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Carlin-Type Gold Deposits in Nevada: Critical Geologic Characteristics and Viable Models

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Abstract

Carlin-type Au deposits in Nevada have huge Au endowments that have made the state, and the United States, one of the leading Au producers in the world. Forty years of mining and numerous studies have provided a detailed geologic picture of the deposits, yet a comprehensive and widely accepted genetic model remains elusive. The genesis of the deposits has been difficult to determine owing to difficulties in identifying and analyzing the fine-grained, volumetrically minor, and common ore and gangue minerals, and because of postore weathering and oxidation. In addition, other approximately contemporaneous precious metal deposits have overprinted, or are overprinted by, Carlin-type mineralization.

Recent geochronological studies have led to a consensus that the Nevada deposits formed ~42 to 36 m.y. ago, and the deposits can now be evaluated in the context of their tectonic setting. Continental rifting and deposition of a passive margin sequence followed by compressional orogenies established a premineral architecture of steeply dipping fluid conduits, shallow, low dipping "traps" and reactive calcareous host rocks. Sedimentary rock sequences that formed following continental margin rifting or in a foreland basin ahead of an advancing thrust front contain reactive pyritic and carbonaceous silty limestones, the primary host rocks in almost every deposit. The largest deposits now lie in the lower plate to the Devonian to Mississippian Roberts Mountain thrust, which placed nonreactive, fine-grained siliciclastic rocks with less inherent rock permeability, above more permeable carbonate stratigraphy, forming a regional aquitard. North-northwest– and west-northwest–striking basement and Paleozoic normal faults were inverted during postrifting compressional events and formed structural culminations (anticlines and domes) that served as depositional sites for auriferous fluids in the Eocene. These culminations are now exposed as erosional windows through the siliciclastic rocks of the Antler allochthon.

During the Eocene, northwesterly to westerly extension reopened favorably oriented older structures as strike-slip, oblique-slip, and normal-slips faults. Fluid flow and mineral deposition appear to have been fairly passive as there is minimal evidence for overpressured hydrothermal fluids, complicated multistage vein dilatancy, or significant synmineralization slip. Geologic reconstructions and fluid inclusions indicate that deposits formed within a few kilometers of the surface. Ore fluids were moderate temperature (~180°–240°C), low salinity (~2–3 wt % NaCl equiv), CO_2 bearing (<4 mol %), and CH_4 poor (<0.4 mol %), with sufficient H_2S (10^{-1}–10^{-2} m) to transport Au. Ore fluids decarbonatized, argillized, and locally silicified wall rocks, and deposited disseminated pyrite containing submicron Au as Fe liberated from wall rock reacted with reduced S in the ore fluid.

Isotopic studies indicate multiple sources for ore fluids and components and require either different models for different districts or call upon meteoric waters to overwhelm a deep ore-fluid signal in most districts. Oxygen and H isotope ratios of minerals and fluid inclusions indicate a deep magmatic or metamorphic fluid source at the Getchell deposit; however, most similar studies in other districts have identified meteoric water. A large range in S isotopes in ore pyrite from all districts suggests derivation from a sedimentary source; yet studies at Getchell and a few studies in the northern Carlin trend are consistent with a magmatic S source. As a result of these inconsistencies, current models relate deposits to (1) metal leaching and transport by convecting meteoric water, (2) epizonal intrusions, and (3) deep metamorphic and/or magmatic fluids.

With the exception of the isotopic studies, compiled data from all Nevada trends and districts indicate compelling similarities, suggesting that all Nevada Carlin-type deposits formed in response to similar geologic processes. We propose a model in which removal of the Farallon slab promoted deep crustal melting that led to prograde metamorphism and devolatilization, thus generating deep, primitive fluids. Such fluids were likely incorporated in deep crustal melts that rose buoyantly and ultimately exsolved hydrothermal fluids, possibly

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containing Au. Metamorphism at midcrustal levels may have contributed fluids, all of which were collected into basement-penetrating rift faults, where they continued to rise and scavenge various components, evolving in composition to become ore fluids. North-northwest–trending paleo-normal faults and northeast-trending paleo-transform faults, preferentially dilated during Eocene extension, controlled the regional position, orientation, and alignment of the deposits. Eventually the ore fluids accumulated in areas of reduced mean effective stress, particularly boundaries of older Jurassic and Cretaceous stocks and structural culminations. The ore fluids were diluted by exchanged meteoric water as extension increased fault permeability in the upper crust. Within a few kilometers of the surface, fluids were diverted by structural and stratigraphic aquitards into reactive host rocks, where they sulfidized host rock iron and deposited Au.

Sedimentary rock-hosted disseminated Au deposits in other parts of the world exhibit many similarities to Nevada Carlin-type Au deposits, yet no district has been discovered anywhere else that approaches Nevada's Au productivity. The deposits found in other parts of the world are products of diverse, well-recognized, hydrothermal systems (e.g., low-sulfidation epithermal, porphyry Cu-Mo-Au, reduced intrusion-related epizonal orogenic, and sedimentary exhalative or sedex). Of these, the deposits in southern China are remarkably similar to Nevada Carlin-type deposits and are interpreted to have formed where metamorphic fluids reacted with wall rocks and local meteoric water.

Introduction

Carlin-type Au deposits are largely replacement bodies with visually subtle alteration dominated by decarbonatization of silty carbonate host rocks, and they contain Au in solid solution or as submicron particles in disseminated pyrite or marcasite. The deposits occur in clusters or along trends (Fig. 1) and exhibit both structural and stratigraphic controls. They were first recognized as a new deposit type after discovery of the Carlin deposit in 1961, even though some deposits had been mined since the early 1900s.

Current production from Carlin-type deposits dominates Au production in the country and constitutes ~9 percent of world production, making the United States one of the

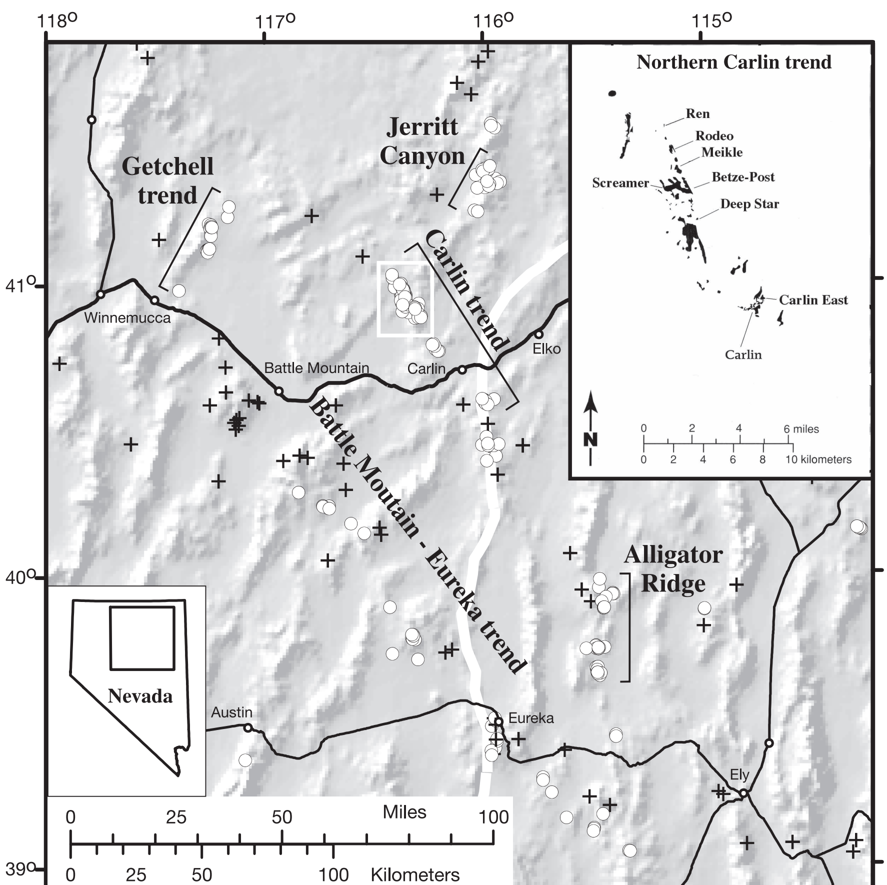


Fig. 1. Digital elevation model of northern Nevada, showing locations of major mineral belts and districts, Carlin-type deposits (circles; Peters et al., 2003), other significant Au, Ag, Pb, Zn, or Cu deposits (crosses; Long et al., 2000), eastern limit of the Roberts Mountain allochthon (bold white line; Crafford and Grauch, 2002), cities (small circles), and highways (black lines). Inset (white box) shows the distribution of Carlin-type deposits in the northern Carlin trend (Thompson et al., 2002). Figure provided by D. Sweetkind, U.S. Geological Survey, and modified from Hofstra et al. (2003).

leading Au-producing countries in the world (Nevada Bureau of Mines and Geology, 2004). Although several Carlin-type Au deposits in Nevada are world class (>100 t Au) and giant (>250 t Au), total contained Au and ore grades vary widely across districts and within deposits (Fig. 2). Exploration drilling through cover has greatly expanded the size of several districts and, along with developed mines, shows that deposits occur in camps or linear groups (trends) and that small districts may have significant resources at depth. Ten deposits in the Carlin, Getchell, and Battle Mountain-Eureka trends contain more than 5 million ounces (Moz) of Au and four deposits contain more than 10 Moz (Fig. 2). The Carlin trend contains the largest endowment of Au identified to date for such deposits in Nevada, and production has now exceeded 50 Moz (Nevada Bureau of Mines and Geology, 2004).

Despite the huge production and more than forty years of mining and research, a generally accepted genetic model for the deposits has yet to emerge (e.g., Joralemon, 1951; Hausen and Kerr, 1968; Bagby and Berger, 1985, Radtke, 1985; Seedorff, 1991; Kuehn and Rose, 1995; Ilchik and Barton, 1997; Hofstra and Cline, 2000; Emsbo et al., 2003). Development of a comprehensive genetic model has been hampered by widely conflicting data and, until recently, lack of agreement on the age of the deposits. Field relationships and conventional geochronological studies (App. Tables A1, A2) now indicate that the deposits formed during a narrow time interval during the mid to late Eocene, between about 42 and 36 Ma (App. Table A3; Hofstra et al., 1999; Tretbar et al., 2000; Arehart et al., 2003). Establishing this timing was critically important as the tectonic regime during deposit formation can now be incorporated into a genetic model. Sometime during the middle Eocene, northern Nevada began to undergo a transition from contractional to extensional deformation of the upper crust with associated calc alkaline magmatism (App. Fig. A1). This transition is considered instrumental to deposit formation.

Current models of deposit formation incorporate (1) meteoric water circulation (Ilchik and Barton, 1997; Emsbo et al., 2003), (2) epizonal plutons that contributed heat and possibly fluids and metals (Sillitoe and Bonham, 1990; Henry and Ressel, 2000; Johnston and Ressel, 2004), and (3) deep metamorphic fluids ± magmatic fluids (Seedorff, 1991; Hofstra and Cline, 2000). Much of the controversy regarding the genesis of Carlin-type Au deposits results from the fact that within some of the districts and trends and at other locations in Nevada, there are sedimentary rock-hosted disseminated Au deposits that have some attributes in common with, and other features that are distinct from, Carlin-type deposits (e.g., Hofstra and Cline, 2000; Hofstra, 2002).

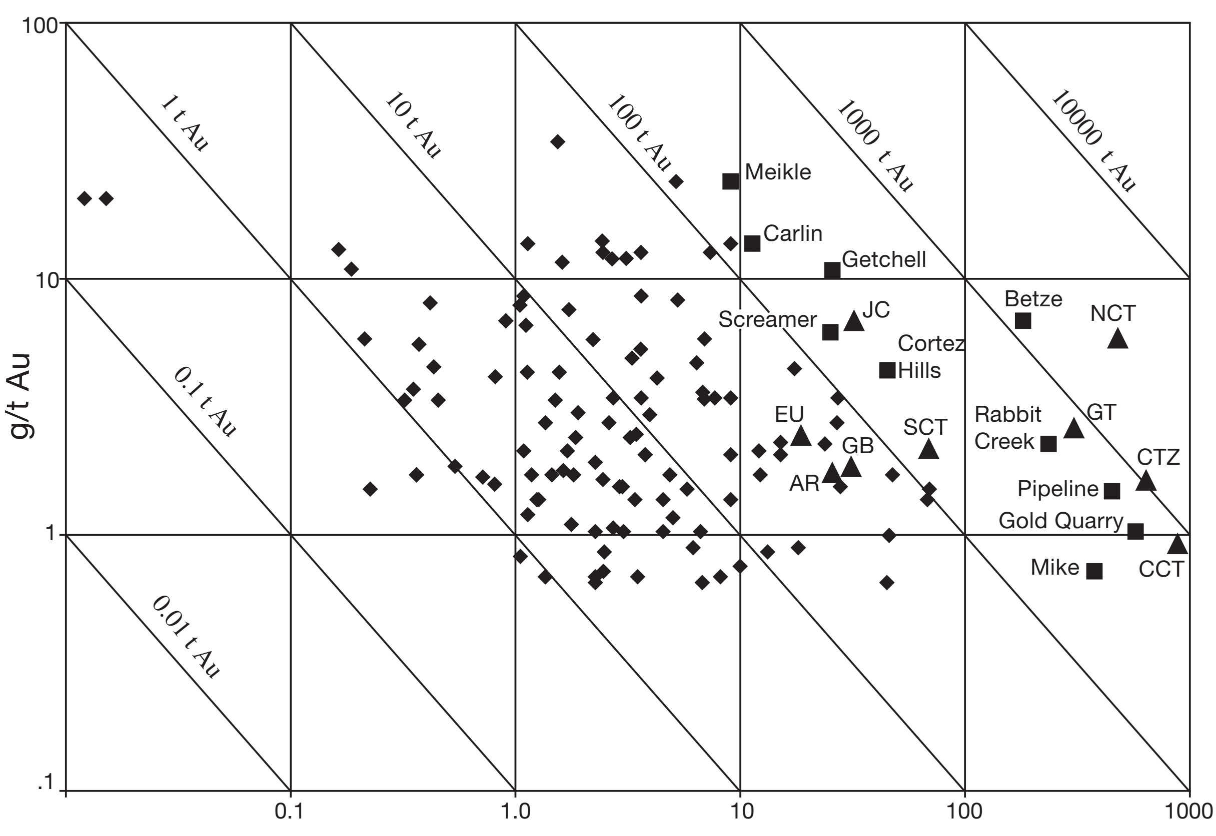


FIG. 2. Ore grade (g/t Au) vs. metric tons of ore for Carlin-type deposits (small diamonds), deposits containing greater than 5 Moz (155.5 t) gold (squares), and major districts (triangles). AR = Alligator Ridge district, CCT = central Carlin trend; CTZ = Cortez district in Battle Mountain-Eureka trend (BMET), EU = Eureka district in BMET, GB = Gold Bar district in BMET, GT = Getchell trend, JC = Jerritt Canyon district, NCT = north Carlin trend, SCT = south Carlin trend.

We believe that these sedimentary rock-hosted disseminated Au deposits are products of several well-recognized and distinctly different types of hydrothermal systems (Table 1; A. Hofstra and P. Emsbo, unpub. synthesis, 2004). Of these, distal disseminated deposits associated with porphyry Cu-Mo-Au systems (Cox and Singer, 1990, 1992) have frequently been called upon to explain Carlin-type deposits (e.g., Sillitoe and Bonham, 1990; Johnston and Ressel, 2004). These intrusion-related deposits exhibit many of the characteristics of Carlin-type Au deposits, but they have higher Ag and base metal concentrations, form from higher temperature and higher salinity fluids, and have clear spatial and genetic relationships with porphyry systems (Cunningham et al., 2004). Outside of Nevada, sedimentary rock-hosted disseminated Au deposits are concentrated in southern China (Li and Peters, 1998, Peters, 2002) but also occur at scattered locations around the world. Although the tectonic settings are different, the sedimentary rock-hosted disseminated Au deposits in the West Qinling belt and Dian-Qian-Gui area of southern China are most similar to Nevada Carlin-type deposits. Much of the controversy over whether or not Carlin-type deposits, distal disseminated deposits, and other sedimentary rock-hosted disseminated Au are essentially the same rather than discrete deposit types results from the assumption that there is only one way to produce a deposit with Carlin-type attributes, which we believe to be incorrect. This debate is not merely academic, as the various deposit types (Table 1) have different endowments of Au and will lead exploration geologists into different geologic environments in search of new discoveries.

This paper builds on earlier studies by incorporating new data and considering critical geologic features of Carlin-type deposits identified during detailed comparisons of the five major trends and districts in Nevada (Carlin, Getchell, Battle Mountain-Eureka, Jerritt Canyon, Alligator Ridge; Tables 2, A1, A2). A review (Hofstra and Cline, 2000) benefiting from knowledge of the age of the deposits evaluated available data in search of a best-fit genetic model but concluded that no single model could accommodate all credible data. Our current goal is to derive, through evaluation of new data and comparison of districts, a genetic model that can be used as a baseline for comparison with other sediment-hosted disseminated Au deposits around the world.

Another goal of this paper is to relate details of deposit geology to the tectonic evolution of northern Nevada. Why are major economic deposits of this type restricted to this area? Has a lengthy and unique geologic history or particular geologic event contributed to their formation? In comparing the geology of districts and trends we are struck by the consistency of geologic features at scales from regional to submicroscopic. These consistent characteristics, the relatively small size of the region that contains the deposits, and the now widely accepted narrow time frame for Au deposition that coincides with incipient extension and widespread magmatism, convince us that deposits in all districts are genetically linked and formed in response to crustal-scale orogenic processes that operated across northern Nevada.

Our study leads us to a genetic model that incorporates features proposed in earlier models and calls upon deep crustal processes to generate some components of a primitive ore fluid. Some ore-fluid components may have been sourced in the lower crust, where they initially accumulated and began their lengthy ascent toward the surface. Metamorphic ± mag-

TABLE 1. Deposits with Geologic Features Similar to the Nevada Carlin-Type Gold Deposits

Deposit type	Examples	References
Distal disseminated deposits associated with porphyry Cu-Mo-Au systems (also Carlin-like deposits) Seedorff (1991)	Lone Tree and Marigold, Nevada	Theodore (2000), McKibben and Theodore (2002)
	Melco and Barneys Canyon, Utah	Presnell and Parry (1996), Gunter and Austin (1997), Cunningham et al. (2004)
	Star Pointer, Nevada	Smith et al. (1988)
	Jeronimo and Silica del Hueso, Chile	Colley et al. (1989), Gale (2000), Simian and Hitzman (2000)
	Bau, Malaysia	Percival et al. (1990), Sillitoe and Bonham (1990, Schuh (1993), Percival and Hofstra (2002)
	Mesel, Indonesia	Turner et al. (1994), Garwin et al. (1995), Corbett and Leach (1998), Turner et al. (2002)
	Zarshuran and Agadarreh, Iran	Daliran et al. (1999, 2002), Mehrabi et al. (1999), Asadi et al. (2000), Lescuyer et al. (2003)
Distal disseminated deposits associated with reduced intrusion-related systems	Bald Mountain, Nevada	Hitchborn et al. (1996), Nutt et al. (2000), Nutt and Hofstra (2002)
	True North, Alaska and Brewery Creek, Yukon	Thompson and Newberry (2000), Flanigan et al. (2000), Poulsen (1996)
Epizonal and disseminated orogenic deposits	Fosterville, Victoria and Reefton, New Zealand	Beirlein and Maher (2001), Bierlein et al. (2002)
Sedex deposits	Rodeo, Nevada	Emsbo et al. (1999, 2002), Emsbo (2000)
Hybrid deposits with sedex and stratiform ore partially overprinted by Carlin-type mineralization	Meikle (footwall dolomitization) and Rodeo (stratiform), Nevada	Emsbo et al. (2003)

TABLE 2. Features Characteristic of Nevada Carlin-Type Gold Deposits

Deposit features	N Carlin trend	Central Carlin trend	S Carlin trend	Battle Mtn-Eureka trend	Getchell trend	Jerritt Canyon	Alligator Ridge
Pre-Eocene structural and stratigraphic architecture	XX	XX	XX	XX	XX	XX	XX
Underlain by Archean or thinned and mixed Paleoproterozoic and Archean transitional crust	XX	XX	XX	XX	XX	XX	XX
Underlain by thick Neoproterozoic to Early Cambrian rift-related clastic rocks	X	X	X	X	X	X	X
Location east of or near continental margin	XX	XX	XX	XX	XX	XX	XX
Proximal to regional thrust fault	XX	XX	XX	XX	XX	XX	-
Proximal to reactivated rift structures	X	X	X	X	X	X	?
High-angle northwest and northeast structures control ore	XX	XX	XX	XX	XX	XX	XX
Low-angle structures control ore	XX	XX	-	XX	XX	XX	-
Rheologic contrast around older stock controls ore	XX	XX	-	XX	XX	-	-
C- and pyrite-rich silty limestone or limey siltstone host rocks	XX	XX	XX	XX	XX	XX	XX
Proximal coeval igneous rocks	X	X	X	X	-	X	-
Characteristic alteration present	XX	XX	XX	XX	XX	XX	XX
Characteristic ore and late ore minerals present	XX	XX	XX	XX	XX	XX	XX
Au in arsenian trace element-rich pyrite or marcasite	XX	XX	XX	XX	XX	XX	XX
Deep magmatic ± metamorphic source identified for He, Pb, Nd, ±Sr	X	nd	nd	nd	XX	nd	nd
Magmatic ± metamorphic water identified in ore fluid	X	-	-	-	XX	-	-
Meteoric water identified in ore fluid	XX	XX	nd	nd	-	XX	nd
Magmatic ore S source identified	X	-	-	-	X	-	-
Sedimentary ore S source identified	XX	XX	XX	XX	-	XX	XX
Postore oxidation	XX	XX	XX	XX	XX	XX	XX

Notes: XX = important deposit feature, X = observed feature, - = not present, nd = no data, ? = unknown

matic processes may have generated or contributed to the original deep ore fluid, which likely evolved significantly as various components, including Au, were added and removed along fluid pathways.

Pre-Eocene Tectonic Framework

The superposition of tectonic events extending back to the Proterozoic in northeastern Nevada produced geologic features that are common to all Nevada Carlin-type deposits. These pre-Eocene events, namely continental rifting followed by several compressional orogenies, built a premineral architecture of steeply dipping fault and fracture meshes that acted as fluid conduits. The resulting low-angle faults combined with impermeable rock units formed aquitards to inhibit fluid ascent, causing the steep faults to channel fluids into permeable and reactive calcareous host rocks (App. Fig. A2).

Geologic history

During the assembly of the Laurentia supercontinent, Paleoproterozoic terranes were accreted to the Archean Wyoming craton. The suture zone, which is well defined in the Cheyenne belt of Wyoming, has a westward trend into northern Nevada where it becomes a broad zone of intermixed Paleoproterozoic and Archean rocks (Lush et al., 1988; Tosdal et al., 2000; App. Fig. A3). In the Meso- and Neoproterozoic, rifting between 1.0 and 1.3 and 0.6 and 0.9 Ga separated Laurentia from an adjoining crustal block (Karlstrom et al., 1999; Timmons et al., 2001), thereby forming a progressively westward-thinning margin of continental crust (Tosdal et al., 2000). A westward-thickening wedge of Neoproterozoic and early Cambrian clastic rocks buried the thinned crystalline basement during the rift phase of extension (Stewart, 1972, 1980; Poole et al., 1992). Following active rifting of the continent, a miogeoclinal sequence developed on the trailing continental margin with deposition of interbedded carbonate and shale on the shelf. To the west, silty carbonate rocks were deposited along the continental slope.

During the Late Devonian and early Mississippian Antler orogeny, the Roberts Mountain thrust (Roberts et al., 1958; Stewart, 1980) formed as eugeoclinal siliciclastic and basaltic rocks were thrust eastward over the shelf-slope sequence along a fold and thrust belt (Fig. A2). Loading by the allochthon, which was up to several kilometers thick (E.L. Miller et al., 1992), onto the continental margin produced a foredeep basin in eastern Nevada that migrated eastward in front of the thrust belt. Early Mississippian synorogenic and Pennsylvanian postorogenic sedimentary rocks filled the foredeep (Poole et al., 1992). Intermittent shortening and extension in the Pennsylvanian and through the Permian (e.g., the south-directed Humboldt orogeny of Ketner, 1977; Theodore et al., 2004) was followed by eastward emplacement of the Golconda allochthon during the Sonoma orogeny in the Early Triassic.

An east-dipping subduction zone was established along the western margin of North America by the Middle Triassic. North-central Nevada lay east of the main magmatic arc, which is represented by the Mesozoic granitic batholiths of the Sierra Nevada Range. Magmatism in north-central Nevada began with emplacement of Middle Jurassic, backarc volcanic-plutonic complexes and lesser lamprophyre dikes. Coeval shortening produced a north-trending belt of east-verging folds and thrusts in eastern Nevada (Elko orogeny of Thorman et al., 1991). Plutons evolved from I-type granitoids in the Early Cretaceous to S-type peraluminous granites in the Late Cretaceous, as the crust was progressively thickened (Barton, 1990) during the Late Cretaceous Sevier and Laramide orogenies (Burchfiel et al., 1992). At about 65 Ma magmatism shifted eastward into Colorado and did not resume in Nevada until about 42 Ma (Lipman et al., 1972; Hickey et al., 2003b). The area of the northern Carlin trend underwent the last significant period of exhumation in the Late Cretaceous and had cooled to <60° to 70°C by ~60 Ma (corresponding to a <2- to 3-km crustal depth; Hickey et al., 2003b).

Pre-Eocene architectural controls on Carlin-type mineralization

Nature of underlying crust: The Carlin-type deposits of Nevada are restricted to the area underlain by Archean crust (northern Carlin trend and Jerritt Canyon) or by the thinned and mixed Paleoproterozoic and Archean transitional crust (Battle Mountain-Eureka trend and Getchell district) lying between the rifted edge of the continental margin and largely intact crystalline basement rocks (Table 2; Fig. A3). The deposits are furthermore localized in areas underlain by thick Neoproterozoic to Early Cambrian rift-related clastic rocks (Table 2; Seedorff, 1991), which are thought to have supplied metals and S to the deposits (Seedorff, 1991; Vikre, 2000), a proposal that is one interpretation of Pb isotope data for ore-stage sulfide minerals (Tosdal et al., 2003).

Basement faults: The alignment of Carlin-type deposits in north-central Nevada reflects major basement fault fabrics, which were established during Neoproterozoic rifting (Roberts, 1966; Tosdal et al., 2000; Grauch et al., 2003). The majority of Carlin-type deposits (Tables 2, A1) are controlled by 330°- to 350°- and 290°- to 310°-trending faults and folds that are interpreted to have been localized by the fault fabric established during continental rifting. The basement-penetrating fault systems are critical in that they served to influence subsequent patterns of sedimentation, deformation, magmatism, and hydrothermal activity and may have acted as collecting points and conduits for hydrothermal fluids sourced from deeper crustal levels.

Favorable host rocks: The miogeoclinal shelf-slope sequence contains a large mass of thin-bedded, pyritic, and carbonaceous silty dolomitic limestone and marl, which are the primary host rocks in most of the Carlin-type deposits, including almost all of the >5-Moz deposits (Table 2; Figs. 3, 4). Carbonate dissolution during ore formation resulted in volume loss and permeability increase; the quartzose silt and partially leached (sanded) dolomite grains maintained an open framework accessible by auriferous fluids.

Transitional facies such as the base of a carbonate bank or basin-margin slope have high initial porosity due to soft-sediment deformation, gravity-driven debris-flow deposits (see below), and diagenetic recrystallization of fossils and turbidites. Other Carlin-type deposits are localized by major changes in rock composition, where thin-bedded silty rocks occur over clean, thick-bedded limestone and are capped by siliciclastic rocks (e.g., Rain, Alligator Ridge, Horse Canyon).

Synsedimentary faults, mineralization, and diagenesis: Paleozoic synsedimentary faults caused abrupt facies and thickness changes and served to localize syngenetic Au and barite deposits (Emsbo et al., 1999, 2003) and rift basalts. These faults are parallel to, and probably link with, the underlying basement fault fabric beneath the sedimentary sequence. The faults reflect reactivation during sedimentary loading or during extension. Soft-sediment deformation and debris-flow breccias on slopes led to increased permeability and primary porosity along these zones. The best example occurs over a distance of less than 800 m and is located between the Meikle and Betze-Post deposits (Fig. 4c). There a northwest-trending zone separates shallow marine massive, oolitic, fossiliferous limestones of the Devonian Bootstrap Limestone, which is host to the Meikle deposit, from laminated muddy limestones and debris flow breccias of the time-equivalent Popovich Formation which hosts Betze-Post (Armstrong et al., 1998; Griffin, 2000). At Getchell, within the Ordovician Comus Formation, a sequence of pillow basalts and underlying sedimentary breccias has a sharp 290°- to 300°-trending southern margin that is an important ore control to the underground Turquoise Ridge deposit. The margin occurs along the northern limb of a monocline interpreted to have formed by syndepositional, normal reactivation of an underlying north-dipping, west-northwest–trending basement fault (Placer Dome Exploration, pers.commun., 1999). Similar synsedimentary rift basalts at Twin Creeks resulted in sills and flows that formed important shallow-dipping aquitards that promoted lateral fluid flow in reactive calcareous rocks in the footwall (Bloomstein et al., 1991; Stenger et al., 1998).

Normal faulting during deposition of the miogeoclinal sequence also led to local shallow-water conditions and subaerial exposure and dolomitization in a sabkha-environment (Armstrong et al., 1998) or potentially caused karst formation and preore collapse breccias. These environments resulted in intrinsic permeability and secondary porosity in the host rocks that enhanced fluid flow and provided sites for younger Au deposition.

Antler orogeny: The main host unit for Carlin-type deposits is the lower plate to the Roberts Mountain thrust. Most giant deposits lie within 100 m of a thrust (or its projection), which in many districts is the Roberts Mountain thrust. Smaller deposits (<1 Moz) occur east of the Roberts Mountain allochthon (e.g., Alligator Ridge) where regional Paleozoic thrust faults are absent. The thrust faults are important in that they formed a regional aquitard to any upward fluid flow; nonreactive, fine-grained siliciclastic rocks with less inherent rock permeability were thrust over permeable carbonate stratigraphy. Aquitards may force fluids laterally away from the steeply dipping fluid conduits into permeable reactive rocks. Otherwise, fluids may deposit Au over a large vertical interval (>1 km) or may discharge at, or near, the surface, as appears to be the case in the Alligator Ridge district (Nutt and Hofstra, 2003).

Folds and inversion of normal faults: Paleozoic shortening led to imbrication in thin-skin fold and thrust belts of the Paleozoic stratigraphy. Cutting across the thin-skinned fold and thrust belts are local and narrow belts of thick-skinned folds and reverse faults with significantly less overall shortening. The thick-skinned belts have northwest trends (Madrid and Roberts, 1991).

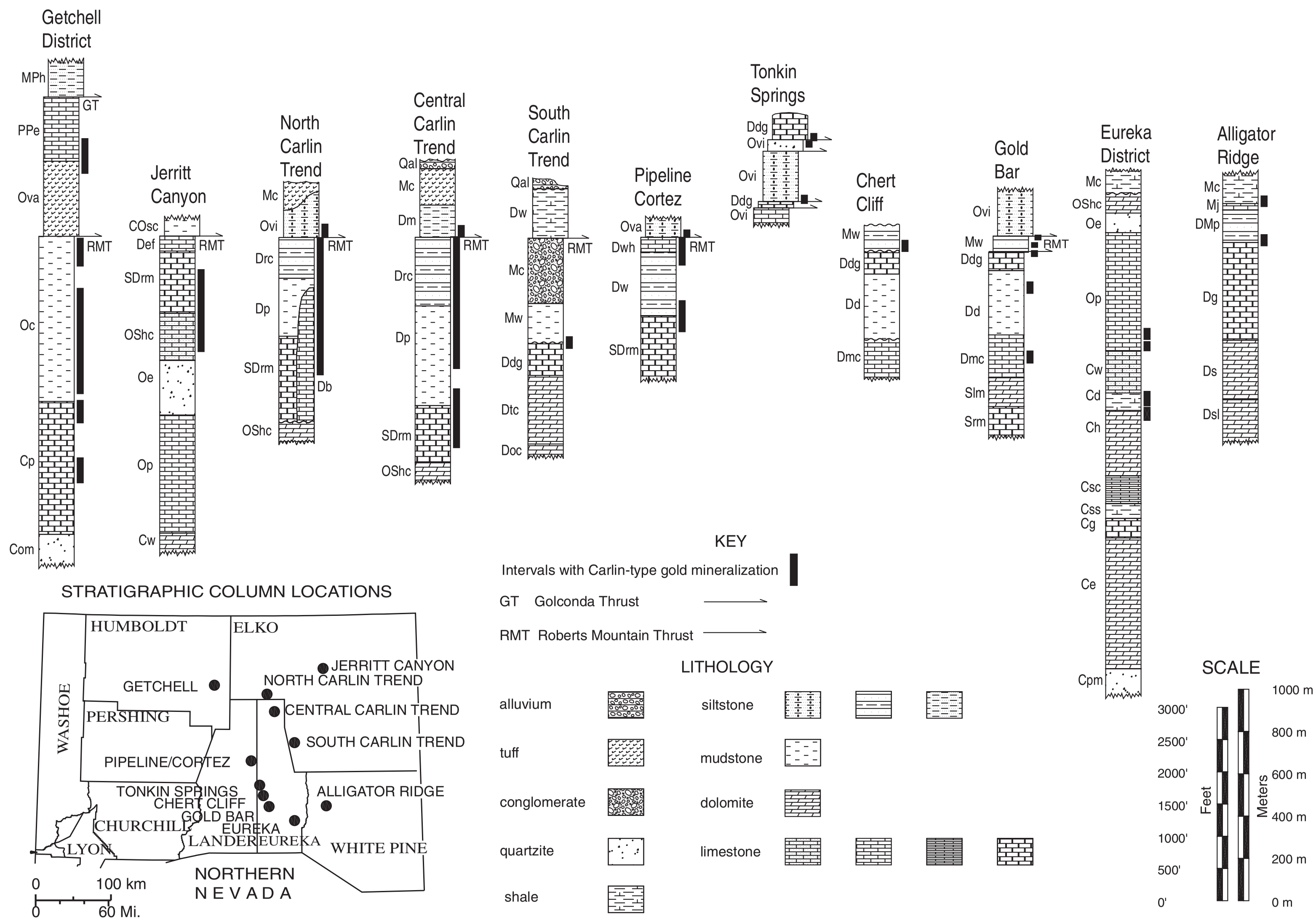


FIG. 3. Stratigraphic columns for districts containing Carlin-type gold deposits. Bars indicate the location of Carlin-type gold mineralization in each section. Formation key: Cd = C Dunderburg, Ce = C Eldorado, Cg = C Geddes, Ch = C Hamburg, Com = C Osgood Mountain, COsc = C-Ord Snow Canyon, Cp = C Preble, Csc = C Secret Canyon, Cpm = C Prospect Mountain, Cw = C Windfall, Db = Dev Bootstrap, Dd = Dev Denay, Ddg = Dev. Devils Gate, Def = Dev Eastern Facies, Dg = Dev Guilmette, Dm = Dev Marys Mountain, Dmc = Dev McColley Canyon, DMp = Dev-Miss Pilot, Doc = Dev Oxyoke Canyon, Dp = Dev Popovich, Drc = Dev Rodeo Creek, Ds = Dev Simonson, Dsl = Dev Sevy Laketown, Dtc = Dev Telegraph Canyon, Dwh = Dev Wenban Horse Canyon, Dw = Dev Wenban (Pipeline-Cortez), Dw = Dev Woodruff (S Carlin trend), Mc = Miss Chainman (Eureka, Alligator Ridge), Mc = Mio Carlin (North and Central Carlin trend), Mj = Miss Joana, MPh = Miss-Perm Havallah, Mw = Miss Webb, Oc = Ord Comus, Oe = Ord Eureka, Op = Ord Pogonip, OShc = Ord-Sil Hanson Creek, Ova = Ord Valmy, Ovi = Ord Vinini, PPe = Penn-Perm Etchart, Qal = Quat Alluvium, SDrm = Sil-Dev Roberts Mountain, Slm = Sil Lone Mountain, Srm = Sil Roberts Mountain. Data from French et al. (1996), Teal and Jackson, (1997), Vikre and Maher (1996), G.A. Ghidotti and M.D. Barton, writ. commun. (1999), McLachlin et al. (2000), Yigit (2001), Bettles (2002), Harlan et al. (2002), Jory (2002), Longo et al. (2002), Emsbo et al. (2003), Nutt and Hofstra (2003), Peters et al. (2003), D. Reid, U.S. Gold Corp, writ. commun. (2003), Placer Dome, unpub. data (2004).

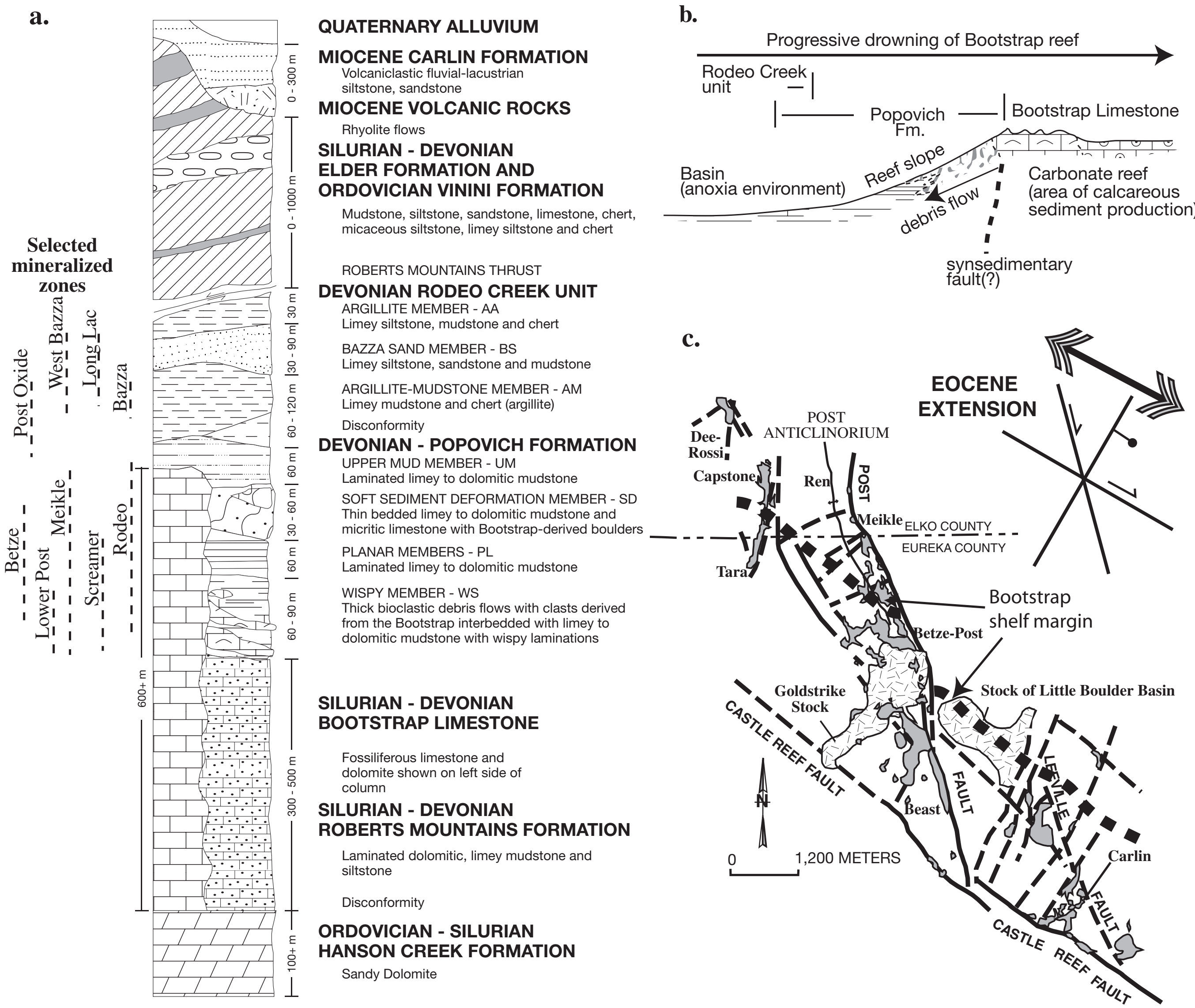


FIG. 4. a. Stratigraphic column through the Goldstrike mine area, showing the localization of the small structural controlled orebodies within the Rodeo Creek unit and the concentration of high gold grades and large structural and stratigraphic controlled orebodies in the Popovich Formation (from Volk et al., 1997; Bettles, 2002). b. Paleoenvironment of the host rocks in the northern Carlin trend, showing distribution of carbonate facies and their formations. The Devonian rocks in the northern Carlin trend represent the reef foreslope facies and the progressive drowning of the Bootstrap reef by the Rodeo Creek unit. c. Location of Bootstrap shelf margin. Also shown is the predicted slip pattern on faults that would be active during Eocene extension.

The fold belts and windows have been attributed to Mesozoic shortening (Madrid and Roberts, 1991; Peters, 2000) or igneous doming (Roberts, 1960). Other evidence suggests a Paleozoic age. In the Carlin trend, northwest-trending folds predate the Jurassic Goldstrike stock (Moore, 2001). In the Piñon Range around the Rain deposit, northwest-trending folds and reverse faults are localized along inverted Early Mississippian normal faults that disrupted the north-striking Antler fold and thrust belt prior to the deposition of late Mississippian postorogenic strata (Fig. 5; Tosdal, 2001). These observations suggest that the narrow northwest-trending fold belts are at least in part the result of inversion of the north-northwest– and west-northwest–striking basement and Paleozoic normal faults, most likely in the late Paleozoic. Other examples of inverted Paleozoic normal faults include (1) the north-northwest–striking Getchell fault and the hanging-wall, asymmetric, west-verging anticlines of folded growth sequences as interpreted from seismic sections; (2) the north-northwest–striking Post fault system in the Carlin trend and west-verging Ernie anticline (Baschuk, 2000); and (3) the west-northwest–striking "flower structures" in the West Bazza pit at Betze-Post (Lauha, 1998) and Rain (Williams et al., 2000). Significantly, these fabrics were also reactivated in the Mesozoic and the Early Tertiary (see below).

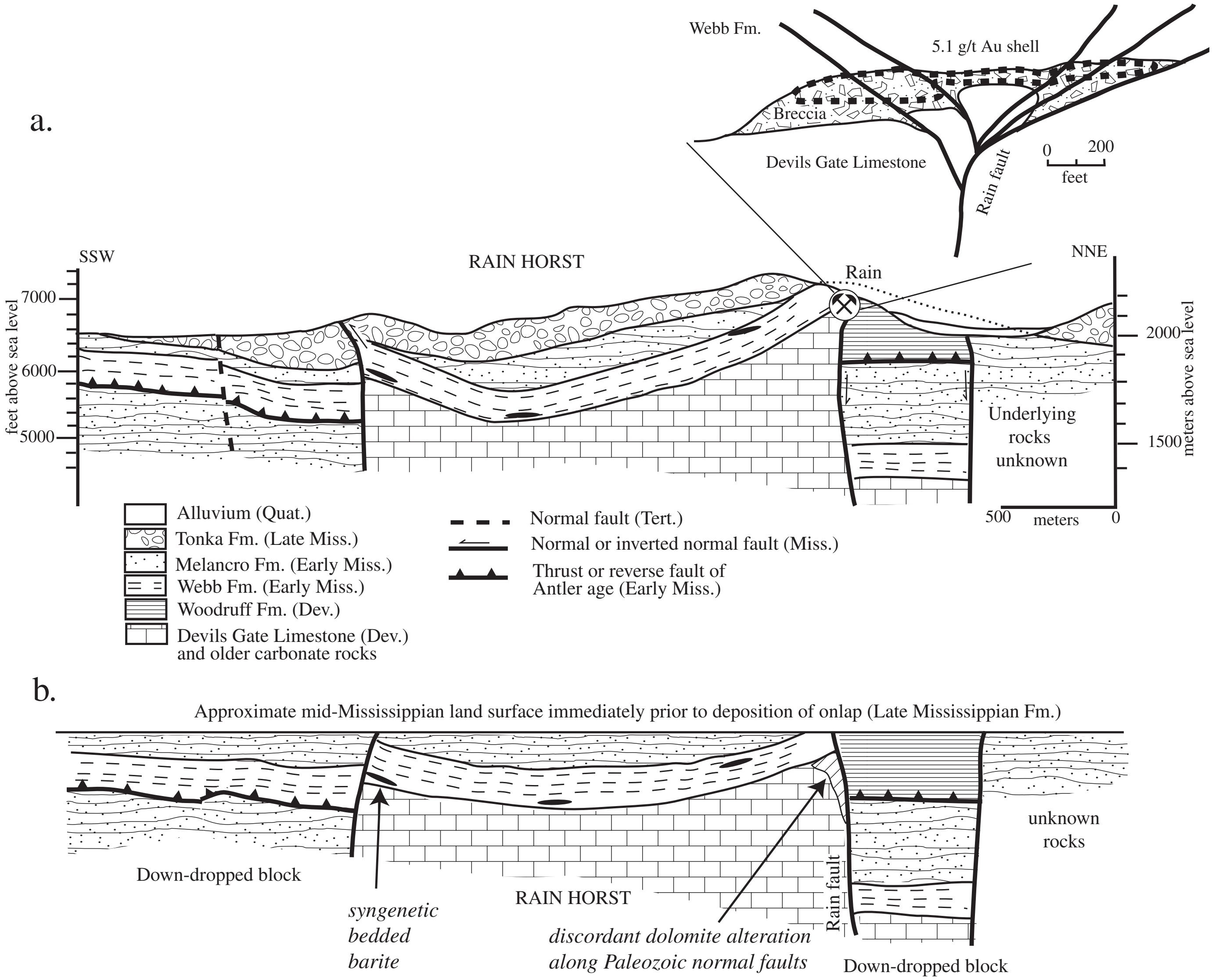
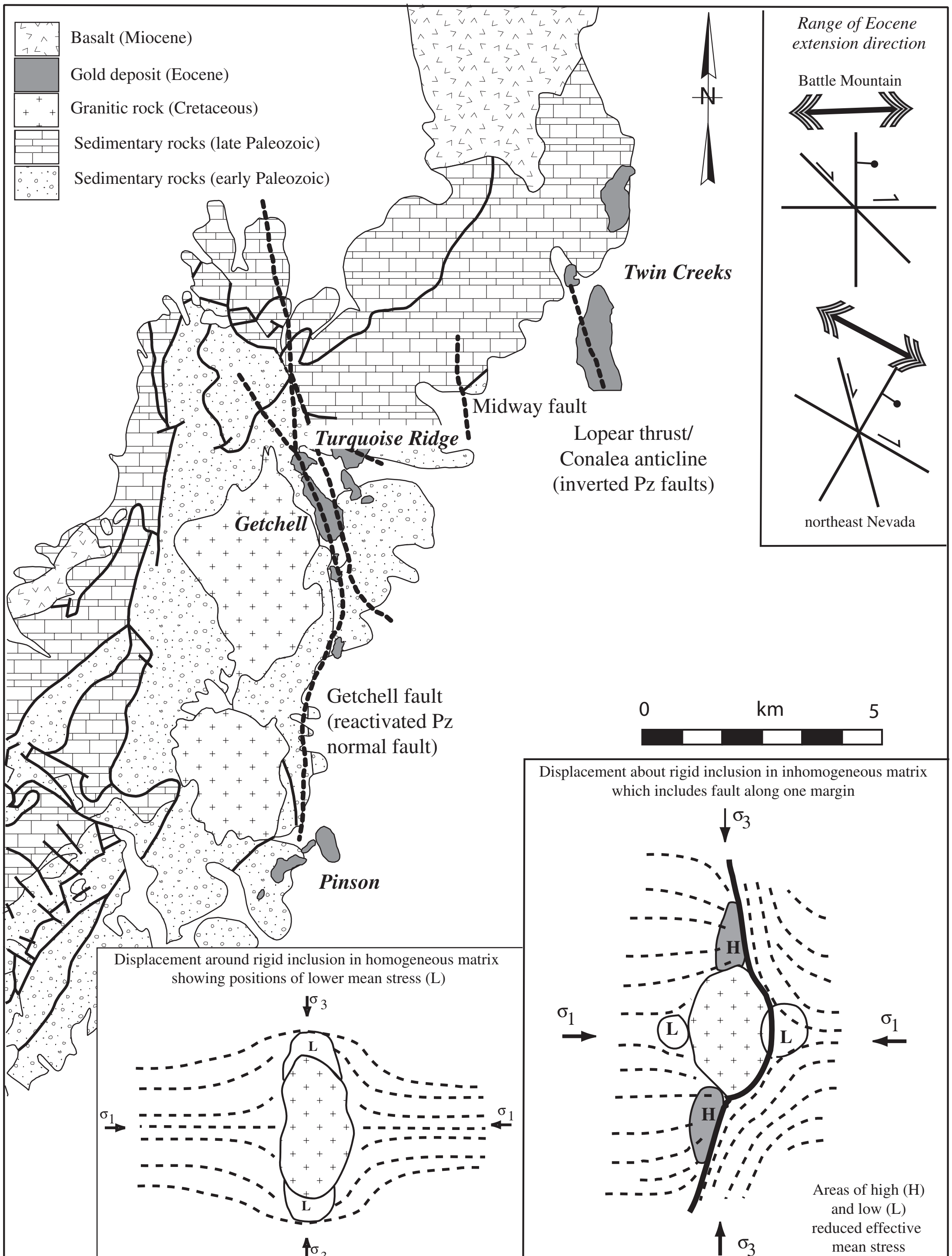


FIG. 5. Simplified (a) cross section and (b) restored cross sections looking west-northwest (~300°) and normal to the Rain fault, northern Piñon Range, Nevada. The Late Mississippian and Early Pennsylvanian Tonka Formation (Trexler et al., 2003), which is part of the post-Antler overlap sequence, was deposited on a paleosurface that beveled different levels of the Early Mississippian Antler fold and thrust belt as well as northwesterly striking Early Mississippian normal faults that cut the fold and thrust belt. The Mississippian normal faults shown in panel b were inverted during late Paleozoic (?) southward-directed shortening as steeply to moderately dipping reverse faults. The Rain horst (Mathewson, 2001) is an elevated block of Devonian Devils Gate Limestone, and older subjacent rocks, that forms either the autochthonous footwall to the fold and thrust belt or represents a structurally deeper thrust sheet as is seen in the Railroad area in the central Piñon Range (Tosdal, 2001). The margins of the Rain horst are constrained by gravity data (Mathewson and Beetler, 1998). Inset diagram shows diverging inverted normal faults and subsidiary accommodation reverse faults at a dip change in the Rain fault, which for the most part dips steeply (Longo et al., 2002). Premineral solution collapse and tectonic breccia lies in the hanging wall of the Rain fault; premineral dolomitized limestone lies in the Devils Gate Limestone below the orebodies. Syngenetic barite lenses are shown schematically in the Webb Formation. The Rain deposit as shown by the 5.1-g/t Au ore shell lies at the top of the breccia and below the calcareous siltstones that compose the early Mississippian Webb Formation. The Devonian Woodruff Formation, which is correlative to the Rodeo Creek unit in the northern Carlin trend (see Fig. 3), consists of organic-rich (>3% C) metalliferous black shale.

Inversion resulted in structural culminations (doubly plunging anticlines, domes), some of which subsequently acted as depositional sites for auriferous fluids. Once exhumed these culminations are now the erosional windows through the siliciclastic rocks of the Antler allochthon. The Lynn window in the Carlin trend is the best-known example.

Preore (Mesozoic) stocks: The three largest Carlin-type districts (northern Carlin trend, Getchell, Pipeline-Cortez) are spatially associated with Jurassic or Cretaceous plutons that are preore. Small satellite stocks and numerous dikes are also common. These bodies were emplaced along the existing fault systems in each district, which subsequently controlled flow of the ore fluids. The widespread presence of intrusions in the abundantly faulted fabric of the districts indicates that the permeability structure provided by the fault and fracture architecture around the Carlin-type deposits was in place prior to ore formation.

The Au deposits of the region are common along the margins of these plutons (Fig. 6) or at the rheologic contrasts provided by the contact between the metamorphic hornfels aureoles and the unmetamorphosed country rocks (Tables 2, A1; e.g., Deep Star, Heitt et al., 2003). The stocks and their surrounding aureoles essentially acted as rigid bodies that controlled and pinned postintrusion deformation. For example, in the Eocene during ore formation, limited extension reactivated favorably with oriented faults, which were largely the north-northeast– to northwest-striking faults. However, because of the low magnitudes of strain, slip on these faults near the older Mesozoic intrusions was forced around the rigid stocks and contact aureoles (Fig. 6) or stepped from one



fault to another (e.g., Deep Star deposit, Heitt et al., 2003; Post fault in Betze-Post, Moore 2001). The net effect was a complex fracture mesh adjacent to the plutons that includes small-scale faults with vertical and horizontal slip vectors, all of which have been recognized in structural studies (e.g., Deep Star; Heitt et al., 2003).

The heterogeneous strain around the stocks essentially formed extensive damage zones marked by increased fault and fracture density. These sites of enhanced vertical permeability are located asymmetrically around the margins of the intrusions (Fig. 6). The local reduced mean stress associated with the enhanced fracture density favored Au deposition in these sites. Intrusions locally resulted in the formation of reactive ferroan carbonates (Twin Creeks, Fortuna et al., 2003; Deep Star, Heitt et al., 2003), which likely facilitated ore deposition by sulfidation, as discussed below. In addition, sill-like apophyses extending from the Mesozoic stocks served as local aquitards that helped to develop high grades in their footwalls (e.g., Meikle, Volk et al., 1995; Getchell, Tretbar, 2004).

Eocene Setting for Carlin-Type Au deposits

The Cretaceous-Paleogene tectonic framework

The western margin of North America underwent near-continuous easterly convergence from the early Mesozoic to the end of the Oligocene (Engebretson et al., 1985). The Cretaceous Sevier-Laramide orogeny was a product of that convergence and northeastern Nevada is generally thought to have occupied the thickened core of the Sevier hinterland (Coney and Harms, 1984; D.M. Miller et al., 1992; Camilleri et al., 1997). From the Late Cretaceous to middle Eocene, the oceanic Farallon and Kula plates were spreading apart while subducting beneath North America (App. Fig. A4). The spreading ridge intersected the North American plate somewhere between British Columbia and Mexico (Engebretson et al., 1985), with the slab window produced by the subducting ridge passing northward through Nevada at the beginning of the Eocene, at ~54 Ma (Breitsprecher et al., 2003). The Farallon plate is thought to have had a subhorizontal geometry resting on the base of the western North American lithosphere (Dickinson and Snyder, 1978; Christiansen and Yeats, 1992).

High K calc-alkaline magmatism within northern Nevada began ~42 Ma and swept southward with time, culminating in Oligocene-Miocene volcanic activity in central-southern Nevada (Fig. A4; Armstrong and Ward, 1991; Seedorff, 1991; Henry and Boden, 1998) in response to the progressive removal or rollback of the Farallon plate and the reintroduction of hot asthenospheric mantle to the base of the North American lithosphere (e.g., Humphreys, 1995; Humphreys et al., 2003). Eocene volcanism was linked to short-lived periods of upper crustal extension and development of broad depressions filled with fluvial-alluvial and lacustrine sediments, volcaniclastic rocks, ash-flow tuffs, and lavas (Solomon et al., 1979; Axen et al., 1993; Potter et al., 1995; Gans et al., 2001; Rahl et al., 2002). Over much of the western cordillera, upper crustal extensional faulting was accompanied by midcrustal flow and the localized development of Oligocene and Miocene metamorphic core complexes. Extension became younger to the south, from the late Paleocene in British Columbia to Oligocene and Miocene in Nevada (Parrish et al., 1988; Gans et al., 1989; Foster and Fanning, 1997; Mueller et al., 1999; McGrew et al., 2000; Rahl et al., 2002). The onset of Eocene extension in northern Nevada has been linked to the removal of the Farallon plate from the base of North American lithosphere, either through gravity-driven collapse of the thickened Sevier hinterland (Jones et al., 1998; Liu and Shen, 1998; Rahl et al., 2002) or transient shear stresses at the base of the brittle crust as lower crustal flow was reestablished (Westaway, 1999).

The spatial and temporal overlap of Carlin-type deposits with the onset of Cenozoic volcanism and extension in northern Nevada suggests a fundamental link between these phenomena (Seedorff, 1991; Hofstra, 1995; Ilchik and Barton, 1997; Henry and Boden, 1998; Hofstra et al., 1999). A full understanding of the patterns of crustal fluid flow responsible for Au mineralization in northern Nevada between ~42 and 36 Ma is dependent upon more complete knowledge of the spatial and temporal relationships between regional extension and volcanism.

Eocene paleogeograpic and tectonic development of northern Nevada

Rocks exposed along the modern erosion surface in the Carlin-Jerritt Canyon region were exposed to a major phase

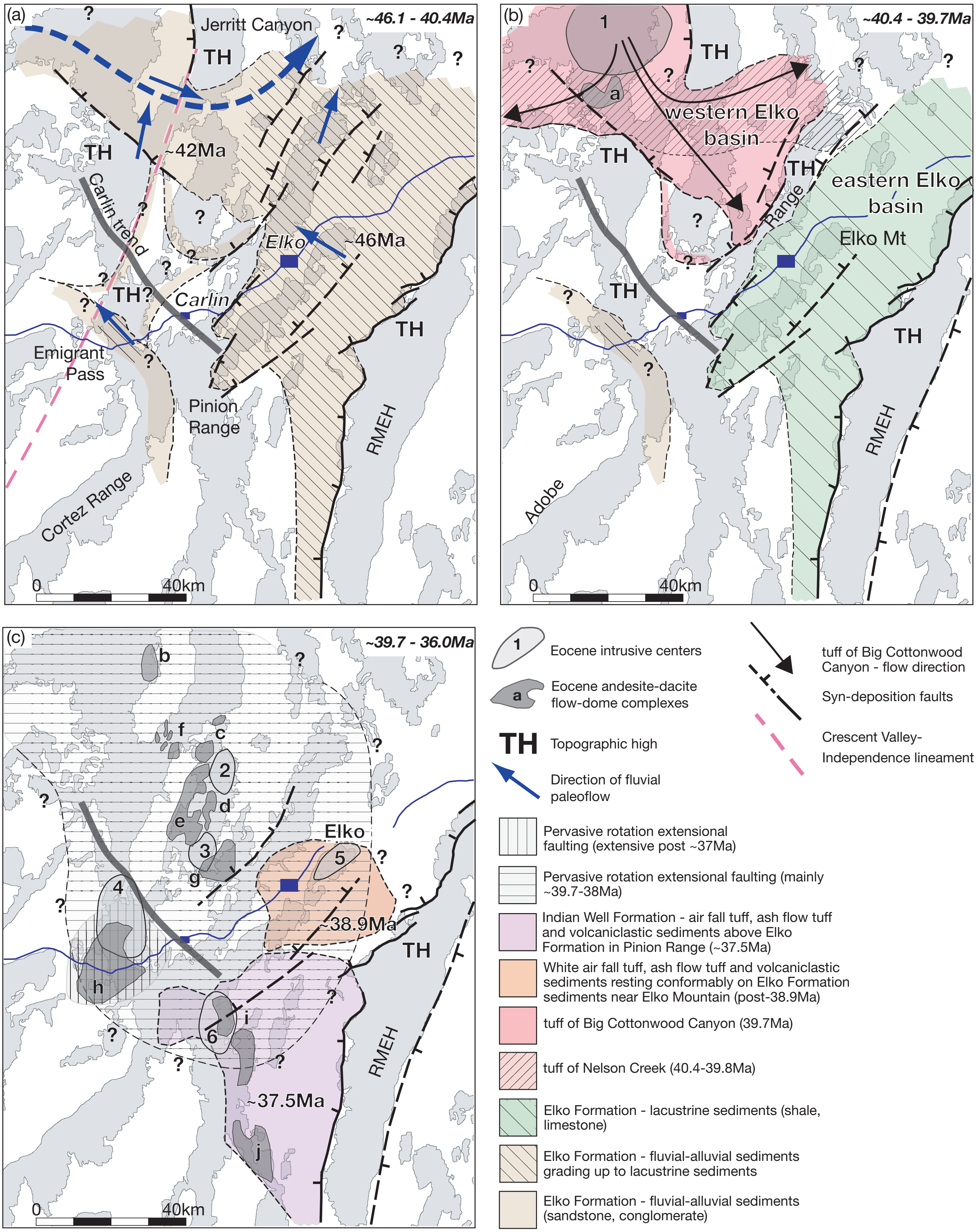
FIG. 6. Simplified geologic map of northern Osgood Mountains adapted from Hotz and Willden (1964), showing asymmetric distribution of Carlin type Au deposits around the Cretaceous Osgood Mountains stock. Also shown are interpreted Paleozoic normal faults. Inset diagrams in upper right shows the Reidel shear fracture and fault mesh that would fail under extension or extensional shear failure and thus be dilatant under the expected northeasterly to easterly extension in the Eocene. Extension directions are based on geologic information from northeastern Nevada (Tosdal and Nutt, 1999) and the Battle Mountain area (Theodore, 2000); no geologic constraints for the strain orientation in the Eocene is known for the Osgood Mountains and thus the range of orientations derived from geologic maps elsewhere in Nevada are shown. In detail, north- to northeast-striking faults would preferentially fail under extension or extensional shear in either orientation and thus be the favored sites of enhanced fluid flow. Inset diagrams in lower right are two-dimensional models of stress trajectories around rigid bodies in softer rocks (Ridley, 1993). Such rigid bodies, which in the case of the Carlin-type deposits are Mesozoic plutons, would perturb the stress field (heavy dashed lines), thereby giving rise to areas of high and low effective mean stress. Two cases are shown: one is where the rigid body lies in homogeneous rocks and no extant faults and the second where an older fault is pinned along the margins of the rigid body. Flow of hydrothermal fluids is enhanced in the areas of low effective mean stress whose relationship to the rigid body varies depending upon the structural architecture. Both models predict that the fluid flow would be distributed at discrete zones on the margins of the rigid body and not uniformly around the body. The model is supported by the distribution of gold deposits in the Osgood Mountains and also the northern Carlin trend (see Fig. 4c).

of exhumation and cooling to <60° to 70°C by the end of the Cretaceous (Hickey et al., 2003a, b). Subsequent low cooling rates (~1°C/m.y.), indicative of very low exhumation rates, suggest there was little or no surface-breaking tectonic activity in northern Nevada from the end of the Sevier orogeny to the middle Eocene (cf. McGrew et al., 2000). Such a history makes it likely that by the middle Eocene, northern Nevada was characterized by a low-relief landscape and an elevation similar to that of today (~1,600–2,800 m; Chase et al., 1998; Wolfe et al., 1998).

Beginning in the middle Eocene, numerous regional extensional basins began to develop across northern Nevada and into western Utah (e.g., Smith and Ketner, 1976; Potter et al., 1995; Rahl et al., 2002; Nutt and Hofstra, 2003). The largest and best studied of these is the Elko basin, which is present along the west side of the entire Ruby Mountains-East Humboldt Range as far west as the cluster of Carlin-type deposits that comprise the Carlin trend and Jerritt Canyon (Fig. 7; Smith and Ketner, 1976; Solomon et al., 1979; Henry et al., 2001; Haynes, 2003). The Elko basin developed as one or more large half-grabens above a shallow west-northwest–dipping growth fault on the west flank of the modern Ruby Mountains-East Humboldt Range between ~46 and ~37 Ma (Satarugsa and Johnson, 2000; Haynes, 2003). Synextensional fluvial-alluvial sedimentation (plus minor volcanic ash) began at ~46 Ma in the eastern part of the basin, proximal to the bounding growth fault. As net displacement across the fault accumulated, the basin expanded and fluvial sedimentation extended northwestward, where a broad, west-northwest– to east-west–trending paleovalley formed between the Carlin trend and Jerritt Canyon from ~42 Ma (Figs. 7, 8). From ~40.4 to 39.7 Ma, explosive volcanic eruptions from the Tuscarora volcanic field, and other unidentified centers, filled the paleovalley in the western part of the basin with varying thicknesses of pyroclastic and volcaniclastic rocks, effectively ending fluvial sedimentation (Figs. 8 and 9a; Hickey, 2003b). In contrast, lacustrine sediments continued to fill the eastern parts of the basin, suggesting that a topographic barrier had developed between the two parts of the basin by ~40 Ma (Fig. 7b, c). After 39.7 Ma, explosive volcanism in the western Elko basin ceased and pervasive rotational extensional faulting began to break up the western part of the basin. South of the Carlin trend in the Emigrant Pass area this phase of extension was initiated some time prior to ~38 Ma (Henry et al., 2001). The emplacement of shallow Eocene intrusions and the extrusion of small, commonly fault-controlled, andesite-dacite flow-dome complexes accompanied the pervasive rotational extension (Fig. 7c; Henry and Ressel, 2000; Ressel et al., 2000; Henry et al., 2001). Lacustrine sedimentation in the eastern Elko basin stopped sometime between ~38.9 to 38 Ma as rotational extensional faulting extended into the eastern part of the basin (Fig. 7c; Haynes, 2003). Rotational extension continued to at least ~38 Ma over much of the Elko basin and in its western parts, particularly the Emigrant Pass area, continued into the late Oligocene (Henry et al., 2001).

The distribution of Eocene ash-flow tuff units and underlying clastic and volcaniclastic rocks in the western Elko basin suggest that the northern Carlin trend occupied a relative topographic high on the southern side of a west-northwest– to east-west–trending paleovalley (Figs. 7b, 8, and 9a). Eocene units thin and pinch out to the south, toward the northern Carlin trend and do not crop out again until south of the Carlin trend, where a succession of Eocene fluvial-alluvial sedimentary rocks are overlain by ~38 to 37 Ma andesite flow-dome complexes without the intervening ash-flow tuffs present in the western part of the paleovalley (Fig. 7; Henry and Faulds, 1999; Henry and Ressel, 2000). A similar situation is present in the southern Independence Mountains, where Eocene ash-flow tuff and underlying clastic sediments pinch out southward to a location where Paleozoic rocks are

FIG. 7. Middle to late Eocene paleogeography of the Carlin trend (modern mountain ranges are shown in gray). a. ~46.1 Ma. The Elko basin, developed on the west flank of the Ruby Mountain-East Humboldt Range (RMEH). Synextensional strata of the Elko Formation filled the eastern part of the basin to a maximum thickness of ~800 to 1,300 m. The Elko Formation comprises fluvial-alluvial conglomerate and sandstone (+ minor volcanic tuff) overlain by a thicker package of fine-grained lacustrine sediments. Sedimentation in the western part of the basin began at ~42 Ma, when fluvial-alluvial basin fill and volcaniclastic sediments accumulated in a west-northwest– to east-west–trending, possibly fault-controlled, paleovalley. b. ~40.4 to 39.7 Ma. A sequence of ash flow tuffs (up to ~900 m thick) was deposited in the western Elko basin filling the paleovalley between the Carlin trend and Jerritt Canyon. The tuffs of Nelson Creek (40.4–39.8 Ma, unknown source area) and Big Cottonwood Canyon (39.7 Ma, sourced from the Tuscarora volcanic field) are regionally extensive units and their absence in the eastern Elko basin and the Emigrant Pass areas suggest that topographic barriers had isolated the western part of the basin by the time they were erupted. c. Post ~39.7Ma. Sedimentation and explosive pyroclastic volcanism in the western Elko basin ceased. Pervasive domino-style extensional faulting and andesite-dacite volcanism were initiated. Lacustrine sedimentation in the eastern Elko basin continued until a locally erupted felsic tuff filled the lake at ~38.9 Ma. Pervasive rotational extensional faulting extended into this part of the basin shortly afterward, prior to the extrusion of andesitic lavas at ~38 Ma. In the Pinion Range, ~37.5 Ma tuff units (Indian Well Formation) were deposited unconformably over faulted Elko Formation. Note that the Crescent Valley-Independence lineament of Peters (1998) does not appear to have had any major surface manifestation during the middle to late Eocene. Intrusive centers and andesite flow-dome complexes, (1) Tuscarora volcanic field, 39.9 to 39.3 Ma, (2) Lone Mountain, ~39 Ma, (3) Swales Mountain, ~39 Ma, (4) North Carlin-Emigrant Pass igneous complex, 39.3 to 36.0 Ma, (5) Elko Mountain hypabyssal rhyolite, 38.6 Ma, (6) Bullion stock, 36.7 Ma, (a) Pleasant Valley volcanic complex, ~39.8 Ma, (b) Sixmile Canyon lavas, 39.3 Ma, (c) Stampede Ranch and Pie Creek andesite, ~39.7 Ma, (d) Maggie Peak andesite-dacite complex, ~39.7 Ma, (e) Andesite of Fish Creek Mountain, (f) Dip Creek andesite, 39.4 Ma, (g) Susie Creek volcanic rocks and intrusions, (h) Emigrant Pass volcanic field, 37.8 to 37.2 Ma, (i) Flow-dome complex east of the Bullion stock, 37.5 Ma, and (j) Robinson Mountain volcanic rocks. The boundaries of the intrusive centers are approximate and are largely based on the regional aeromagnetic data of Hildenbrand and Kucks (1988). Geochronology after Henry and Boden (1998), Henry and Ressel (2000), Henry et al. (2001), Haynes (2003), K.A. Hickey, unpub. data, and R.M. Tosdal, unpub. data.


(a)
~46.1 - 40.4Ma
Jerritt Canyon
TH
~42Ma
Carlin trend
TH?
Carlin
Elko
~46Ma
Emigrant Pass
Pinion Range
Cortez Range
RMEH
0
40km
(b)
~40.4 - 39.7Ma
1
a
western Elko basin
eastern Elko basin
Range
Elko Mt
Adobe
RMEH
TH
0
40km
(c)
~39.7 - 36.0Ma
b
f
c
2
d
e
3
g
4
h
Elko
5
~38.9Ma
6
i
~37.5Ma
j
RMEH
TH
0
40km
Eocene intrusive centers
Eocene andesite-dacite flow-dome complexes
TH Topographic high
Direction of fluvial paleoflow
tuff of Big Cottonwood Canyon - flow direction
Syn-deposition faults
Crescent Valley-Independence lineament
Pervasive rotation extensional faulting (extensive post ~37Ma)
Pervasive rotation extensional faulting (mainly ~39.7-38Ma)
Indian Well Formation - air fall tuff, ash flow tuff and volcaniclastic sediments above Elko Formation in Pinion Range (~37.5Ma)
White air fall tuff, ash flow tuff and volcaniclastic sediments resting conformably on Elko Formation sediments near Elko Mountain (post-38.9Ma)
tuff of Big Cottonwood Canyon (39.7Ma)
tuff of Nelson Creek (40.4-39.8Ma)
Elko Formation - lacustrine sediments (shale, limestone)
Elko Formation - fluvial-alluvial sediments grading up to lacustrine sediments
Elko Formation - fluvial-alluvial sediments (sandstone, conglomerate)

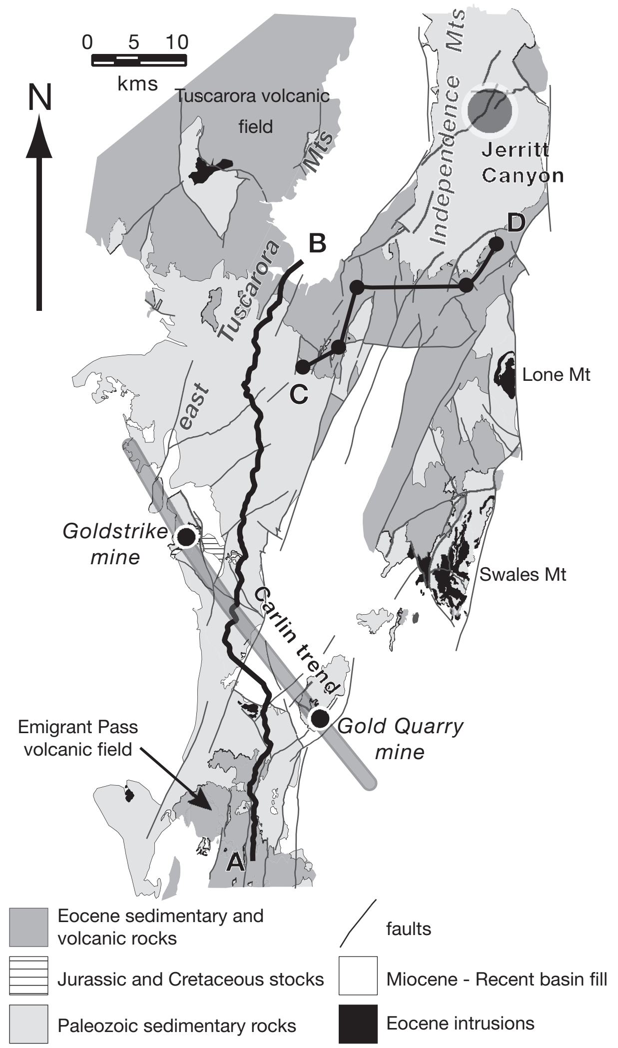


FIG. 8. Simplified geologic map of the Carlin-Jerritt Canyon region in northeastern Nevada, showing the locations of sections AB and CD in Figure 9.

directly overlain by andesite lavas near Swales Mountain (Fig. 8). Eocene ash-flow tuff sequences also thin northward from the paleovalley toward Jerritt Canyon, suggesting there may have been an Eocene topographic high in the Independence Mountains (Figs. 8 and 9a).

The maximum amount of relief generated by Eocene extension can be estimated from the thickness of synextensional sediments that accumulated in the extensional basins. In general, the absolute extent of footwall uplift that accompanies half-graben–style extension is similar to or less than the amount of absolute hanging-wall subsidence. (e.g., Schlische, 1991). Assuming that sediment supply did not significantly lag behind the growing capacity of the basin, the maximum amount of tectonically driven local relief should have been less than the maximum thickness of synextensional Eocene strata, which in the Elko basin is <800 to 1,300 m (e.g., Smith and Ketner, 1976; Satarugsa and Johnson, 2000; Haynes, 2003). Given the synextensional character of sedimentation and allowing for some footwall erosion, the maximum instantaneous amount of Eocene relief development may have been significantly less than this. For comparison, the late Miocene to Holocene basin in Crescent Valley has ~2 km of infill but has a footwall that rises only ~1,000 to 1,300 m above the valley floor (Friedrich et al., 2004).

The general pattern and middle to late Eocene timing of sedimentation and volcanism in the Elko basin is repeated in several smaller, satellite depocenters in the region. One such basin is preserved in the Alligator Ridge area at the southern end of the Ruby Mountains, associated with several shallow-formed Carlin-type Au deposits that define the southern end of the Carlin trend (Nutt and Hofstra, 2003). Other examples include the Copper basin (e.g., Rahl et al., 2002) and the Bull Run basin (Clark et al., 1985) to the northeast and north-northwest of Jerritt Canyon, respectively. Neither of these basins is associated with known Carlin-type mineralization. In contrast, Eocene sedimentary basins are not known to be associated with the Carlin-type Au deposits in the Cortez-Pipeline and Getchell-Twin Creeks, although middle to late Eocene and early Oligocene felsic tuff units are preserved in paleodepressions proximal to both regions (e.g., Wallace, 1993; Muntean et al., 2001).

The nature of Eocene extension in northern Nevada

Extension in Eocene strata: The nature of Eocene extension in northern Nevada is most apparent within and bounding the Eocene basins. The early phase of extension associated with basin development was largely accommodated by widely spaced growth faults. The detailed geometry of these structures is generally poorly constrained, although the main basin-forming faults bounding the eastern and western margins of the Elko and Copper basins, respectively, are large, shallow-dipping, north- to northeast-striking structures (e.g., Satarugsa and Johnson, 2000; Rahl et al., 2001). Other basin-bounding faults in the Carlin-Jerritt Canyon region appear to have had a west-northwest to east-west orientation (Fig. 7a).

The second phase of Eocene deformation that dismembered the Elko basin after ~39.7 Ma was characterized by pervasive domino-style faulting with northeast- to north-trending, gentle- to moderate-dipping fault blocks (e.g., Henry et al., 2001; Moore, 2001; Hickey et al., 2003b). Slip was relatively homogeneously distributed and total extension is estimated to have been ~15 to 40 percent. Similar styles of rotational faulting, but with varying amounts of extension, also developed in the Copper basin (Rahl et al., 2001), the late Eocene-early Oligocene Caetano tuff south of the Cortez area (Muntean et al., 2001), the Robinson mining district west of Ely (Gans et al., 2001), and the White Sage basin on the Nevada-Utah border (Potter et al., 1995), suggesting a regionally developed style of extension (e.g., Howard, 2003).

Extension in pre-Eocene bedrock: Although the northeast-trending rotational fault set developed in the Elko basin may be locally developed in pre-Eocene bedrock (Moore, 2001), there is little evidence to support the initiation of extensive new Eocene fault or fracture meshes. Rather, Eocene extension was largely accommodated by heterogeneous shear and tensional reactivation of older, variably oriented pre-Eocene structures. Eocene extension in northern Nevada was oriented broadly northwesterly to westerly (280°–330°), as indicated by the strike of Eocene-age domino-style fault blocks, by the

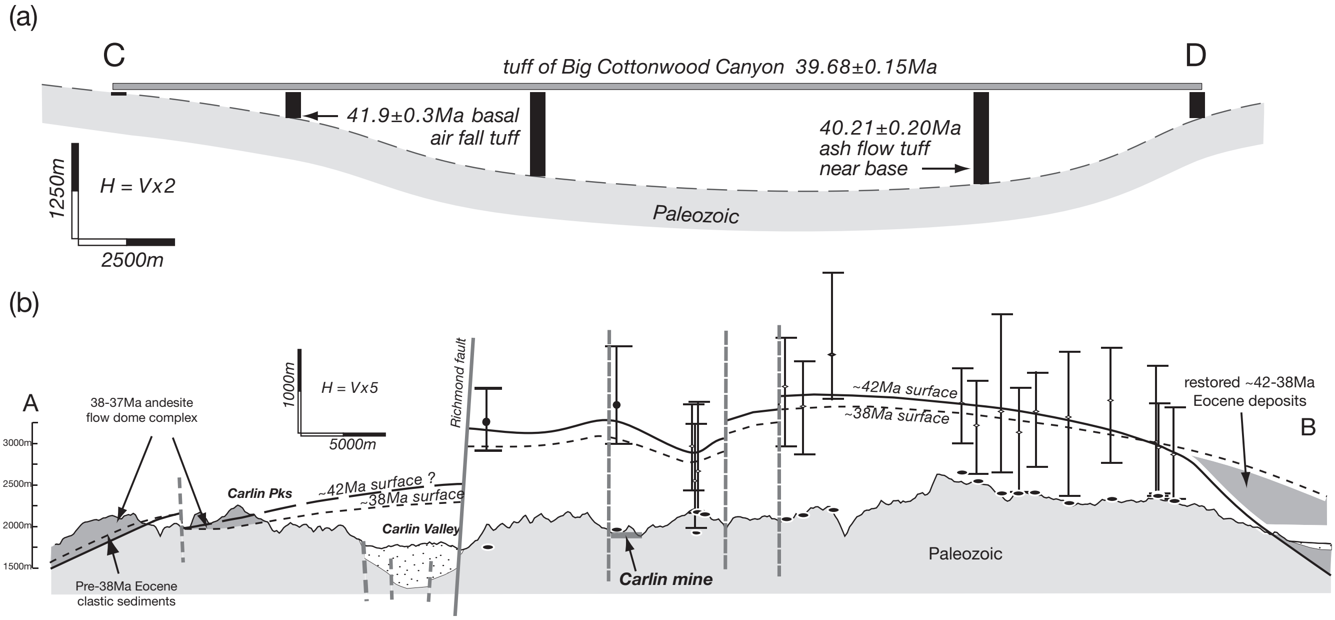


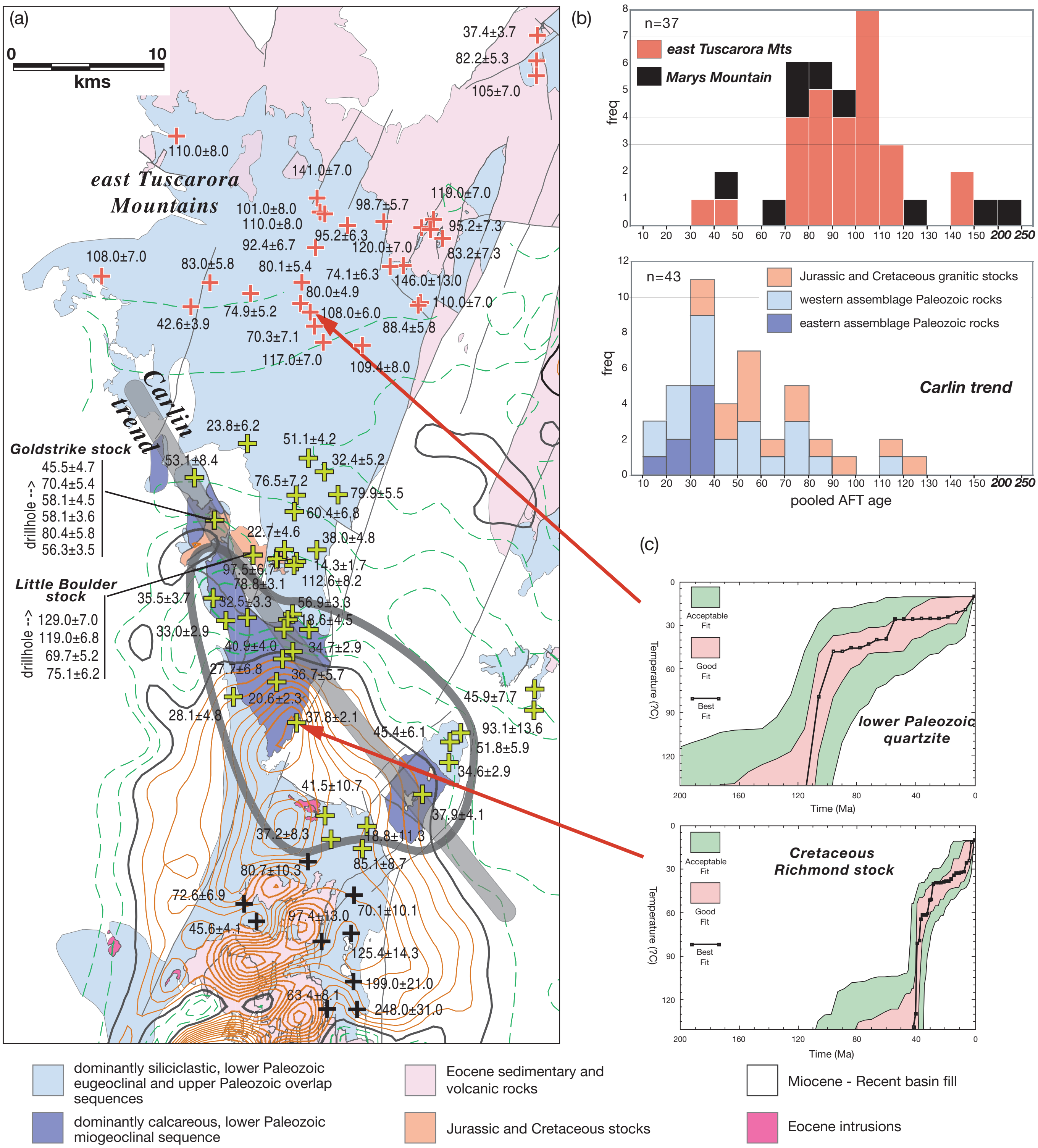
FIG. 9. a. Restored section across a west-northwest–trending Eocene paleovalley between the northern Carlin trend and Jerritt Canyon (2× vertical exaggeration). The tuff of Big Cottonwood Canyon is a regional marker unit and was used as a datum to restore post-39.7 Ma extensional faulting. The depth to the base of the Eocene sequence at five points along the section is shown by the vertical black bars. Isotopic ages from tuff units at, or near, the base of the sequence are also shown (K.A. Hickey, unpub. data). b. Restored Eocene landscape projected along the crest of the east Tuscarora Mountains (section AB, ×5 vertical exaggeration). The restored elevations of the 42 and 38 Ma erosion surfaces were derived from the thermal modeling of apatite fission tracks, assuming a surface temperature of 10°C and a near-surface geothermal gradient of 30°C/km (modeling was undertaken using AFTSolve; Ketchum et al., 2000). The vertical bars represent the range of calculated maximum elevations at 42 Ma that contain 95% of the modeled data with a >5% probability of matching the observed data. The weighted mean elevation of the 100 best (highest probability) models is shown by the symbol near the center of each bar. The position of the ~38 Ma surface is constrained by the restored thickness of Eocene units in the paleovalley at the north end of the section, and a pre-38 Ma erosion surface beneath 38 to 37 Ma andesite flows at the south end of the section. Landscape restoration suggests that the present erosion surface through the Carlin mine lies, at the most, ~1 to 2 km below the Eocene erosional surface at 42 to 38 Ma. These estimates increase to <~2 to 3 km for near-surface geothermal gradients of 20° to 25°C/km. Black ellipses mark sample locations projected into the topographic section.

general orientation of Eocene dikes (Presnal, 1997; Ressel et al., 2000; Henry et al., 2001; Theodore, 2002), and by the direction of midcrustal flow that is demonstrably of Eocene age in the Ruby Mountains-East Humboldt Range metamorphic core complexes (Howard, 2003, and references therein). The underlying Proterozoic rifted plate margin and northwest-oriented Paleozoic fault fabrics thus lay subparallel to the extension direction, and were favorably oriented to be reactivated as low-magnitude strike-slip (WNW-striking faults; Tosdal and Nutt, 1999; Nutt and Hofstra, 2003) or oblique-slip faults (NNW-striking faults; Heitt et al., 2003). Northeast-oriented pre-Jurassic fault fabrics that lie at high angles to the tensional stress (e.g., the Crescent Valley-Independence lineament of Peters, 1999) were reactivated in extension (Fig. 7) and may have locally controlled the geometry of Eocene basins. Shallow-dipping pre-Jurassic thrust faults were favorably oriented to reactivate as low-angle extensional faults and may have locally accommodated much of the pervasive rotational extension evident in the overlying Eocene basins (Moore, 2001; Muntean et al., 2001).

Tosdal and Nutt (1998, 1999) proposed that the Carlin trend acted as an accommodation zone during Eocene extension. Slip along the zone was a manifestation of minor lateral displacement caused by the crustal blocks on either side undergoing differing magnitudes of Eocene extension. Evidence for such lateral displacement includes (1) en echelon alignment of ore-controlling faults and deposits (Dee-Rossi, Capstone-Bootstrap-Tara, and Ren-Meikle-Beast Group; Fig. 4c); (2) change in strikes of major ore-controlling faults from west-northwest to more northerly along a trend from southeast to northwest (Teal and Jackson, 1997); (3) consistent northeast strikes of secondary faults controlling Au deposition (summarized by Teal and Jackson, 1997); (4) sinistral deflection of north-trending Mesozoic folds and thrusts in the Alligator Ridge area (Nutt and Good, 1998; Tosdal and Nutt, 1999; Nutt and Hofstra, 2003); and (5) the broad-scale deflection of older structural fabrics. Opposing senses of upper crustal slip in crustal blocks on either side of the Carlin trend reflect differential extension within each block (Tosdal and Nutt, 1998, 1999). The Battle Mountain-Eureka trend may also correspond to an accommodation zone across which the Eocene to Miocene extensional history differs in bulk extension, style, or orientation (Muntean et al., 2001).

Thermochronological evidence for Eocene hydrothermal flow

Apatite fission tracks in samples collected from the northern Carlin trend were partially to fully annealed by heterogeneous thermal pulses from ~40 to 15 Ma. A zone of pervasive annealing preserves evidence for an episode of rapid cooling from >100°C at ~40 Ma (Fig. 10; Chakurian et al., 2003), with localized annealing of fission tracks as late as ~20 Ma. The zone of annealing is broadly parallel to the Carlin trend and becomes more heterogeneously distributed and less pervasive farther to the northwest, toward the Post-Betze-Screamer and Meikle deposits, the largest deposits in the district (Hickey et al., 2003b). The spatial and temporal overlap of fission-track annealing and Carlin-type Au mineralization (Table A3) has been interpreted to reflect a genetic relationship (Chakurian et al., 2003; Hickey et al., 2003a). This association is, however, not perfect (Fig. 10), as significant deposits lie outside the area of pervasive annealing and, in the



Gold Quarry area, the thermal anomaly extends laterally beyond the Carlin trend (Fig. 10). The trend may reflect the more focused flow of Au-bearing fluid relative to those fluids responsible for the zone of apatite annealing.

Paleogeographic and tectonic controls on Eocene hydrothermal fluid flow and Au mineralization

The paleogeographic and tectonic development of northern Nevada outlined above has several implications for models of Eocene hydrothermal fluid flow and Carlin-type Au mineralization.

1. The limited local paleorelief associated with the Elko basin (<1,000 m) and the location of the northern Carlin trend on a relative paleogeographic high suggest that local topographic hydraulic gradient is unlikely to have been the main driving force for the upflow of deeply circulating meteoric water into the northern Carlin trend during mineralization. Therefore, it seems more probable that thermal energy provided the main impetus for hydrothermal circulation, with local relief influencing near-surface flow paths.

2. The estimated 42 to 36 Ma age of Carlin-type mineralization in the Carlin-Jerritt Canyon region implies that the Au deposits formed after the Elko basin began to form. Mineralization did, however, overlap with the stage of low-magnitude pervasive rotational extension that dismembered the basin after ~39.7 Ma (Fig. 7). This extension coincided with the cessation of pyroclastic volcanism and the extrusion of small, commonly fault-controlled, andesite-dacite flow-dome complexes. Extension fractured the crust, enabling more intermediate-composition magmas to vent prior to their evolution to a more felsic composition. Formation of closely spaced normal faults would also have enhanced the permeability of the region, enabling deeper and more pervasive circulation of meteoric water through the crust. Mineralization appears to have been temporally associated with this phase of extension (Table A3).

3. Fluid flow and mineral deposition mainly appears to have been a passive phenomena, as there is little evidence for widespread overpressured hydrothermal fluids, complicated multistage vein dilatancy, or significant synmineralization slip, except locally where tectonic and hydrothermal breccias have been recognized (Kuehn and Rose, 1992; Cline and Hofstra, 2000; Hofstra and Cline, 2000; Williams et al., 2000; Heitt et al., 2003). Instead, mineralization was associated with the heterogeneous shear and tensional reactivation of older, variably oriented pre-Eocene structures and with incompatible strain around bodies of greater rheological competence, such as plutons or their hornfels aureoles. The magnitude of dilatancy was controlled by the orientation of such features with respect to the bulk northeast-oriented displacement (Fig. 6).

Post-Mineral History

Subsequent to Au deposition, heterogeneous extension of the Great Basin, accompanied by magmatism, continued into the Oligocene and early Miocene (34–17 Ma). Ductile flow of the middle crust, evident in the localized development of metamorphic core complexes east of the Carlin-type deposits, was accommodated in the upper crust by low- and high-angle normal faults (Wernike, 1981; Gans et al. 1989). At about 17 Ma, there was a fundamental switch in extensional style, with the regional development of widely spaced, steeply dipping normal faults that resulted in a series of variably tilted fault blocks separated by alluvial basins (Stewart, 1980; Zoback et al., 1981). Basin and Range-style faulting was accompanied by bimodal volcanism and hydrothermal activity that produced epithermal Au, Ag, and Hg deposits (Christiansen and Lipman, 1972; John, 2001). Some of these deposits occur where the northern Nevada rift intersects the Carlin (Ivanhoe Au-Hg district; Wallace, 2003) and Battle Mountain-Eureka (Buckhorn Au-Ag district; John and Wallace, 2000) trends. Intersections are represented by small Hg deposits near Meikle (Purple Vein; Emsbo et al., 2003) and Gold Quarry in the overlying Miocene Carlin Formation. Very young (≤2 Ma) breccias cemented with calcite and barite, like those at Meikle (Hofstra et al., 2000; Emsbo and Hofstra, 2003), are the product of still younger hydrothermal circulation that used the same fracture mesh that localized Carlin-type Au deposits.

The Eocene landscape reconstruction shown in Figures 7 and 9b suggests that there has been ~1,000 to 2,000 m of denudation in the Carlin-Jerritt Canyon region since ~42 to 38

FIG. 10. a. Pooled apatite fission-track ages for pre-Cenozoic samples from the region around the Carlin trend (uncertainties are 1σ). A zone of pervasive Eocene-Miocene thermal annealing of apatite spatially associated with the Carlin trend is outlined in gray. Samples from the zone of annealing have mainly Eocene and younger apatite fission-track ages. Aeromagnetic contours are the total residual field anomaly adapted from Hildebrand and Kucks (1988) and define a subsurface, ~38 Ma, intrusive complex (Henry and Ressel, 2000). Contour intervals are 50 nanoteslas; thick solid contour nT = 0; thin orange contour nT >0; thin dashed green contours nT <0. b. Frequency distribution of apatite fission-track ages from the east Tuscarora Mountains, the Marys Mountain area, and the Carlin trend (red, black, and green crosses in (a), respectively). The older, Cretaceous, age cluster is a product of denudation during the Sevier orogeny and represents the regional background cooling history. The younger, Cenozoic, age cluster reflects partial to complete annealing of fission tracks as a consequence of spatially restricted convective thermal pulses between ~40 and 15 Ma. Annealing was concentrated around the window of lower plate miogeoclinal rocks that hosts most of the Au deposits in the northern Carlin trend. Annealing in areas away from the Carlin trend (within the eugeoclinal sequence) tends to have been less pervasive and less complete. Samples of the Jurassic Goldstrike and Little Boulder stocks (see a) adjacent to the Carlin trend were not by Eocene annealing and preserve mainly Cretaceous to Paleocene apatite fission-track ages. This suggests that there was little convective circulation of hydrothermal fluids through these intrusions during the Eocene. c. Temperature-time (T-t) cooling paths derived from thermal modeling of apatite fission-track data from a sample of Paleozoic sandstone in the east Tuscarora Mountains and the Cretaceous Richmond stock (modeled using AFTSolve, Ketcham et al., 2000). The Paleozoic sample, like most samples distal from the Carlin trend, underwent its last major phase of cooling at ~100 to 110 Ma, whereas the sample from the Richmond stock cooled from >100°C at ~40 Ma.

Ma. Most of this erosion is likely to have occurred prior to the middle Miocene because a pre-Basin and Range low-relief landscape characterized by broad, flat summits caps the mountain ranges in the immediate vicinity of the deposits. Summit-flat erosion is controlled by hillslope processes involving bedrock weathering and surface runoff. Estimates of long-term denudation rates on low-relief summit flats in temperate to semiarid, or arid, environments (like that in NE Nevada from the Miocene onward) are <50 to 100 m/m.y. and typically between 2 and 30 m/m.y. (e.g., Dohrenwend, 1990; Small et al., 1997; Ellis et al., 1999). These latter values suggest that the summit flats in the Carlin-Jerritt Canyon region are 30 to 450 m below the erosion surface that was present at the initiation of Basin and Range extension and that the Au deposits in the northern Carlin trend were exhumed to within several hundred meters of the surface prior to ~17 Ma.

Supergene modification and oxidation of the orebodies was contemporaneous with exhumation. This postmineral oxidation was critical to the initial development and mining of many deposits because it liberated Au from pyrite in otherwise low-grade hypogene material, making upper levels of deposits amenable to open-pit mining and Au recovery by cyanide heap-leach methods. Evidence that indicates an important but poorly understood role for supergene processes includes karst features, argillic alteration (halloysite, disordered 1A kaolinite, alunite) with isotopic signatures of Neogene meteoric water, the presence of subhorizontal Au zones in some deposits that parallel redox fronts (Mike, Gold Quarry), enrichment blankets composed of supergene sphalerite with extremely negative $\delta^{34}S$ values that lie below but parallel to oxidized Au zones (Mike), and local remobilization and coarsening (e.g., to 1–100 μm) of Au where auriferous pyrite was oxidized to goethite (Hofstra et al., 1999; Phillips et al., 1999; Hofstra and Cline, 2000; Bawden et al., 2003; Heitt et al., 2003). Supergene alunite deposition began as early as 30 Ma (Gold Quarry) and continued to about 8 to 10 Ma, with some limited alunite deposition as recent as 3 Ma (Hofstra et al., 1999; Table A1).

Description of Orebodies

Carlin-type ore is characterized by disseminated, Au-bearing, and trace element-rich pyrite that occurs in replacement bodies in carbonate host rocks. The various forms of individual orebodies (e.g., tabular, strata-bound, carrot, T-shaped, irregular) reflect local zones of porosity and permeability that result from favorable lithologic features, high- and low-angle faults, and especially intersections of these features. The largely low-angle and stratiform bodies may have root zones that project toward high-angle feeder faults. Permeable features contributing to the geometry of orebodies include high-angle faults, thrust faults, low-angle normal faults, hinge zones of anticlines, lithologic contacts, reactive carbonate units, debris-flow deposits, facies changes, brecciated zones between rocks of differing lithology (especially along zones of flexural slip), and contacts of sedimentary rocks with metamorphic aureoles related to Mesozoic intrusions. Aquitards include structures and less-permeable rocks such as shales and intrusive rocks.

Deposits are commonly elongated in north or northwest directions parallel to high-angle structures, although intersections with northeast high-angle faults can be important and influenced some deposit forms. Deposits are up to 3 km in length with ore occurring over a kilometer vertically, although such orebodies are generally stacked and are not vertically continuous (Fig. 3). Orebodies are typically capped by less-permeable horizons, and the best ore grades are commonly concentrated beneath domes or anticlines where high-angle structures acted as feeders in a style similar to that of petroleum reservoirs.

While igneous rocks are present in some districts, features that would genetically link ore fluids with coeval, epizonal magmas are lacking. Preore Jurassic to Tertiary dikes and sills intruded along structural pathways that subsequently acted as conduits for ore fluids (Teal and Jackson, 2002). Within other deposits dikes and sills acted as aquitards, largely owing to clay alteration, enhancing grade in sedimentary rocks adjacent to or beneath dikes and sills (e.g., Meikle, Emsbo et al., 2003; Getchell, Tretbar, 2004). Such features result in ore zones correlating with older igneous bodies. Dikes, sills, and/or intrusions that are approximately coeval with mineralization have been identified in some but not all districts. Where such intrusive rocks are present (Carlin trend, Heitt et al., 2003; Jerritt Canyon, Hofstra et al, 1999), metals, ore minerals, and alteration assemblages lack zoning relationships with respect to the igneous rocks. Mineral assemblages, such as Eocene skarns and polymetallic replacement bodies, that would be expected between intrusions and more distal Carlin-type deposits, have not been identified in the northern Carlin trend, Getchell-Twin Creeks, and Pipeline-Cortez, despite deep drilling. The lack of zoning and high-temperature assemblages suggests that strong thermal or chemical gradients between Carlin-type deposits and proximal intrusions were absent.

Deposit mineralogy and geochemistry reflect the association of Au with As, Sb, Tl, and Hg in preference to base metals and Ag. Most Au was deposited with main ore-stage minerals, including Au-bearing arsenian pyrite and marcasite, quartz, kaolinite, dickite, and illite (Fig. 11). These minerals are fine grained and typically volumetrically minor to insignificant (Arehart et al., 1993b; Arehart, 1996; Hofstra and Cline, 2000) in comparison to relict host-rock minerals that include quartz, micas, and clay minerals, dolomite, calcite, pyrite, and various forms of C. However, in the highest grade ore zones or deposits (e.g., Meikle, Deep Star), ore-stage hydrothermal minerals can comprise much of the rock (Emsbo et al., 2003; Heitt et al., 2003). Late ore-stage minerals, which are generally macroscopic, precipitated in voids or crosscutting fractures as hydrothermal systems collapsed and cooled (Hofstra et al., 1991) and are generally dominated by calcite, pyrite and/or marcasite, quartz, orpiment, realgar, and stibnite, with realgar and calcite typically having precipitated last (Fig. 11). Other minerals, including galkhaite, fluorite, cinnabar, sphalerite, and tellurides, as well as microscopic-size trace minerals (e.g., Ferdock et al., 1997), are present in some but not all deposits (Hofstra and Cline, 2000). Barite, generally described as late-ore stage and abundant and spatially related to ore at Rain, is postore where timing has been established, as are some calcite veins (Getchell, Cline, 2001; Meikle, Emsbo and Hofstra, 2003; Betze-Post, Lubben, 2004). Botryoidal pyrite and/or marcasite and sphalerite with low $\delta^{34}S$ values are postore and supergene in origin (Bawden et al., 2003).

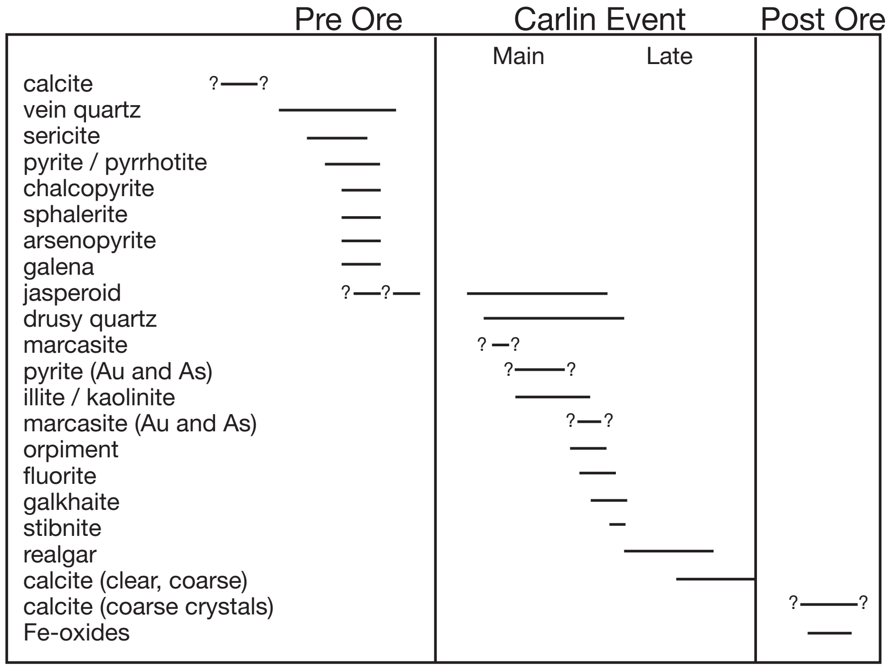


FIG. 11. Paragenetic sequence of pre-ore, Carlin and postore-stage minerals determined for the Getchell deposit. The sequence illustrates the minerals and their relative timing typical of many Carlin-type deposits (modified from Cline and Hofstra, 2000).

Wall-Rock Alteration

Host rocks are typically decarbonatized, argillized, and variably silicified in addition to being sulfidized and enriched with Au (App. Fig. A5; e.g., Hausen and Kerr, 1968; Bakken and Einaudi, 1986; Kuehn and Rose, 1992; Stenger et al., 1998; Hofstra and Cline, 2000). As the reactions that formed Au-bearing pyrite are generally not the same reactions that formed alteration minerals (Hofstra et al., 1991), ore and alteration zoning within the deposits is irregular. This lack of a temporal, and thus spatial relationship, along with the fine-grained nature and sparse abundance of ore and alteration minerals, limits the use of alteration to identify ore.

Decarbonatization

Carbonate rocks were dissolved in nearly all deposits and most ore zones and locally replaced with quartz to form jasperoid. Fluid acidity, the amount of cooling, and the degree of fluid-rock interaction controlled the extent of ore zone decarbonatization (Hofstra and Cline, 2000), which varies from minimal (e.g., Gold Bug, Screamer) to intense (e.g., West Leeville, Carlin, Meikle). Intense decarbonatization produced collapse breccias in some deposits (Bakken, 1990), significantly enhancing porosity, permeability, and fluid-rock reaction, leading to formation of high-grade ore (Emsbo et al., 2003).

Zones of complete carbonate dissolution near fluid conduits commonly are zoned outward, first to dolomite- and then to calcite-stable zones, and farther out to zones with abundant calcite veins. The presence of various carbonate minerals—calcite, siderite, ankerite, or dolomite—reflects the preore composition of wall rocks, preferential dissolution of calcite, sulfidation of ferroan carbonate minerals, or carbonatization of silicate minerals beyond the sulfidation halo (Hofstra and Cline, 2000). Calcite veins in the outer zone generally have $\delta^{13}C$ values that are similar to the carbonate host rocks, suggesting they formed from CO_2 released by decarbonatization or collapse of cool CO_2-bearing meteoric water into the hydrothermal system; however, at least some calcite veins are postore (Hofstra and Cline, 2000; Emsbo et al., 2003; Emsbo and Hofstra, 2003; Lubben, 2004).

Argillization

Wall rocks are argillized where moderately acidic ore fluids reacted with older alumino-silicate minerals and formed assemblages of kaolinite ± dickite ± illite (Folger et al., 1998; Hofstra and Cline, 2000). Basalts, lamprophyres, and other igneous rocks are commonly intensely argillized, whereas argillization is minimal in relatively pure carbonate rocks.

Although it is difficult to confidently distinguish clay minerals formed at the time of Au deposition from clays related to earlier hydrothermal and later supergene events, 1Md kaolinite and 2M1 dickite were determined to be coincident with mineralization in some deposits (Alligator Ridge, Ilchik, 1990; Vista orebody at Twin Creeks, Osterberg and Guilbert, 1991; Carlin, Kuehn and Rose, 1992; Getchell, unpub. Placer Dome data, 1999; Deep Star, Clode et al., 1997; Altamirano-Morales, 1999; Heitt et al., 2003). Kaolinite and dickite proximal to ore-fluid conduits are zoned outward to intergrown fine muscovite-illite (2M1 or 1M) or smectite, and farther out to K-feldspar–stable zones (Carlin, Kuehn and Rose, 1992; Betze-Post, Arehart et al., 1993b; Deep Star, Heitt et al., 2003), in response to reduced fluid-rock reaction and consumption of H ions (Fig. A5).

A few detailed studies have identified increasingly crystalline illite proximal to ore, suggesting a genetic association (Gold Bar, Masinter, 1991; Chimney Creek, Osterberg and Guilbert, 1991; Jerritt Canyon, Folger et al., 1998; Pipeline, Placer Dome, unpub. data, 1999; Getchell, Cail and Cline, 2001). Fine muscovite and some illite, however, have provided a wide range of ages, many of which are older than and unrelated to Au but which previously contributed to controversy over the age of the deposits (see Hofstra et al., 1999). The range of ages indicates that preore muscovite and illite remained stable through all but the most intense alteration, also retaining their preore Ar signatures.

Silicification

Silicification accompanied some but not all Au deposition and is manifested by the presence of jasperoid and, to a lesser extent, by fine quartz druses lining vugs; ore-stage quartz veins are relatively uncommon in these deposits. Jasperoid is spatially associated with ore at the district scale, yet jasperoids range from being barren to containing high-grade ore (Bakken and Einaudi, 1986; Ye et al., 2002). Subeconomic grades reflect, in part, the decoupling of processes that precipitated jasperoid from those that deposited Au. For example, some upwelling ore fluids cooled and formed jasperoids in fluid conduits below ore zones and prior to Au deposition, whereas other fluids precipitated jasperoid above or lateral to ore zones as fluids exited sites of Au deposition (Hofstra and Cline, 2000). Where relatively pure limestones were silicified, massive jasperoids were produced. In contrast, where limey siltstones, sandstones, or shales were silicified, fine calcite laminae were replaced by "micro" jasperoids.

Quartz druses line vugs formed by decarbonatization or in collapse breccias and, rather than veins, are a characteristic textural feature of Carlin-type deposits (Cline and Hofstra, 2000; Cline, 2001; Emsbo et al., 2003; Lubben, 2004). Gold-bearing pyrite occurs both within jasperoid and incorporated in the base of drusy quartz crystals at the jasperoid-drusy quartz interface (Cline, 2001; Lubben, 2004). Postore drusy quartz, distinguished by cathodoluminescence studies at some deposits, has overgrown ore-stage quartz druses (e.g., Meikle, Emsbo et al., 2003; Betze-Post, Lubben, 2004).

The generally low abundance of ore-stage quartz and lack of one-to-one correlation between quartz and Au in Carlin-type deposits may be related to the relatively low temperature of ore formation (see Hofstra and Cline, 2000, and references therein) and the fact that quartz precipitation is inhibited by kinetics below ~180°C (Henley and Ellis, 1983; Fournier, 1985; Rimstidt, 1997). As temperatures declined, quartz precipitation likely diminished although Au precipitation may have continued.

Jasperoid textures are related to minerals that initially replaced carbonate minerals and, as such, indicate approximate temperatures and depths of formation (Lovering, 1972). Jasperoid displaying reticulate or xenomorphic textures (e.g., Getchell; Cline, 2001) results from carbonate replacement by quartz (Lovering, 1972), indicating formation at somewhat elevated temperatures (>180°C) and greater depths, and such "high-temperature" jasperoid is more likely to have been related to Au deposition in Carlin-type deposits. By contrast, jasperoid that displays jigsaw or chalcedony textures indicates that lower temperature forms of silica, such as amorphous or opaline silica, initially replaced calcite, though they may have been subsequently recrystallized to quartz (Lovering, 1972). Observed textures suggest that amorphous or opaline silica precipitated at some deposits (Alligator Ridge, Beast, South Carlin, Gold Bar, Eureka; Nutt and Hofstra, 2003; Table A2), probably at shallow depths and from depleted ore fluids or fluids unrelated to the Carlin system.

Au-bearing arsenian pyrite

Carlin-type Au deposits are perhaps best known for the consistent occurrence of submicron, so-called invisible Au in trace element-rich pyrite and marcasite, even in samples in which Au exceeds several ounces per ton. Gold-bearing pyrite and marcasite occur as discrete grains, generally less than a few micrometers in diameter, or as narrow rims on earlier formed pyrites (Wells and Mullens, 1973; Arehart et al., 1993b; Simon et al., 1999; Hofstra and Cline, 2000). Visible Au has been observed locally (Wells and Mullens, 1973; Bakken et al., 1989; Emsbo et al., 2003) but is interpreted to have been liberated and coarsened during weathering and oxidation of ore-stage pyrite (Wells and Mullens, 1973) or deposited by earlier Devonian sedex or later events (Emsbo et al., 1999, 2003), although some have suggested direct Au deposition from hydrothermal fluids (Rain, Shallow, 1999; Gold Quarry, Rodeo, T. Thompson, pers. commun., 2004).

Ore-stage pyrite contains a range of Au concentrations that reach ~9,000 ppm (Wells and Mullens, 1973; Henkelman, 2004; Palenik et al., 2004; Reich et al., 2005), and the overall grade of a particular ore zone is a function of the abundance of ore-stage pyrite and its Au concentration. Arsenic is the most abundant nonstoichiometric element in the pyrite; concentrations reach, or in a few cases exceed, ~15 wt percent (Cline, 2001, unpub. data; Reich et al., 2005), although 2 to 8 wt percent is more common and As/Au ratios are highly variable. Elevated As does not, however, distinguish Carlin-type pyrite, as As concentration is also high in older intrusion-related, late ore-stage, and postore-stage pyrite (Arehart et al., 1993b; Hofstra and Cline, 2000; Emsbo et al., 2003; Henkelman, 2004). Other trace elements in pyrite and marcasite include as much as ~0.25 to 1.0 wt percent Sb, Hg, Tl, Te, and Cu, and variable but generally lower Pb, Mo, Zn, Mn, Bi, Ni, W, Ag, and Co (Wells and Mullens, 1973; Hofstra and Cline, 2000; Cline, 2001; Emsbo et al., 2003; Henkelman, 2004).

Metal patterns in zoned pyrite rims document an evolving ore-fluid chemistry and suggest that early fluids contained the highest Au. Early fluids at Getchell and Meikle (Cline et al., 2003; Emsbo et al., 2003), represented by inner pyrite rims, had elevated concentrations of As, Hg, Tl, Te, Cu, ±Pb in addition to Au. With time, concentrations of most metals declined. Outermost pyrite rims, representing later fluids, contain lower concentrations of Au and the ore-metal suite with some metals below detection but elevated concentrations of other metals including Pb and W, which are interpreted to have been leached from wall rocks (Cline et al., 2003).

In an effort to understand the mechanisms of Au precipitation, numerous studies have speculated on whether Au occurs in pyrite as structurally bound ionic Au^{+1} or as submicron inclusions of native Au^0 (Bakken et al., 1989; Arehart et al., 1993b; Fleet and Mumin, 1997). More recently, Simon et al.

(1999) identified both Au^0 and Au^{+1} in arsenian pyrite from Twin Creeks and concluded that structurally bound Au^{+1} was adsorbed from ore fluids that were not saturated with respect to native Au. Recent high-resolution transmission electron microscopy (TEM) studies of samples from several Carlin trend deposits (Palenik et al., 2004; Reich et al., 2005) determined that most Au occurs in solid solution in pyrite, but nano-sized particles of Au^0 are present where Au exceeds a solid-solution molar solubility limit that is a function of As concentration (Au/As >0.02; Reich et al., 2005). Au^{+1} is present in solid solution below this solubility limit. These results are interpreted to indicate that Au was adsorbed onto pyrite grains from ore fluids undersaturated in native Au.

Organic C

Black carbonaceous rocks occur in every district and commonly host some of the ore (Hofstra and Cline, 2000). Indigenous C occurs as intergranular films and particles along sedimentary layers and is concentrated in stylolites. Carbon is also concentrated in pores, along fractures, and in quartz and calcite veins. Rocks with high concentrations of indigenous C (~>2 wt %) were deposited in restricted anoxic basins and are potential petroleum source rocks, whereas those with high S/C ratios are indicative of euxinic conditions during deposition (Leventhal and Giordano, 2000). Some rocks with high concentrations of C in the Carlin trend and Alligator Ridge district provide evidence of petroleum generation, migration, and accumulation in structural culminations (e.g., Kuehn, 1989; Hulen and Collister, 1999).

In rocks below the Roberts Mountains allochthon, both indigenous and concentrated organic C are generally overmature relative to petroleum generation and typically consist of cryptocrystalline graphite (Poole and Claypool, 1984; Hofstra, 1994; Leventhal and Giordano, 2000). Field relationships indicate that petroleum migration occurred prior to emplacement of Jurassic intrusions (e.g., northern Carlin trend, Emsbo et al., 2003), and thermal modeling suggests that petroleum generation and subsequent catagenesis was due to emplacement of the Roberts Mountains allochthon (Gize, 1999). Laser Raman spectra show that C crystallinity is the same in ore and waste and only increases adjacent to intrusions such as the Goldstrike stock in the northern Carlin trend (Hofstra, 1994; P. Emsbo, pers. commun., 1998).

In rocks east of the allochthon, the thermal maturity of indigenous and concentrated C is generally lower and more variable, with higher maturity near igneous intrusions, Carlin-type deposits, and active geothermal areas (Poole and Claypoole, 1984; Ilchik et al., 1986, Hitchborn et al., 1996; Hulen and Collister, 1999). At Yankee, paragenetic relationships indicate there were two episodes of petroleum generation, one during Carlin-type mineralization and one after in response to Basin and Range tectonism (Hulen and Collister, 1999; Nutt and Hofstra, 2003).

Hofstra and Cline (2000) found no consistent relationship between organic C content and ore grade. They also noted that many orebodies occur in rocks with low C contents and concluded that C played little or no role in Au precipitation. At upper crustal levels in these systems, organic matter may have been a direct source of H_2S and a critical ingredient for H_2S generation by thermochemical sulfate reduction (Hunt, 1996). Organic matter also maintained a reduced condition in the fluid, and thus H_2S as the predominant form of S in the fluid, enabling the hydrothermal solutions to scavenge and transport Au to the sites of ore formation (Hofstra and Cline, 2000). During exhumation and weathering of the deposits, C consumed O in supergene fluids such that enclaves of carbonaceous ore were locally preserved.

Lithogeochemistry

Lithogeochemical studies show fairly consistent patterns of elemental flux into and out of deposits in several districts during ore formation (Jerritt Canyon, Hofstra, 1994; Twin Creeks, Stenger et al., 1998; Getchell, Cail and Cline, 2001, Gold Bar, Yigit et al., 2003; Meikle, Emsbo et al., 2003; Screamer, Kesler et al., 2003a; Table A2). Gold/silver ratios, commonly >10 in sedimentary rocks and ~10 in igneous dikes (Table A2), demonstrate consistently high transport and precipitation of Au relative to Ag into the deposits and are a key signature that distinguish Carlin-type Au deposits from most other precious metal systems.

Elements added to host rocks in all deposits include S, As, Au, Sb, Hg, and Tl. Elements that are variably added to altered rocks, or identified as added in some deposits but not all, include Te, Cu, W, Mo, Se, Fe, Ag, Pb, Si, Ba, Cs, and Zn. As demonstrated by ore-stage pyrite compositions, some of these elements, including Cu and Te, appear to be components of the early, high Au ore fluid whereas others may have been leached from wall rocks. Base metals are typically not quantitatively added to rocks. Lead appears to have multiple origins, with deeply sourced Proterozoic rocks contributing Pb to early ore fluids, and wall rocks contributing Pb to late ore fluids (Tosdal et al., 2003). Although trace metals other than Au are significant from a genetic standpoint, concentrations are low and elements generally occur only as trace impurities in pyrite and marcasite.

Elements removed from rocks by ore fluids include carbonate C, Ca, Mg, Sr, Mn, Na, and Sc. Generally immobile elements include Al, Ti, Th, and Zr (Hofstra and Cline, 2000). These patterns reflect alteration, particularly decarbonatization and argillization. Potassium is variably immobile or locally added or removed in trace amounts (Cail and Cline, 2001).

The mobility of Fe is important as it reflects the degree to which sulfidation controlled pyrite and Au deposition. Iron was relatively immobile in sedimentary rocks and mafic dikes at Jerritt Canyon (Hofstra et al., 1999), in wall rocks at Twin Creeks and Gold Bar (Stenger et al., 1998, Yigit et al., 2003), in zebra dolomite and igneous rocks at Meikle and Post-Betze (Emsbo et al., 2003), and in most ore zones at Getchell (Cail and Cline, 2001). However, Fe was added to ore at Screamer (Kesler et al., 2003a), in some ore zones at Getchell (Cail and Cline, 2001), and in minor amounts at Alligator Ridge (Nutt and Hofstra, 2003). Iron transport and the addition of pyrite to the rock are locally evident by the presence of abundant ore-stage pyrite in high-grade zones in sedimentary rocks below basalt flows (Getchell, Twin Creeks) and other Fe-bearing aquicludes and in argillized tuffs (Tonkin Springs).

Ore Fluids and Ore Deposition

Fluid inclusions in ore-stage minerals in Carlin-type deposits are small, sparse, and difficult to relate to Au mineralization.

Nevertheless, studies that provided similar results for several districts (App. Fig. A6) show that ore-stage fluids were low- to moderate-temperature (180°–240°C), low-salinity (~2–3 wt % NaCl equiv) aqueous fluids that contained CO_2 (<4 mol %) and CH_4 (<0.4 mol %), and sufficient H_2S (10^{-1}–10^{-2} m) to transport Au and other bisulfide-complexed metals (Cline and Hofstra, 2000; Hofstra and Cline, 2000; Lubben, 2004). There is no fluid inclusion, mineralogic, or textural evidence for fluid boiling or immiscibility.

CO_2 dissolved in the ore fluid controlled fluid pH and, at depth, where fluids were hot (i.e., >330°C) and in equilibrium with wall rocks, pH would have been near neutral (Giggenbach, 1997). As ore fluids ascended and cooled, H_2CO_3 progressively dissociated, generating increasingly acidic fluids that dissolved carbonate and hydrolized feldspars to kaolinite, dickite, or illite. A simple equilibrium speciation and reaction path model (REACT, The Geochemists Workbench; Bethke, 2001) was used to simulate reaction of an ore fluid at 200°C containing 4 mole percent CO_2 and 0.6 wt percent NaCl with a calcite (1 kg) plus dolomite (0.2 kg) host rock containing pore fluid at a pH of 7 and 60°C (e.g., 2-km depth assuming geothermal gradient of 30°C/km; App. Fig. A7). Model results indicate that initial reaction between the fluid and wall rock causes pH to rapidly decrease to 4.3 to 4.5. Five hundred and twenty kg of ore fluid are required to remove 1 kg of limestone, 785 kg are required to dissolve 0.2 kg of dolomite, and significant CO_2 is generated (log fugacity ±250 bars). Following limestone and dolomite dissolution, pH decreases to 3 to 3.5 where kaolinite is stable. The lack of evidence for boiling or fluid immiscibility in Carlin-type systems suggests that confining pressure exceeded the PCO_2 generated from carbonate dissolution, preventing H_2O-CO_2 immiscibility, pH increase, and precipitation of calcite. These observations provide minimal initial constraints for deposit depth.

At temperatures >400°C, a near-neutral pH and chloride-poor, sulfide-bearing ore fluid would have transported Au as $Au(HS)_2^-$; Seward, 1973; Benning and Seward, 1996; Stefansson and Seward, 2004). As ore fluids ascended to the upper crust, cooled to ~240°C, and became acidic, the ligand responsible for Au transport may have switched (Hofstra and Cline, 2000) to AuHS° (Seward, 1973; Benning and Seward, 1996; Stefansson and Seward, 2004). The solubility of Fe, Ag, and base metals as chloride complexes was likely suppressed by elevated H_2S and low chloride concentrations. Solubility calculations (e.g., Hofstra et al., 1991; Hofstra, 1994) indicate that typical Carlin ore fluids with H_2S concentrations greater than ~10^{-2} m would have Au/Ag ratios >3, and this ratio would increase with increasing H_2S (Gammons, 1997). Such ratios are consistent with Au/Ag ratios of ore-stage pyrite and entire Carlin-type orebodies.

Ore and gangue deposition

The overwhelming occurrence of submicron Au in trace element-rich pyrite suggests that a single process controlled precipitation of most Au in Carlin-type deposits. Numerical modeling (Hofstra et al., 1991; Woitsekhowskaya and Peters, 1998) and geochemical studies (Jerritt Canyon, Hofstra, 1994, Phinisey et al., 1996, Hofstra et al., 1991, 1999; Twin Creeks, Stenger et al., 1998; Getchell, Cail and Cline, 2001; Gold Bar, Yigit et al., 2003) indicate that sulfidation processes best replicate ore-mineral assemblages, alteration assemblages, and Au grades. Sulfidation accounts for the residence of Au in pyrite and lack of introduced Fe that characterizes most ore zones (Hofstra and Cline, 2000). For example, particularly high-grade ore at Meikle correlates strongly with decarbonatized and sulfidized ferroan dolomite (Emsbo, 1999). Pyritization is indicated in portions of some deposits where Fe was clearly added to the rocks in addition to S. In these deposits, Fe may have been present in a resident solution that derived its Fe from reaction with nearby rocks. This resident fluid may have mixed with ore fluids in an "extended sulfidation" process, as proposed at Screamer (Kesler et al., 2003a). The general lack of visible Au in Carlin-type deposits and its ubiquitous residence in pyrite suggest Au coprecipitated with, or was adsorbed on, arsenian pyrite by a variety of reactions as a_{H_2S} decreased (Sha, 1993; Simon et al., 1999; Reich et al., 2005).

Reaction-path modeling, along with fluid inclusions in sequentially younger minerals that generally record declining temperatures (Cline and Hofstra, 2000; Hofstra and Cline, 2000), demonstrate that late-ore stage orpiment, realgar, and stibnite precipitated in response to cooling after deposition of Au-bearing pyrite (Hofstra et al., 1991; Woitsekhowskaya and Peters, 1998). Cooling at temperatures above ~180°C is consistent with formation of jasperoid with reticulate textures and drusy quartz. At temperatures significantly below ~180°C, however, fluids would have become supersaturated with respect to quartz (Rimstidt, 1997), leading to the precipitation of chalcedony or amorphous silica (Lovering, 1972; Nutt and Hofstra, 2003).

Depth of ore deposition

Early fluid inclusion studies suggested that Carlin-type deposits formed at depths generally greater than ~5 km, largely due to the presence of CO_2-rich fluid inclusions identified in quartz (Osterburg, 1990; Kuehn and Rose, 1995; Groff, 1996; Lamb and Cline, 1997). These inclusions are now, however, considered to be products of Mesozoic intrusive activity and are not indicative of conditions associated with Carlin-type mineralization (Cline and Hofstra, 2000; Hofstra and Cline, 2000). The lack of evidence for fluid boiling and related ore-fluid phase-equilibria constraints (summarized by Hofstra and Cline, 2000) indicate that ranges of minimum depths of formation were ~1.7 to 6.5 km and maximum depths probably no greater than 5 to 8 km. In the Getchell trend, fluid inclusions have higher CO_2 contents than most other districts (Cline and Hofstra, 2000), suggesting that deposits in this trend may have formed at greater depths.

Recent fission-track studies in the Carlin-Jerritt Canyon region yield estimates that overlap with the shallower depths indicated by fluid inclusion analyses. Thermal modeling of apatite fission-track age and track-length data yield time-temperature histories indicating that rocks currently exposed at the surface in the ranges north and south of the northern Carlin trend had largely cooled to <60°C by the Eocene (Fig. 10; Hickey et al., 2003a). Assuming a mean surface temperature of 10°C and a near-surface paleogeothermal gradient of 30°C/km, these data suggest that the present-day erosion surface lies, at most, ~1 to 2 km below the Eocene erosional surface (Fig. 9b; Hickey et al., 2003a). Lower paleogeothermal gradients of 20° to 25°C/km would increase this estimate to <2 to 3 km. Such shallow depths are supported by the presence of a spherulitic core and ~1-m glassy selvedges in the mineralized Deep Star rhyolite (T. Thompson, pers. commun., 2004).

The most shallow formed Carlin-type deposits in northern Nevada are those exposed at Alligator Ridge south of the Ruby Mountains. A late Eocene reconstruction of this area indicates that these relatively small deposits formed at depths of <300 to 800 m, and silicified Eocene conglomerates suggest that ore fluids discharged at or near the paleosurface (Nutt and Hofstra, 2003). Such shallow ore fluids must have contained less CO_2 than indicated by fluid inclusion studies at other deposits. Collectively, fluid inclusion data, apatite fission-track modeling, jasperoid textures, and geologic reconstructions indicate that the upper parts of the currently defined deposits formed over a range of depths from ~0.3 to ~3 km, and the deeper portions of some deposits may have formed as deep as 5 km (Hofstra and Cline, 2000, and references therein; Hickey et al., 2003a; Nutt and Hofstra, 2003).

Sources of ore fluid components: Isotopic studies

O and H sources: Oxygen and H isotope studies were conducted on fluid inclusions, hydrous minerals, and quartz at several deposits but have not produced results that identify a single ore-fluid source (App. Fig. A8). The lack of agreement could result from difficulties in identifying, collecting, and/or analyzing pure separates of ore-stage minerals or may reflect real differences in different districts, including greater or lesser dilution of a deep ore fluid or the presence of different proportions of different fluids in different deposits. Unresolved problems in determining the original H isotope signatures of fluid inclusion water in quartz may also contribute (Faure, 2003), although the indicated errors would not significantly change the interpreted sources of fluids in most studies.

Low δD_{H_2O} values, indicating that the water was largely meteoric in origin, were obtained in most studies (Hofstra et al., 1988, 1999; Yigit, 2001; Emsbo et al., 2003; Lubben, 2004). However, high δD_{H_2O} values from both fluid inclusions and clay minerals from Getchell (Groff, 1996; Shigehiro, 1999; Cline and Hofstra, 2000) indicate that much of the water of this deposit had a deep source, either magmatic or metamorphic (Fig. A8). Mantle-derived He in Getchell ore-stage fluid inclusions (Cline et al., 2003) supports this interpretation, although mantle He is also released along crustal-scale structures and in sedimentary basins undergoing extension (Kennedy et al., 1992, 1997). Other studies of clay minerals from two northern Carlin trend deposits yield conflicting data; some kaolinite analyses (Deep Star, Heitt et al., 2003; Deep Post, T. Thompson, pers. commun., 2004) indicate magmatic or metamorphic water, whereas other analyses (Deep Star, Hofstra, unpub. data) identified only meteoric water in kaolinite but magmatic or metamorphic water in pre-ore illite or smectite (Fig. A8). $\Delta^{18}O_{H_2O}$ values obtained from clay minerals and fluid inclusions at all deposits decrease from ore to late ore stage. The values and trend reflect late ore-stage infiltration of relatively unexchanged meteoric water that mixed with and diluted either deep magmatic or metamorphic water or an evolved meteoric water (Fig. A8; Radtke et al., 1980; Hausen and Park, 1986; Groff, 1996; Hulen and Collister, 1999; Hofstra and Cline, 2000). Although fluid inclusion and isotopic data provide evidence of mixing between more and less exchanged meteoric waters (e.g., Jerritt Canyon, Hofstra et al., 1988, 1999; Hofstra, 1994; Gold Bar, Yigit, 2001; Meikle, Emsbo et al., 2003) or between metamorphic and/or magmatic and variably exchanged meteoric waters (e.g., Getchell, Cline and Hofstra, 2000), mineral parageneses, mass-transfer studies, and chemical modeling indicate that most Au was deposited prior to mixing with local ground water, mainly in response to sulfidation of host-rock Fe (Hofstra et al., 1991; Woitsekhowskaya and Peters, 1998; Cline and Hofstra, 2000).

Results of studies to date provide good evidence for two waters, evolved meteoric and pristine meteoric, that are common to all Carlin-type deposits, and for a third, deeply sourced water in the Getchell trend and possibly in the northern Carlin trend. These results can be interpreted as indicating that deep magmatic or metamorphic water may have constituted the ore fluid that transported metals to all deposits but was diluted by evolved meteoric water during the ore stage in most deposits and districts, or ore-stage fluids in the Getchell trend and possibly some parts of the northern Carlin trend were dominated by deep magmatic or metamorphic water, whereas meteoric water dominated the ore fluids of most other systems. Less evolved meteoric water diluted ore fluids during the late ore stage in all districts.

S sources: As reduced S was likely the principal ligand responsible for Au transport in Carlin-type systems, S isotope data can provide insight into Au sources as Au cannot have been sourced at depths greater than S. The range of $\delta^{34}S$ values reported for main ore-stage pyrite in Carlin-type systems does not, however, identify a single, unambiguous reduced S source (App. Fig. A9). Evaluation of the data is complicated because analytical techniques vary from in situ ion-probe analyses of 10- to 35-μm-diameter spots to conventional analyses of pyrite separates or small to large bulk samples, and samples in many studies are not well described. Conventional analyses may be valid where the large majority of pyrite in a sample can be demonstrated to be ore stage. However, some studies undoubtedly analyzed mixtures of ore-stage and earlier or later formed pyrites since the fine Au-bearing pyrite is seldom the only pyrite generation in mineralized rock. In situ ion-probe analyses can provide meaningful results but only if the spots analyze single generations of paragenetically constrained pyrite.

The most reliable results from published studies have a broad S isotope range (Fig. A9), which requires a sedimentary source of S (Arehart et al., 1993a; Hofstra, 1997; Hofstra and Cline, 2000; Emsbo et al., 2003). In addition to potential analytical problems, the broad range of reported values may reflect S derived from different stratigraphic intervals or paleoenvironments or different mechanisms of H_2S generation, including dissolution of pyrite, destruction of organosulfur compounds, thermochemical sulfate reduction, desulfidation of pyrite to pyrrhotite, and inorganic sulfate reduction at different depths in the system. Ion-probe spot analyses of ore-stage pyrite from the Getchell deposit (Cline et al., 2003; Cline, unpub. data), and similar studies at the Betze-Post deposit (Screamer, Kesler et al., 2003a, b; Betze-Post, Henkelman, 2004), exhibit a small range of values (–1 to +7 ‰), consistent with magmatic S. Thus, a magmatic source may account for the lower values but not the higher values (Fig. A9) unless magmas assimilated sedimentary S, as in some Mesozoic igneous systems (Vikre, 2000).

Late ore-stage minerals are coarser and thus easier to sample than sulfides related to Au ore, and are generally not subject to the described problems; however, they do not provide information about main ore-stage fluids (Fig. A8). These minerals commonly have high $\delta^{34}S_{H_2S}$ values (Fig. A9) that likely were caused by incursion of relatively unexchanged meteoric water with H_2S leached from carbonaceous matter in sedimentary rocks (e.g., Emsbo and Hofstra, 2003). Alternatively, H_2S generated by thermochemical sulfate reduction of sedimentary barite may have been a source of some late ore-stage S (e.g., Arehart et al., 1993a; Hofstra, 1997; Emsbo et al., 2003). A postore shift to values that extend as low as ~–30 per mil in botryoidal pyrite and/or marcasite ± sphalerite veinlets (Fig. A9) in some deposits is consistent with cooling and oxidation of H_2S in hydrothermal fluids or previously deposited sulfide minerals by local ground waters as the systems collapsed (e.g., Hofstra and Cline, 2000) or during subsequent supergene weathering (e.g., Bawden et al., 2003), followed by bacterial sulfate reduction of the generated sulfate (e.g., Plumlee and Rye, 1992). Postore barites generally have $\delta^{34}S$ values less than syngenetic, diagenetic, and early vein barite values and probably obtained S largely by dissolution of preexisting barites. However, 2 Ma barites in the northern Carlin trend with $\delta^{34}S$ values near 15 per mil are interpreted to have formed by oxidation of H_2S derived from organic S in carbonaceous sedimentary rocks (Emsbo and Hofstra, 2003).

Pb and Os sources: Two sources for Pb were identified in the paragenetic sequence from Getchell and Turquoise Ridge (Tosdal et al., 1998, 2003; Hofstra et al., 2000). One source is represented by the Pb in most ore-stage pyrite, which has an isotopic composition similar to the average Neoproterozoic and Cambrian clastic rocks in eastern Nevada. This implies derivation from those rocks via fluid-rock interaction, a conclusion consistent with Sr and Nd isotope data for late ore-stage fluorite (Hofstra et al., 2000).

A second Pb source, typical of Ordovician to Devonian siliciclastic and calcareous rocks in the Carlin trend, dominates late ore-stage minerals from the north Carlin trend, Getchell, and Jerritt Canyon. $^{187}Os/^{188}Os$ ratios of ore-stage iron sulfide minerals and later ore-stage stibnite, orpiment, and realgar from Jerritt Canyon (Hofstra et al., 1999) are also within the range of average crustal rocks, consistent with sedimentary sources of Os.

Discussion

Deposit similarities and inconsistencies

Our detailed comparisons demonstrate that there is remarkable similarity between Carlin-type Au deposits in all districts in Nevada (Tables 2, A1, A2). Geologic processes active since the Neoproterozoic produced a geologic architecture composed of host rocks, structures, and physical and chemical traps conducive to mineralization. Geochronologic and geologic studies show that deposits formed during a short time interval across a broad region exposed to incipient upper-crustal extension and deformation as well as a thermal pulse generated by magmatic activity. Although isotopic differences exist at some deposits, detailed studies show that all districts display broadly similar styles of mineralization and alteration. All of these observations lead to the conclusion that regional processes produced ore fluids that reacted with favorable and generally similar host rocks in all districts and similar processes precipitated Au.

Nevertheless, isotopic studies imply that water of various sources contributed to form ore fluids in different deposits or districts or, in some instances, a single deposit. Interpreted simply, Getchell provides the strongest argument for magmatic S and magmatic or metamorphic water contributions to an ore fluid. If a "deep fluid" model applies to all districts, all deposits are related to crustal-scale magmatic and/or metamorphic processes, and meteoric water overwhelmed the systems at upper crustal levels. Alternatively, meteoric water may have evolved to become an ore fluid via reactions with sedimentary rocks (Seedorff and Barton, 2004) and deposited ore in several districts that is similar to ore deposited by magmatic or metamorphic waters at Getchell.

Viable models

Inconsistencies in the data have led to a number of models that relate mineralization to (1) metal leaching and transport by convecting meteoric water (Ilchik and Barton, 1997; Emsbo et al., 2003), (2) epizonal intrusions (Henry and Boden, 1998; Chakurian et al., 2003; Johnston and Ressel, 2004), or (3) deep metamorphic and/or magmatic fluids (Seedorff, 1991; Hofstra and Cline, 2000; Heinrich, 2005; App. Table A4).

Meteoric water model: Leaching models call upon either lateral flow of meteoric water with scavenging of Au from a sedex source (e.g., Emsbo et al., 2003) or deep crustal meteoric convection and leaching of Neoproterozoic source rocks in the crust (e.g., Seedorff, 1991; Ilchik and Barton, 1997). Much isotopic data is consistent with such models, and there is evidence for circulation of synextensional meteoric water in metamorphic core complexes to depths of 10 to 15 km (Lee et al., 1984; Fricke et al., 1992). These models do not, however, account for the presence of deeply sourced magmatic and/or metamorphic water or metals identified in some districts.

Epizonal intrusion model: Models that relate deposits to epizonal plutons suggest that Carlin deposits are either sedimentary equivalents of volcanic-hosted epithermal deposits or are distally related to porphyry-style intrusions (e.g., Cunningham et al., 2004; Table A4). Such models are regarded as viable because Carlin-type deposits in some districts are spatially associated with dikes of similar age (Ressel et al., 2000; Heitt et al., 2003) and because the deposits are similar to distal disseminated (Cox and Singer, 1990, 1992) or Carlin-like deposits (Seedorff, 1991).

A genetic relationship between Carlin-type deposits and epizonal porphyry-type intrusions requires a significant component of lateral flow of ore fluids (e.g., Fig. 1, Hedenquist et al., 2000), which would generate mineral and alteration zoning. The Carlin districts of Nevada have been exposed through exploration and mining for distances of several tens of kilometers and at least a kilometer vertically, yet no genetic relationships, zoning patterns, or vectors toward known coeval hypabyssal stocks that would support a magmatic connection have been identified. Carlin-type deposits and districts exhibit no regional-scale mineralogical or alteration zoning, whereas distal disseminated deposits are zoned peripheral to temporally related intrusive centers, even though distances from the intrusion may be significant (Cunningham et al., 2004). Furthermore, there is no evidence of any Eocene hydrothermal activity that exceeded ~250°C along the 20- to 35-km-long districts of very similar deposits, despite widespread and continuous drilling. Although Carlin-type deposits have trace element signatures similar to some epithermal deposits, Carlin deposits do not show any zonation at the present-day surface, unlike all epithermal deposits, and there is no evidence for fluid boiling, suggesting that they formed at greater depths than epithermal systems.

The Getchell deposit displays the greatest affinity for an igneous source based on stable isotope studies. However, Eocene igneous rocks have not been identified in this district although Eocene volcanic rocks are present ~7 km northeast of Chimney Creek (Wallace, 1993). Alternatively, in other districts where Eocene dikes are present, the water that comprised the ore fluids, with a few exceptions, appears to be meteoric, and the wide range of S isotope values is interpreted to reflect a dominant sedimentary source. All of these observations and data suggest that intrusions, if present, were remote and deep.

Deeply sourced ore-fluid model: Models that call upon deep magmatic or metamorphic sources for water and other components of the ore fluids are supported by isotopic data from some deposits, as indicated by ion-probe analyses of some ore pyrites, the spatial association of deposits and districts with crustal faults, and the huge scale of hydrothermal activity and amount of Au in the larger districts. Arguments against a deep-source model include stable isotope data that indicate sedimentary and meteoric sources for some components in some deposits, the lack of identified coeval igneous activity or intrusive centers in some districts, and the lack of demonstrated metamorphism during the Eocene.

A metamorphic fluid model requires prograde metamorphic devolatilization of sedimentary rocks to generate water to form ore fluids, with focused discharge along crustal fault zones to epizonal depths during episodes of extension. Such a model is appealing because it explains the spatial association of deposits and crustal-scale structures along an ancient continental margin, as well as radiogenic and stable isotopes that indicate deep sources for some ore-fluid components. Such a model can also account for Au mineralization of the scale found in the Nevada districts and deposits. However, peak metamorphism in midcrustal rocks exposed in metamorphic core complexes in the region was achieved during the Cretaceous or earlier. Storage of Cretaceous metamorphic fluids has been proposed (Seedorff, 1991) but is difficult to demonstrate.

A deep magmatic model is consistent with geologic reconstructions and apatite fission-track data that, respectively, indicate that some of the largest deposits are located in regions of paleotopographic highs and are associated with thermal anomalies; magmatic heat may have driven fluids to those topographic highs. Recent numerical modeling (Heinrich et al., 2004; Heinrich, 2005) suggests that magmatic fluids generated at depths perhaps as deep as 10 km (Lowenstern, 2001; Heinrich, pers. commun., 2004) can evolve into Carlin ore fluids through aqueous fluid immiscibility and separation of vapor containing volatiles and bisulfide-complexed Au, from brine containing chloride-complexed base metals. The low-density vapor ascends to upper-crustal levels along a high-pressure and near-neutral pH fluid pathway, which limits further fluid immiscibility and retains Au in solution until the ore fluid sulfidizes Fe derived from host rocks (Heinrich et al., 2004). Problems with a deep magmatic model include lack of evidence for cooling of the ore fluid, such as skarn mineralization or quartz veins, in conduits below the deposits.

Au-only deposits

Phillips and Powell (1993) noted that Au-only deposits have many characteristics in common with Carlin-type deposits, including low salinity, elevated CO_2, high Au/Ag, high Au/base metals, elevated Au-As-Sb, temperatures between 200° and 450°C, lack of consistent alteration zoning, associated Au and C, and Au deposition coincident with regional thermal events. They further conclude that Au-only systems formed from a uniform ore fluid that required a large and uniform source and suggest that deep, crustal-scale processes could best generate such a fluid. We concur and conclude that deep, crustal-scale processes are significant in generating ore fluids in Carlin-type systems, given the size of the Carlin Au belt in Nevada, the volume of Au in the deposits, and the large-scale geologic features that control deposit locations. We also recognize, however, that the pyritic, carbonaceous, ±baritic nature of the stratigraphic sequence in this part of Nevada, together with the presence of older Au deposits and enrichments, permits additional scavenging of metals and S from upper-crustal rocks by fluids rich in H_2S. For these reasons, leaching and reprecipitation of Au may have been important in some districts and may have contributed to the huge endowment of Au in these deposits.

Unique geologic history?

One of the most important questions is why only a single region in the world has been discovered that contains this type of sediment-host disseminated Au deposit, with this magnitude of contained Au, the largest after the Witwatersrand region in South Africa. One hypothesis is that a unique geologic history in northern Nevada was essential for this major Au district to form. If this hypothesis is correct, geologic processes from the Neoproterozoic to the late Eocene, necessary to produce ideal host rocks and structures, converged with optimum geologic processes during the Eocene to deposit Au. The critical aspects of the tectonic history include continental rifting and formation of a passive margin, which placed silty and pyritic carbonate rocks above basement-penetrating fault systems. Subsequent compressional events were important in forming regional thrusts that placed less permeable and less reactive siliciclastic rocks above the reactive carbonate rocks. Inversion of rift-related faults during compression created structural culminations directly over basement-penetrating faults that later served as major conduits for deep ore fluids. Also critical was a period of major continental extension and associated magmatism that drove circulation of deep crustal fluids into the upper crust. Similar but smaller or isolated deposits found elsewhere in the world may have formed from processes similar to those in the late Eocene of Nevada, but without the requisite host rocks and structure, or may have resulted from less ore fluid or thermal energy, a smaller Au budget, or some combination of these and other limiting factors. If this hypothesis is correct, the possibility of discovering districts and deposits of similar size elsewhere in the world may be limited.

Alternatively, other major districts of Carlin-type deposits may remain to be discovered, unrecognized as of yet because of the subtle nature of the mineralization and alteration, particularly if the deposits have not been exposed to supergene alteration. Our observations suggest that opportunities to discover Carlin-type districts outside of Nevada are increased in areas where passive margin carbonate sequences were compressed and subsequently extended with associated magmatism. Future opportunities for discovery may improve after we identify the critical ingredients—preexisting stratigraphic and structural architecture, earlier Au mineralization, deep metamorphism and/or magmatic activity, coeval upper-crustal igneous activity, incipient extensional tectonism—essential to the formation of large Carlin-type deposits.

Genetic Model for Nevada Deposits

We propose the following model for the formation of the deposits (Fig. 12), based on similarities observed in Carlin-type deposits in all districts in Nevada (Tables 2, A1, A2) and which attempts to reconcile the differences for some deposits. The model is consistent with geologic observations and our current understanding of the complex geologic history of this part of Nevada and allows for Au to be sourced from several locations in the crust.

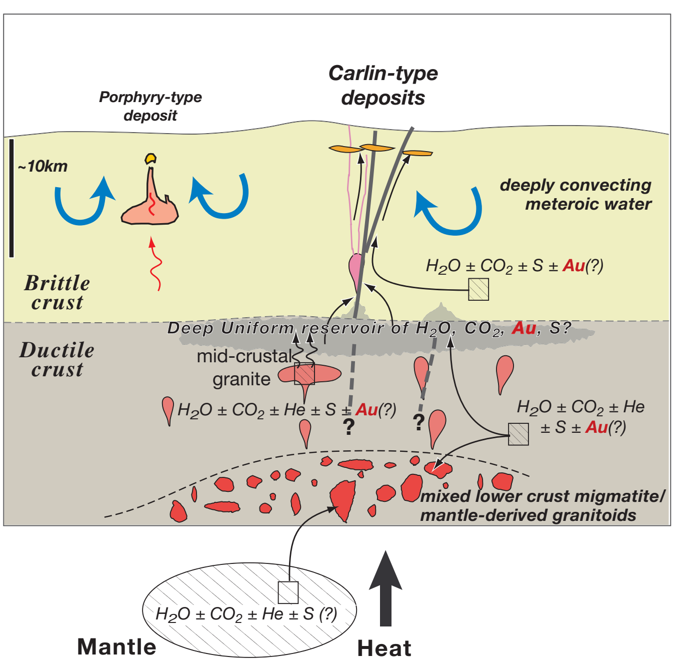


FIG. 12. Model illustrating processes associated with generation of hydrothermal fluids that evolved to become ore fluids for Carlin-type deposits, showing locations within the crust that may have contributed Au and other ore-fluid components. See text for discussion.

During the Eocene, asthenosphere was reintroduced to the base of North American lithosphere as the shallow Farallon plate was removed (Humphreys, 1995; Westaway, 1999; Humphreys et al., 2003). This activity generated high K calc-alkaline magmatism that swept southward through the latitude of the Carlin deposits at ~42 Ma. As the Farallon plate was removed, mantle-derived mafic magmas were injected into lower crust, generating partial melting and transferring mantle-derived volatiles with juvenile isotopic signatures into the crust (Fig. 12). Associated prograde metamorphism and devolatilization of lower crust probably released additional volatile constituents, possibly including Au, which were incorporated into lower crustal melts. As the melts rose buoyantly, they eventually became saturated with volatiles and exsolved hydrothermal fluids that may have transported bisulfide-complexed Au. Exsolved hydrothermal fluids, with possible metamorphic fluid contributions, continued to move upward and evolve compositionally as they scavenged or fixed various components along fluid pathways. Au along with As, Sb, Hg, S, and trace metals such as Pb may have been scavenged from Neoproterozoic rocks, particularly pelitic beds (Seedorff, 1991).

As overthickened upper crust began to extend, aqueous hydrothermal fluids migrated into and rose along dilatant faults associated with reopened Proterozoic rift-related structures (Fig. 12). Reactions between carbonaceous, pyritic, baritic, calcareous rocks and ascending fluids may have increased H_2S concentrations, thus increasing the capacity of the fluid to scavenge Au. In the northern Carlin trend, ascent of auriferous fluids appears to be temporally associated with a transition from early, broad-scale extension, to a stage of pervasive rotational faulting. Ore fluids here further appear to have been driven to topographic highs, perhaps in response to thermal input from upper crustal (6–10 km deep) plutonic complexes. In most districts, ore fluids were diluted by deeply convecting meteoric waters.

Ore fluids accumulated in areas of reduced effective mean stress along boundaries of older Jurassic and Cretaceous stocks, and in structural culminations where aquitards focused, diverted, or trapped fluids, promoting increased fluid/rock reaction. Reactive fluids decarbonatized and argillized wall rocks, further enhancing permeability and exposing and sulfidizing available reactive host-rock Fe. Pyrite precipitation decreased the a_{H_2S} in the ore fluids, thereby causing coprecipitation or adsorption of Au and other bisulfide-complexed metals, and Au was incorporated in trace element-rich pyrite as submicron particles of native Au or structurally bound Au.

Eventual reduced flow of ore fluids and collapse of unexchanged meteoric water into the system caused fluid mixing and cooling and precipitation of late ore-stage minerals. Metals derived from local siliciclastic and calcareous rocks were incorporated in late ore-stage minerals and outer pyrite rims. Late calcite veins precipitated above deposits or overprinted ore-stage mineralization as fluid reactivity was neutralized. Spent, dilute, low-temperature ore fluids exited ore zones, locally forming unmineralized jasperoids in exhaust structures.

Implications of global occurrences for the origin of Carlin-type deposits

One of the unique attributes of Carlin-type deposits and districts is their large size (100s to 1000s t Au). Although porphyry Cu-Mo-Au and reduced intrusion-related systems can have comparable endowments of Au (e.g., Bingham Canyon district with 1,600 t, Fairbanks district with 1,250 t), most Au is proximal to the intrusion (Krahulec, 1997; Thompson and Newberry, 2000), and none of the distal disseminated deposits associated with intrusion-related systems have Au endowments comparable to large Carlin-type deposits and districts. To explain the trends of Carlin-type deposits in Nevada, a Bingham Canyon- or Fairbanks-size system or systems would have to discharge and fix their entire endowment of Au in sedimentary rocks at distal, shallower levels and leave uncertain or no evidence of its magmatic origin.

The only sedimentary rock-hosted disseminated Au deposit types in other parts of the world with similar endowments of Au are epizonal orogenic deposits (A. Hofstra and P. Emsbo, unpub. synthesis, 2004). Large orogenic Au districts also have ore fluids with chemical and isotopic compositions that are similar (Ridley and Diamond, 2000; Goldfarb et al., 2005) to those in Carlin-type deposits. But orogenic deposits typically occur in deformed eugeoclinal or flysch basin sequences on the forearc side of convergent margins (Goldfarb et al., 2001), rather than in carbonate rocks on the foreland side where Nevada Carlin-type districts are located. Although they are not as large, the sedimentary rock-hosted disseminated Au deposits in the West Qinling belt (Gu, 1996; Wang and Zhang, 2001; Mao et al., 2002, Vielreicher et al., 2003) and Dian-Qian-Gui area of southern China (Ashley et al., 1991; Liu et al., 1994; Wang, 1996; Hu et al., 2002; Zhang et al., 2003; Wilde et al., 2003) are similar to both epizonal orogenic deposits and Nevada Carlin-type deposits and formed primarily from metamorphic fluids that mixed with local meteoric water (Liu et al., 1994, Wang, 1996, Wang and Zhang, 2001, Hu et al., 2002, Zhang et al., 2003; Hofstra et al., unpub. data) While the tectonic settings in southern China and Nevada differ, it appears that deep metamorphic plus magmatic fluids were generated in each region and channeled into miogeoclinal sedimentary rocks where they produced deposits with similar characteristics. The large size of some epizonal Au deposits and the involvement of metamorphic fluids in the formation of sedimentary rock-hosted disseminated Au deposits in southern China, by analogy, support a deep fluid source model for Carlin-type deposits.

Critical Questions and Areas for Future Research

The preferred genetic model proposed here is supported by permissive, rather than conclusive evidence, and other models for these deposits remain viable. The model that each geologist applies in the search for these deposits will depend on the relative value they give to the information currently available and to their interpretations and biases in areas where available information is conflicting. A uniform model for all deposits in Nevada, which seems desirable given the similar nature of the deposits and districts, and the relatively small region over which they occur, currently requires some apparently credible data to be discounted. Acceptance of all available well-documented data requires different models for different districts and perhaps for different deposits within a single district. By designing future studies to resolve the inconsistencies in the available database, we may be able to resolve many of our remaining questions. Future researchers can improve our understanding of these deposits by conducting studies that address the following issues:

1. Refine timing relationships between mineralization, extension, and igneous activity at various deposits and the differences or similarities from district to district to increase our understanding of the trends and to identify critical mechanisms that triggered influx of ore fluids to the sites of metal deposition.
2. Prepare palinspastic reconstructions of mineralized districts to improve our knowledge of the geologic history during the period of ore formation, aid in constraining the hydrology of the ore system, and contribute to refining the genetic model.
3. Expand geological and geochemical studies to document the presence of other styles of mineralization in rocks lateral to the deposits, or in deeper parts of the stratigraphic package within major trends and districts, to identify potential source rocks responsible for the deposits.
4. Conduct detailed studies at the deposit scale, starting with careful determination of mineral paragenesis followed by fluid inclusion and stable, noble gas, and radiogenic isotopic studies to improve our understanding of the interplay of fluid mixing, cooling, fluid-rock interaction, and sulfidation, and their effects on Au deposition and the sources of fluids and metals. Microanalyses of minerals clearly related to various paragenetic stages can be key to answering these questions.
5. Conduct integrated hydrologic and chemical reaction-path modeling to distinguish viable models from models that are demonstrably untenable.
6. Continue to study deposits with similar characteristics elsewhere in the world to better understand differences and similarities between these systems.

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