberlain, 1997). Cenozoic extension appears greater south of this zone, which may mark a boundary in the exten-

sional pattern (Newman and Witter, 1992; Mueller and North boundary of the core complex—A transfer zone?

The north boundary of the core complex forms a significant element in the extensional picture (Mueller and Snoke, 1993a). Below I speculate that an extensional transfer fault occurs at this boundary, which could bear on restoring the amount of separation on the extensional fault system.

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sional pattern (Newman and Witter, 1992; Mueller and
the core complex (Fig. 5) and the northward increase in thick-
ness. As estimated from the grain size, requires many assumptions
about texture-strain relations, mylonite thickness, fault throw,
and initial fault dip; and any estimate of net slip rate is highly
uncertain (Hacker et al., 1990, 1992). Another approach at-
ttempted in Table 3 incorporates my earlier speculation that
the "Wells fault" is a Tertiary transfer zone whose offset
(~60–70 km) may approximate the extensional separation.

Yet another approach to finding the horizontal separation
makes use of the westward-younging patterns of biotite and
apatite mineral dates observed by Kistler et al. (1981) and
Reese (1986) across the footwall metamorphic rocks. The pat-
terns may simply reflect an east-tilted section through a zone
of partial age retention (Kistler et al., 1981; Blackwell et al.,
1985). If instead the westward younng patterns record se-
quential quenching below an unroofing fault (McGrew and
Snee, 1994), the indicated net rate of horizontal separation
are ~0.5 to ~1.2 km/m.y. (Table 3). This rate is a half to a
whole order of magnitude slower than at core complexes in
the Colorado River extensional corridor or the Aegean Sea
(Foster et al., 1993; John and Foster, 1993; John and Howard,
1995). Thermochronologic transect studies in progress across
the core complex are attempting to test further for any fast
early Miocene spurt of westward-directed unroofing. Other-
wise the present evidence is consistent with a rate of lateral
unroofing that was slower than for other core complexes. The
various results in Table 3 are consistent with net separation of
the Ruby-East Humboldt metamorphic core from its roof
rocks of 50 km or more, comparable to the core complexes in
the Colorado River extensional corridor (Reynolds and
Spencer, 1985; Howard and John, 1987; Hillhouse and Wells,
1991). Location of the hanging wall of the Ruby-East Hum-
boldt metamorphic complex ~50 km westward or west-north-
westward would suggest that the high-grade rocks once lay
approximately beneath the upper crustal rocks exposed in the
Piñon Range and Adobe Range as earlier suggested (Jansma
The following discussion assumes this possible restoration as
a guide to integrating the geology of the upper and middle
crust.

Separation of hanging wall and core-complex footwall

If dominant westward-directed unroofing is assumed
(Table 2), the metamorphic rocks of the core complex once
were overlain by rocks exposed somewhere to the west. How
far west these rocks were transported depends on the total
separation on the detachment fault system, on whether any of
the unroofing sequence was east directed, as Mueller (1993)
indicated for areas north of the Holborn fault, and on the par-
tition of extension into post-Eocene and earlier phases. The
northward increase of maximum structural burial depths in
the core complex (Fig. 5) and the northward increase in thick-
ness of the mylonitic shear zone (Mueller and Snoke, 1993a)
both imply that early tectonic unroofing was greatest in the
north. Perhaps detachment-fault displacement was greatest
there too. As extensional faults in the Basin and Range
province appear to sole in plastic middle crust at depths of
about 10 to 15 km (e.g., Wernicke, 1992), it is unclear how
rocks were exhumed structurally from >30 km depth. Either
cold geotherms allowed initial fault systems to penetrate this
deep, or sequential exhumation proceeded in stages along
multiple fault systems. The early phase of exhumation, which
halved the tectonic load above the north part of the core com-
plex before Miocene (?) time (Fig. 5), therefore may have
been along structures other than the now-prominent west-
rooted shear zone.

Table 3 calculates partial estimates of horizontal separation
or net heave on the detachment system using a variety of ap-
proaches and assumptions, although none is definitive. The
50-km width of the metamorphic belt beneath extensional
tkliplpen suggests a net horizontal separation of at least 50 km.
This separation includes any contributions from eastward-di-
rected or northward-directed unroofing (Mueller and Snoke,
1993a). An approach using shear-strain rate in the mylonite,
as estimated from the grain size, requires many assumptions
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Integrated Crustal Section

Two stacked sections

A suggested Eocene north-south cross section (Fig. 6) uses a section of the Piñon Range and Adobe Range to portray the upper crust and a section along the exposed core complex for the deeper crust. Vertical separation is greater in the north than in the south, but for this simple restoration no north-south variation in horizontal separation is assumed. The two sections portrayed in Figure 6 may not have been exactly stacked, but they likely typify the crustal levels portrayed. The Carlin gold trend lies mostly west of the Piñon and Adobe Ranges and obliquely intersects section B-B’ in Figure 6.

Eocene depths of the metamorphic and igneous rocks in section A-A’ (Fig. 6) were assumed from the geobarometric estimates graphed in Figure 5. Important caveats are that an Eocene timing for the geobarometry of the mylonitic mineral assemblages (in the northern third of the core complex) is uncertain, and that the geobarometric results yield a broad range of depth estimates.

Extensional unroofing was least in the southern area depicted in Figures 5 and 6, so the southern Piñon Range may not restore as far east as the Ruby Mountains. Sinistral faulting and counterclockwise rotation in the southernmost Ruby Mountains and the hanging wall to the west may reflect accommodation to the differential extension (Palmer et al., 1991; Nutt and Good, 1998).

Eocene depths as shown for the East Humboldt Range (northern end of A-A’) correspond to about half of the maximum calculated pressures (Fig. 5), following the conclusion of McGrew et al. (2000) that sequentially lower pressures resulted from partial unroofing by late Eocene time. Crustal thickness in the Eocene before thinning to the present thickness is unknown. Storerz and Smithson (1998) modeled present crustal structure below the Ruby Mountains as a series of deep and middle crustal layers that thicken northward at the expense of shallower, lower-velocity layers, and a Moho that deepens northward from 32 to 35 km. Saratunga and Johnson (1998) offered a less asymmetric velocity model. Such models probably serve as poor guides to the Eocene crustal structure, because the deep crust likely experienced major change during accommodation to post-Eocene extension (Gans, 1987; Wernicke, 1992; MacCready et al., 1997).

Structure in the deep Eocene crust

The late Eocene Harrison Pass granodiorite pluton in the central Ruby Mountains intruded a boundary zone separating a homoclinal section of low metamorphic grade Paleozoic strata to the south from deformed metamorphosed strata and leucogranite to the north (Fig. 6). Mesozoic andalusite and sillimanite metamorphic assemblages in Cambrian and Neoproterozoic rocks in the central Ruby Mountains pass northward to sillimanite assemblages in the northern Ruby Mountains and to kyanite-bearing assemblages in the East Humboldt Range, Clover Hill, and Wood Hills (Thorman, 1970; Hodges et al., 1992; Hudec, 1992; Snoke, 1992; Jones, 1999; McGrew et al., 2000). Mesozoic leucogranites inflate the exposed infrastructure by a factor of two, decreasing upward (Howard et al., 1979; Howard, 1980; 2000; McGrew et al., 2000). The leucogranites were derived by crustal anatexis of McGrew et al. (2000) that sequentially lower pressures resulted from partial unroofing by late Eocene time. Crustal thickness in the Eocene before thinning to the present thickness is unknown. Storerz and Smithson (1998) modeled present crustal structure below the Ruby Mountains as a series of deep and middle crustal layers that thicken northward at the expense of shallower, lower-velocity layers, and a Moho that deepens northward from 32 to 35 km. Saratunga and Johnson (1998) offered a less asymmetric velocity model. Such models probably serve as poor guides to the Eocene crustal structure, because the deep crust likely experienced major change during accommodation to post-Eocene extension (Gans, 1987; Wernicke, 1992; MacCready et al., 1997).
indicate that high-temperature, H$_2$O-rich fluid infiltration (McGrew et al., 2000). Stable isotope and petrologic studies (Kistler et al., 1981; Lee and Barnes, 1997; Batum, 1999; McGrew et al., 2000). The miogeoclinal section is duplicated at A-A' and B-B' levels, and further structural duplications are inferred to occupy the mostly unexposed intervening interval. Strata shown in the northern two thirds of A-A' are inflated 30 to 70 percent by Cretaceous leucogranite. The upper part of section B-B' is interpreted from a cross section of the Piñon Range by Smith and Ketner (1978), and maps and sections of the Adobe Range by Ketner (1973), Silitonga (1974), Solomon and Moore (1982), Coats (1987), and Ketner and Alpa (1992). Lower part of section (A-A') is interpreted from cross sections of the Ruby Mountains by Howard et al. (1979), of the East Humboldt Range by Snake et al. (1997), and from a projection of the Harrison Pass pluton (assuming the pluton is tilted west 30°) after Burton et al. (1997). Rock units are as follows: €Z = Cambrian and Neoproterozoic quartzite, DOW = allochthonous Devonian to Ordovician western and transitional facies, DS = Devonian and Silurian eastern facies, Jg = Jurassic leucogranite, Kg = Cretaceous leucogranite gneiss and rafts of country rock, M = Pennsylvanian and Mississippian, OC = Ordovician and Cambrian eastern-facies carbonates and shale, P = Permian and Pennsylvania strata, Ti = Eocene to early Oligocene intrusions, Ts = Eocene sedimentary rocks (folded and eroded from view in the left part of the section), Tr = Triassic, Tv = Eocene to lower Oligocene volcanic rocks, W = Archean and Paleoproterozoic (?) basement gneisses, Z = Neoproterozoic pelite and quartzite.

Tectonic complexity in the Ruby-East Humboldt Range infrastructure increases northward (Fig. 6). Northward structural plunges and the presence of stacked nappes in the north that include Precambrian basement are consistent with the greater tectonic loading in the north that is implied by mineral assemblage data (Fig. 5).

The exposed structures in the metamorphic complex include thrusts, fold nappes, and recumbentally folded thrusts involving miogeoclinal strata (Fig. 6). The structures show strikingly diverse vergence. During metamorphism, Late Cretaceous (?) folding produced the southeastward overturned Lamoille Canyon fold nappe and the north-northeastward-overturned King Peak fold nappe in the Ruby Mountains, and northwestern-overturned folds in the Wood Hills (Fig. 4; Thorman, 1970; Howard, 1980, 1987; Camilleri and Chamberlain, 1997; Jones, 1999; McGrew et al., 2000). The initially Mesozoic (?) Soldier Creek fold-thrust nappe (SN in Fig. 4) and the next-higher recumbent synclines in the northern Ruby Mountains (Howard, 1980; Snake, 1980) are stretched owing to Tertiary extensional shearing, but they show westward vergence that is probably inherited from Mesozoic folds. Metamorphosed thrust (?) slices that duplicate the section a few kilometers north of the Secret Creek nappe (Snake, 1980; Valasek et al., 1989), and the Archean basement-cored Winchell Lake fold nappe (Fig. 4) together with fault sheets that it folds (Lush et al., 1988; McGrew et al., 2000), are so modified by Tertiary extensional shearing that the original vergence, facing, and attenuated thicknesses are indeterminate.

A middle crust inflated by pre-Eocene tectonic thickening

Figure 5 portrays the overthickened north part of the section as substantially attenuated by Eocene (?) time (McGrew and Snee, 1994; McGrew et al., 2000). McGrew et al. (2000) deduced that pressures and depths for rocks in the East Humboldt Range had by 40 to 50 Ma decreased by a factor of two or more from their maxima. Mesozoic tectonic loading was less for rocks to the south in the Ruby Mountains, which are not known to have undergone attenuation by Eocene time, because hornblende barometry implies that rocks exposed in Lamoille Canyon were still at substantial depth when late Oligocene granitoids were emplaced. The tectonic load resulted from episodic crustal thickening by telescoping from mid-Paleozoic through Mesozoic time, especially in the Cretaceous (Riva, 1970; Smith and Ketner, 1976a, 1978; Thorman et al., 1991; Hodges et al., 1992; Smith et al., 1993; Miller and Hoisch, 1995; Camilleri et al., 1997; Jones, 1999; Taylor et al., 2000). The abundance of Mesozoic granitoids is greatest on the north side of the Harrison Pass pluton, decreases northward, and decreases structurally upward.
in the northern Ruby Mountains and East Humboldt Range; it decreases eastward into the lower-amphibolite facies rocks of the Wood Hills, which contain only isolated pegmatites. This abundance pattern suggests that granites contributed only a fraction of the now-removed tectonic load from above the Ruby Mountains core complex. Instead, the regional geology suggests that this tectonic load consisted dominantly of cryptic Mesozoic thrust sheets and other nappes.

Extensional klippen that structurally overlie the deep crustal rocks of the core complex provide glimpses of the now-dismembered crustal interval that once overlay the deep rocks. These klippen include stacks of thin slivers in the southern East Humboldt Range, each placing lower-grade rocks on higher-grade rocks (Snoke, 1980, 1998; Snoke et al., 1997). Most place younger strata above older strata, but some show the opposite relation, suggestive of preextension thrust-duplicated sections. The klippen probably are displaced westward many kilometers relative to the metamorphic foothills they overlie. Incomplete stratigraphic sections and the scarcity of allochthonous granites beheaded from the nongratitic footwall indicate that these klippen record only a small, fragmentary sample of the preextension load that geobarometry results imply was above the metamorphic complex. Camilleri and McGrew (1997) pointed out that preextension structural duplication is required in the Wood Hills where an extensional fault separates a klippe of unmetamorphosed Devonian to Permian strata from a footwall where the same strata are metamorphosed. Mueller et al. (1999) interpreted a klippe of Ordovician migmatochon rocks, faulted in the northern Pequop Mountains onto Permian rocks in the core-complex footwall (KP in Fig. 4), to be superposed by means of extensional faulting from an inferred higher, older-above-younger thrust plate of Ordovician and younger rocks.

The core-complex domain lacks thick, up-ended, intact, allochthonous tilt blocks that at other highly extended corridors in the Basin and Range province commonly expose sections throughout 8 to 15 km of the upper crust (Anderson, 1971; Proffett, 1977; Howard and John, 1985; Fryxell et al., 1992; Howard and Foster, 1996; Howard et al., 2000). In their absence, I conclude that upper crust denuded from the Ruby-East Humboldt core complex remains largely upright and hidden, and it underpins exposed hanging-wall rocks to the west.

Camilleri and Chamberlain (1997, 1998) proposed that facies evidence that the deep burial of core-complex rocks occurred during Cretaceous east-transported thrusting. They judged that a limey facies of the Devonian Roberts Mountains Formation, now in a Tertiary extensional klippe in the Pequop Mountains, had first been deposited farther west and thrust eastward over a correlative dolomitic facies in the underlying metamorphosed footwall (but see Wise, 1998, for a different interpretation). Cryptic preextension east-directed thrusting can similarly explain a possible western provenance for facies of Carboniferous rocks that occupy west-transported extensional klippen in the central and southern Ruby Mountains (Willden et al., 1967; Armstrong, 1972; Burton, 1997).

Any structural load over rocks in the central Ruby Mountains was small by late Eocene time if the ~36 Ma Harrison Pass pluton was emplaced at ~3 kbars as suggested by Barnes et al. (2001). This value may be a minimum for the upper part of the exposed pluton, given conflicting geobarometric evidence. Geobarometric estimates for plutonic hornblende average ~5.4 kbar for some subunits of the pluton and ~3.6 kbars for others. Possible miarolitic cavities identified near the pluton's roof could imply pressures less than 2 kbars; and metamorphic andalusite in adjacent wall rocks imply intrusions less than 4 kbars if the andalusite is due to contact metamorphism (Barnes et al., 2001), although Burton (1997) cited textural evidence that the andalusite may predate the pluton. If the pluton intruded at ≥10 km depth, the cover for its roof would include about 5 to 7 km of tectonic load added to the normal stratigraphic sequence (Fig. 6).

If the complexity of thrust and fold nappes exposed in the metamorphic rocks is any guide to structural style in the denuded roof, many telescoping structures, not just one thick sheet, provided the tectonic burial required to explain the geobarometric results. The northward increase in Mesozoic geobarometric burial depth implies that a larger structural load was present there, perhaps by more or thicker nappes than were piled above southern parts of the metamorphic complex. Northward plunges of exposed structures (Fig. 6, A-A') support the presence of a greater structural load to the north.

Camilleri and Chamberlain (1997) hypothesized an unexposed, large-throw, east-directed thrust-fault plate to explain the deep tectonic burial in the East Humboldt Range. They also implied in cross section that the basal fault of this plate dives eastward beneath the Goshute-Toano Range (Fig. 2). The small stratigraphic relief of pre-Cretaceous strata unconformably below Paleogene deposits in the hinterland domain (Armstrong, 1968), however, argues against exposure in the Eocene of such a prominent thrust fault. The base of the Paleogene section unconformably overlies Mesozoic strata in the southern Pequop Mountains area, which seem structurally tied to the metamorphic complex as it was mapped by Camilleri and Chamberlain (1997). Furthermore, no clear regional thrust having tens of kilometers displacement is obvious in the exposed hinterland upper crustal section.

Models of tectonic wedging or blind structures (Snoke and Miller, 1988, Fig. 7), or duplexes or a steeply downthrusted prong (Miller and Hoisch, 1995; Lewis et al., 1999) have been proposed to explain structural thickening in the region and at the same time attempt to avoid large stratigraphic relief at the surface. Tectonic wedging and associated backthrusting (Wentworth et al., 1984; Price, 1986; Wentworth and Zoback, 1989; Janssen, 1993) could be consistent with the multiple directions of Jurassic and Cretaceous structural transport. As an example, Price's (1986) model for the Mesozoic Canadian Cordillera explained shortening, thickening, and structures showing opposing vergence as caused by a wedge pushed eastward into and under North American sedimentary rocks (Fig. 7).

Eocene Landscape and Upper Crust
To develop an integrated crustal framework for the late Eocene entails a survey of upper crustal features and processes, followed by analysis of processes in the deep crust.
In middle and late Eocene time, northeastern Nevada exposed a substrate of Paleozoic and Mesozoic rocks onto which Eocene sedimentary rocks and volcanic rocks were deposited in broad alluvial plains and lake basins. In the hinterland domain, Paleogene deposits rest exclusively with little angularity on upper stratigraphic levels of the eastern-facies miogeoclinal sequence (Armstrong, 1968, 1972; Coats, 1987; Brooks et al., 1995). In the core complex domain, the substrate unconformably below Eocene rocks ranges 2 to 4 km in stratigraphic level, from allochthonous Triassic strata in the East Humbold Range to Devonian strata in the Pequop Mountains (Brooks et al., 1995; Camilleri and Chamberlain, 1997).

In contrast, the Eocene deposits in much of the Elko-Carlin domain lap over a substrate of eroded thrust sheets and folds—mostly of western-facies Paleozoic deep-water and continental-slope deposits and locally of upper Paleozoic overlap facies (Riva, 1970; Oversby, 1972; Smith and Ketner, 1976a, 1978; Miller et al., 1981; Ehman, 1985; Coats, 1987; Ketner and Evans, 1988; Thorman and Brooks, 1988; Smith and Miller, 1990; Ketner and Alpha, 1992). The pre-Eocene substrate in the Cortez Range area consists of Jurassic volcanic rocks and Cretaceous lacustrine and fluvial deposits (Muffler, 1964; Smith and Ketner, 1976b). The Peklo Hills (Fig. 2), which likely restore above the northern part of the core complex, expose Eocene(?) strata resting on Paleozoic strata that were complexly folded and thrust faulted in the Mesozoic (Ketner and Evans, 1988; Thorman et al., 1991). Thin-skinned, Mesozoic thrust sheets that carried siliceous, western-facies strata east or south may have capped the northern part of the crustal section—judging, for instance, by the five thin sheets present in the Snake Mountains and the five in the HD Range just north of the section in Figure 6 (Riva, 1970; Coats and Riva, 1983; Thorman and Brooks, 1988; Smith et al., 1990). Such sheets in the Adobe and Pìon Ranges were eroded by Eocene time and overlapped by Eocene sedimentary rocks (Fig. 6; Smith and Ketner, 1978; Ketner and Ross, 1983).

The upper-crustal substrate to Eocene deposits in northeastern Nevada is little metamorphosed in the hinterland domain and in the area between the Ruby-East Humboldt core complex and the Pìon and Adobe Ranges. West and north of Elko and Carlin, some Paleozoic strata record paleotemperatures of 300°C or more (Harris et al., 1980; Poole et al., 1983; Cunningham, 1985; Ranson and Hansen, 1990), which may indicate that a substantial cover was eroded before the Eocene. Strata in the area of the Independence district yielded Precambrian fission-track ages on clastic zircon grains that indicate that Phanerozoic paleotemperatures remained less than 200°C (Hofstra et al., 1999).

Eocene lakes and volcanism

Lacustrine deposits of the Elko Formation record a broad Eocene lake system in the Elko-Carlin domain (Fig. 7; Smith and Ketner, 1976a; Solomon et al., 1979; Ketner and Alpha, 1992; Schalla, 1992; Solomon, 1992). The deposits contain oil shale and limestone and are up to 1 km or more thick. Small lakes containing flora indicative of adjacent forested hills were present in the northern part of the Elko-Carlin domain (Axelrod, 1966). Eocene lakes in the hinterland domain formed depocenters for the White Sage Formation along the Utah border and the Sheep Pass Formation in east-central Nevada near Ely (Fig. 7; Fouc, 1979; Dubiel et al., 1996). The Elko Formation mostly is found in the Elko-Carlin domain west and north of the Ruby Mountains. An exception is the Sohio Ruby Valley no. 1 well encountered a comparable oil-shale-bearing section 20 km east of the Ruby Mountains (Fink and van de Kamp, 1992) and ~20 km west of a breakaway zone of the west-rooted extensional fault system. Vandervoort and Schmidt (1990) suggested that extension following the Sevier orogeny provided the basins for Cretaceous to Eocene lakebeds in eastern Nevada. The Elko Formation thins westward from the Ruby Mountains (Sitaruga and Johnson, 2000), it occupies the hanging wall of the west-dipping Ruby-East Humboldt detachment-fault system, and its basin may mark an early phase of extensional faulting (Henry et al., 2001).

Conglomerate at a similar stratigraphic level underlies Eocene volcanic rocks and lies across faulted upper and middle Paleozoic strata in the northeast part of the core complex domain. In the Pequop Mountains (Brooks et al., 1995; Camilleri and Chamberlain, 1997) the conglomerate is earlier than 41 Ma, and it contains ostracod-bearing siltstone and large boulders derived from granodiorite and the Diamond Peak Formation (Carboniferous). Prevolcanic conglomerate dated between 41 and 39 Ma in the southeastern East Humboldt Range also contains ostracod-bearing clasts derived from lakebeds (Brooks et al., 1995). The stratigraphic and geographic position of these conglomerates suggests they may record Eocene scarps that developed along the eastern border of the Elko Formation lake system.

Calc-alkaline volcanic flows and tuffs dated as late middle Eocene and late Eocene overlie and are interbedded with the Eocene lacustrine strata (Brooks et al., 1995; Henry and Boden, 1998; Henry and Ressel, 2000). The volcanic rocks
and their numerous eruptive centers—the Tuscarora magmatic belt of Christiansen and Yeats (1992) and northeast Nevada volcanic field of Brooks et al. (1995)—mark a midpoint in the early Eocene to Miocene migration of arc-like magmatism southward from British Columbia to southern Nevada (Armstrong and Ward, 1991; Christiansen and Yeats, 1992). The southward migration may relate to steepening or delamination of a formerly shallow-dipping subducting slab (Humphreys, 1995; Sonder and Jones, 1999).

Although the widespread lake deposits suggest that broad basins and low prevolcanic Eocene topographic relief were present in much of eastern Nevada, local unconformities within Eocene sections suggest that active faulting occurred at basin margins (Ehman, 1985; Potter et al., 1995; Mueller et al., 1999). Sedimentary breccias near the East Humboldt Range record late Eocene or Oligocene slumping derived from incipient unroofing of low-grade metamorphic rocks (but not mylonites) in the core complex (Snoke et al., 1990, 1997; Mueller and Snake, 1995b).

**Eocene upper-crustal deformation and dike orientation**

Structural and stratigraphic evidence of Eocene to early Oligocene extensional faulting and associated unconformities is scattered through northeastern Nevada (Zoback et al., 1981; Ehman, 1985; Miller, 1990; Seedorff, 1991; Smith et al., 1991; Brooks et al., 1995; Potter et al., 1995; Mueller et al., 1999; Gans et al., 2001; Henry et al., 2001; Munthe et al., 2001). Steep to moderate westward to northwestward pre-Miocene dikes of Eocene rocks in the Mount Velma, Bull Run, and Copper Basin areas north of Elko (Coash, 1967; Ehman, 1985; Coats, 1987; Rahl et al., 2002), and a southeastward pre-Miocene dip of Eocene rocks on the west flank of the Ruby Mountains (Smith and Howard, 1977) indicate the strike of Paleogene tilt axes and imply that Paleogene extension was oriented east-west to northwest-southeast. Eocene north- to northwest-trending folds were described near Elko, in the Piñon Range, and in the southernmost Ruby Mountains area (Ketner and Alpha, 1992; Nutt and Good, 1998); they may include local transpressional structures related to extension (Nutt et al., 2000) and drag folds along extensional faults.

Eocene dikes near Tuscarora strike mostly northeastward, whereas Eocene to early Oligocene dikes in the Piñon Range and Emigrant Pass areas strike mostly northward (Smith and Ketner, 1978; Henry and Ressel, 2000; Henry et al., 2001). These orientations suggest that in the Eocene the least compressive stress was oriented northwest-southeast to east-west in the Elko-Carlin domain.

In summary, the regional extent and magnitude of late Eocene extension in northeast Nevada remain indistinct (Seedorff, 1991). Crustal thinning may have included broad subsidence (Henry et al., 2001) and eventual partial dismemberment. The evidence, including cooling-age hints of a Paleogene stage of core-complex exhumation, is consistent with the idea that crust that had been shortened and overthickened by Mesozoic telescoping began to collapse in Late Cretaceous or Paleogene time (Coney, 1987; Vandervoort and Schmidt, 1990; Livacari, 1991; Hodges and Walker, 1992; Costenius, 1996; Sonder and Jones, 1999). Extensional faulting in the Bull Run and Copper Basin areas north of Elko accompanied Eocene volcanism (Ehman, 1985; Rahl et al., 2002). The northwest-southeast to east-west orientation of late Eocene to Oligocene extension implied by tilt axes and by dikes was approximately perpendicular to the older Paleozoic and Mesozoic contractional orogen.

**Deep Eocene Processes**

**Magmatism and fluid flow**

Deep counterparts to the Eocene calc-alkaline volcanic rocks in the Elko-Carlin area were intruded at depths ~210 km into the metamorphic rocks of the core complex. The belljar-shaped ~36 Ma Harrison Pass pluton of granodiorite and granite (Fig. 6) partly downfolded its wall rocks, and magmatic stoping partly engulfed its roof of Paleozoic strata (Wright and Snook, 1993; Burton et al., 1997). Granodiorite magma may have accumulated at a depth ~20 km (5.4 ± 0.6 kbars) before rising to the site of the pluton’s emplacement at perhaps 10 to 15 km depth (~3 kbars; Barnes et al., 2001). The composite pluton evidently collected several other batches of magma from partial melting sites in Mojave province crust, mixed in some cases with mantle-derived magmas that provided heat for melting (Barnes et al., 2001). Downfolded wall rocks (Fig. 6) suggest that the pluton’s rise may have been matched by return flow in the wall rocks. The pluton’s intrusion was close to the time of the first erupted tuffs exposed in adjacent ranges (Palmer et al., 1991; Nutt and Good, 1998; Gordee et al., 2000). Any future evidence that volcanic rocks and shallow intrusions exposed in the Piñon Range were physically connected to the pluton would improve estimates of the extensional fault displacement and would inspire reconstructions of an Eocene magmatic plumbing system.

Farther north in the core complex, sills dated as Eocene by U-Pb methods intruded the metamorphic rocks. The two widest sheets are peraluminous granite (gneiss of Thorpe Creek, 36–39 Ma) and quartz diorite (40 ± 3 Ma), each many kilometers broad (Fig. 4); smaller quartz diorite and granodiorite bodies are dated as ~35 to 38 Ma (Howard et al., 1979; McGrew, 1992; Wright and Snook, 1993; MacCready et al., 1997). The intrusions correlate temporally with eruption of volcanic rocks from widespread centers in the region (Brooks et al., 1995; Henry and Ressel, 2000). In contrast, younger, widespread, and small late Oligocene (~29 Ma) granites in the core complex lack many nearby volcanic or intrusive correlations. The dated Eocene sheets in the core complex are attenuated and mylonitized by postinjection deformation. The gneiss of Thorpe Creek (located in the northern Ruby Mountains) averages less than 100 m thick, and the less extensive largest quartz diorite sheet (in the East Humboldt Range) may average 200 m. Together their bulk is less than an order of magnitude that of the Harrison Pass pluton. Assuming that the pluton averages 3 km thick, a rough volume estimate suggests it is ~50 times as voluminous as each of the two sheet intrusions. If the average thickness of crustal section exposed in the metamorphic complex of the Ruby and East Humboldt Ranges is taken as 3 to 4 km, the dated Eocene intrusions account for roughly 10 vol percent.

Stable isotope data from the core complex provide evidence that mantle-derived fluids isotopically homogenized
the deepest exposed metasedimentary rocks during magmatism, and that exchange with meteoric water affected rocks down to the levels of mylonitic rocks close below detachment faults (Gruner and Wickham, 1991; Fricke et al., 1992; Wickham et al., 1993; McGrew and Peters, 1997). The timing of these processes and how they relate to Eocene events remain uncertain.

**Flow, extension, shear, and thermal gradient**

The subhorizontal Eocene sills that intruded into the metamorphic rocks contrast in their orientation to the steep, commonly north-south dikes in the upper crust. The high Eocene temperatures (≥500°C) of deep rocks in the East Humboldt Range (McGrew et al., 2000) suggests that they were less prone than the upper crust to maintain elastic stresses for dike intrusion.

MacCready et al. (1997) proposed from fabric evidence that rocks in the Ruby Mountains were undergoing north-south plastic flow below and simultaneously with mid-Tertiary west-directed extensional shearing, which is best dated as younger than 29 Ma. They interpreted the proposed flow as accommodation to isostatic rise in response to greater unroofing in the north than in the south part of the core complex. Such flow may also be consistent with bulging of the middle crust below the northern part of the exposed infrastructure as modeled from seismic data by Stoerzel and Smithson (1998). Whether such plastic flow occurred in Eocene time is unknown.

Pressure-temperature conditions, fabrics, and mineral dates established for rocks in the East Humboldt Range suggested to several authors, however, that both crustal attenuation and extensional mylonitic shearing may have been operating there by Eocene time (Dallmeyer et al., 1986; Hurlow et al., 1991; Hodges et al., 1992; Wright and Snoke, 1993; McGrew and Snee, 1994; McGrew et al., 2000). McGrew et al. (2000), using 40Ar/39Ar data on hornblende and mineralogic pressure-temperature calculations, found a hot and increasing geothermal gradient and proposed its timing to be 50 to 30 Ma, about 30° to 50°C/km at ~40 Ma. Those authors pointed out that their clockwise pressure-temperature path, which lowered temperature little during decompression, resembles the path expected for areas undergoing extension and thinning of the lower crust.

The Eocene intrusive magmas were evidently 300° to 400°C hotter than their host rocks in the core complex, and they brought much more magmatic advective heat to rocks in the Harrison Pass area than to elsewhere in the core complex. A hot Paleogene geothermal gradient as inferred for the East Humboldt Range by McGrew et al. (2000), whether it resulted from Eocene magmatism or exhumation, would be expected to affect Eocene thermal conditions, hydrothermal circulation, and mineralization processes in the upper crust.

**Discussion**

Late Eocene magmatism and mineralization in northeastern Nevada overlapped temporally with east-west to southeast-northwest extension of the structurally overthickened crust. Earlier Eocene magmatism and coincident extension affected northern parts of the Cordillera before sweeping southward to Nevada (e.g., Janecke et al., 1997). If the inception of magmatism marks steepening of the underlying subduction zone, slab rollback may have helped to drive the extension (Humphreys, 1995; Sonder and Jones, 1999), or upwelling asthenosphere behind a delaminated slab may have driven magmatism and thermally weakened the overthickened crust to the point of extensional collapse (Muntean et al., 2001).

Dispersed Eocene igneous activity and a geotecton further warmed by crustal thinning can be expected to have powered geothermal circulation of fluids in the upper crust. Down-dropping or sagging in the hanging wall of an early phase of the west-rooted fault system that eventually unroofed the Ruby-East Humboldt core complex may have helped form the Eocene basin into which the Elko Formation was deposited. Eocene lakes signal availability of water for hydrothermal circulation systems with a potential for mineralization.

The Carlin trend of gold deposits (Fig. 2) is mostly west of the section portrayed in Figure 6. No recognized major Cenozoic structure clearly controls the north-northwest orientation of the entire Carlin trend. The trend may instead follow a favorable distribution of facies of Paleozoic host strata (Armstrong et al., 1998), perhaps related to the shape of the Paleozoic continental margin (Tosdal et al., 2000), and capped by impermeable Paleozoic allochthons or strata. The crustal depths of gold mineralization lay well above the structural level of rocks now exposed in the core complex, but late Eocene magmas that ponded in the middle crust could have provided regional deep-seated magmatic or metamorphic fluids for upper crustal mineralization (Holstra and Cline, 2000).

Nearly isothermal depressurization, raising of geothermal gradients, and some mylonitic shearing in the middle crust all have been attributed to Eocene or older crustal thinning. If this equivocal timing is correct, Figure 8 suggests how some of these conditions may be related to processes in the upper crust. Faulting in the upper crust would be expected to be compatible with an east-west to northwest-southeast direction of Eocene upper crustal extension. Fluid circulation systems driven by normal faulting, by an elevated geotherm, and by heat from intrusions may have interacted with favorable host strata and structures to enhance mineralization at shallow crustal depths.

To the extent that the rocks in the core complex are representative, middle crustal rocks under the Elko region include metamorphosed and structurally duplicated miogeoclinal sections consisting of carbonate, quartzite, and pelite that were shuffled structurally with underlying Archean and Proterozoic basement gneisses and invaded by abundant Mesozoic granite generated by crustal anatexis at depth. None of these rocks has been well documented as providing a source for metals involved in Carlin-type gold deposits.

The Mesozoic crustal thickening required by the large burial depths of rocks in the core complex suggests that a complex of fold and thrust nappes was present, perhaps involving crustal wedging. These structures would be expected to greatly complicate the geometry and blur the expression of any old, deeply penetrating or persistent crustal flaws such as are commonly assumed to form steep electrical conductors or to be responsible for mineral alignments such as the Carlin trend (Rodriguez, 1998; Crafford and Grauch, 2002;
The surface position of any old, deep structures that could have influenced Eocene mineralization would be expected to be displaced or blurred not only by contractional and extensional allochthons but also by flow and melting deep in the crust. The nearly flat Moho observed throughout the Basin and Range province requires deep crustal flow to accommodate large geographic variations in amount of upper crustal extension (Gans, 1987; Wernicke, 1992). In light of the structural shuffling and flow, it is no surprise that proposed isotopic and geophysical crustal boundaries that might relate to mineral belts (Tosdal et al., 2000; Grauch et al., 2003) appear to be gradational through widths of tens of kilometers.

The tectonically shuffled character of the Eocene middle crust implies that the upper-crustal Carlin gold trend did not and does not root in any simple structure or rock body at middle crustal depth. The Carlin trend furthermore strikes obliquely relative to regional Eocene extensional or igneous structures in the upper crust. These structural mismatches could be explained if the structural setting for the Carlin trend of mineralized rocks relates not so much to deeply penetrating Eocene structures but to a relict northwest-striking belt of favorable host-rock conditions in the upper crust. Favorable host rocks may relate to deposition along a Paleozoic continental margin the roots of which have since been disrupted. Low Tertiary dips and low degrees of tectonic extension in the Elko-Carlin domain suggest that it and its contained linear Carlin trend occupy an upper crustal block that, although detached from the middle crust, is relatively little disrupted internally. Regional Eocene magmatic intrusion, normal faulting, enhanced geothermal gradient, crustal thinning, and ponding of surface water in active structural basins are all processes that can be inferred to have helped drive Eocene hydrothermal systems and mineralization in an upper crustal belt of favorable host rocks.

Acknowledgments

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Wannamaker and Doerner, 2002; Grauch et al., 2003). Radiogenic isotopes signal the underlying presence of Archean basement in the northern East Humboldt Range, in contrast to most of the Ruby Mountains where isotopic values suggest an underlying more heterogeneous basement province of mixed Archean and Paleoproterozoic heritage (Wright and Snoke, 1993; Barnes et al., 2001). Whether or how the basement-cored nappe, deeper Mesozoic burial, and greater early denudation in the East Humboldt Range compared with the Ruby Mountains may reflect this indistinct relict boundary between provinces is unknown.

Regional isotopic, gravity, and magnetic maps have been interpreted as showing deep crust boundaries aligned along and parallel to the Carlin trend and reflecting the continental edge resulting from Proterozoic rifting (Tosdal et al., 2000; Grauch et al., 2003). Those authors suggested that these features have persisted through time and influenced younger structures and mineralization. The structural shuffling demonstrated in the exposed core complex and its involvement of allochthonous basement, however, predicts much disruption of any such old regional crustal boundaries and detachment from their roots.

The surface position of any old, deep structures that could have influenced Eocene mineralization would be expected to be displaced or blurred not only by contractional and extensional allochthons but also by flow and melting deep in the crust. The nearly flat Moho observed throughout the Basin and Range province requires deep crustal flow to accommodate large geographic variations in amount of upper crustal extension (Gans, 1987; Wernicke, 1992). In light of the structural shuffling and flow, it is no surprise that proposed isotopic and geophysical crustal boundaries that might relate to mineral belts (Tosdal et al., 2000; Grauch et al., 2003) appear to be gradational through widths of tens of kilometers.

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