GREAT BASIN-COLORADO PLATEAU TRANSITION IN CENTRAL UTAH: AN INTERFACE BETWEEN ACTIVE EXTENSION AND STABLE INTERIOR

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ABSTRACT

A fundamental tectonic boundary appears to have existed below the site of the present-day Colorado Plateau to Great Basin Transition Zone since Precambrian times. The Plateau proper has seen little deformation since Middle Proterozoic continental assembly apart from Cenozoic uplift and limited thick-skinned contraction and calc-alkaline plutonism. In contrast, the Great Basin region has been subject to repeated episodes of both contractional and extensional tectonism, and extensional activity continues into the modern day. Evidence exists that the Colorado Plateau at its western margin is being converted to lithosphere with rifted Great Basin properties. Some models for migrating extension call upon progressive gravitational collapse of thicker crust of the plateau margin as it warms, possibly aided by hardening of the previously rifted lithosphere (i.e., Great Basin interior) via crustal thinning and cooling.

However, this rather homogeneous and temporally gradual model of deformation has only partial applicability to evolution of the western Colorado Plateau and eastern Great Basin. On the one hand, the limited degree of block style faulting, high elevation, and high apparent elastic thickness of the Transition Zone resemble properties of the Colorado Plateau. On the other, heat flow, upper mantle velocities, and deep electrical conductivity of the transition are more like those of the active eastern Great Basin. We suggest that non-uniform extension occurs below the Transition Zone, with thinning of the lower crust and mantle lithosphere by a degree greater than that of the upper crust. Transition Zone extension may be distinctly earlier than that of the eastern Great Basin, with exhumation of the latter occuring quite uniformly in the Early to Middle Miocene, but the former perhaps initiating significantly later (Late Miocene-Early Pliocene). Estimates of total extension and lower crustal material movement for the eastern Great Basin in the Middle Miocene range widely depending upon the assumed role of the Sevier Desert detachment, but present-day extension appears unrelated to this feature.

At present, we lack basic geophysical and geological data to resolve even gross contributions of force versus strength, let alone individual force or strength components, in driving deformation of the Transition Zone. Gravitational forces are expected to dominate, while Pacific plate boundary forces are considered negligible, but impact of the Colorado Plateau lithospheric keel upon elevated Great Basin asthenosphere may contribute. Strength indicators for the Transition Zone are conflicting. Uncertainties about the roles of low-angle detachments versus lower crustal flow in Transition Zone rifting reflect a limited knowledge of crust/mantle structure, and thus of processes of extensional consumption of formerly stable lithosphere here. The role of magmatism in the Transition Zone remains cryptic also. Some seismic models suggest a high-velocity “rift pillow” near the Transition Zone, which would imply focused extension and magmatism here, while others show only gradual progression in crustal thickness from west to east. The southern part of the Transition Zone has been highly modified by prior calc-alkaline plutonism which significantly affects upper crustal deformation style, but its influence on whole-lithosphere strength is unclear.

INTRODUCTION

Deformation of continental crust and upper mantle under extensional stress is a fundamental issue in Earth evolution, but the process shows a wide range of manifestations (e.g., Buck, 1991; Ruppel, 1995). Nowhere does an interplay between force and strength in this context appear more likely than in the Great Basin-Colorado Plateau (GB-CP) tectonic transition in Utah. To the west, the Great Basin easily comprises the widest active rift zone in the world, with high heat flow, mafic magmatism, thin crust, and high elevation (Figures 1 and 2; Parsons, 1995). To the east lies the apparently competent lithosphere of the CP interior which has experienced little surface deformation since the Proterozoic. The interface between these provinces comprises an uplifted rift shoulder (the Wasatch Front and southward extension) and a 100 km-wide Transition Zone (TZ), and vicinity. These areas have experienced previous episodes of extensional, compressional and magmatic activity since the Precambrian, usually with north-south orientation similar to current deformation. History suggests either that the Colorado Plateau to the east is unusually and persistently strong, or that it has somehow escaped the deformational forces which have so perturbed its surroundings. Nevertheless, direct evidence on process is ambiguous; this cryptic transition zone

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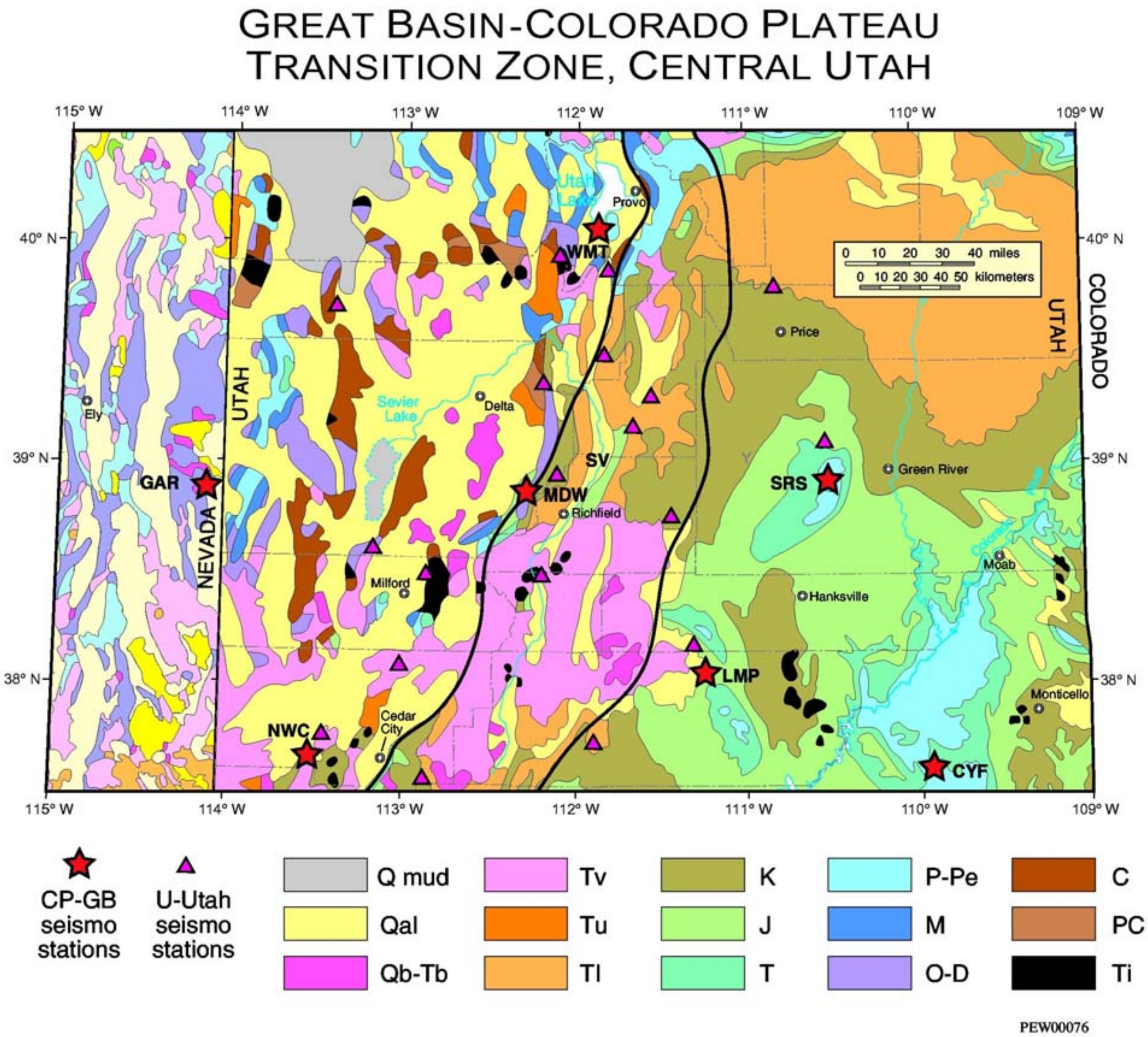

Figure 1. *Generalized geologic map of central Utah and easternmost Nevada. Red stars show 1994-5 CPGB broadband seismometers (Sheehan et al., 1997) and pink triangles show University of Utah permanent seismic network. Geological color scheme applies only to Utah. SV is Sevier Valley. Dark lines in central Utah enclose GB-CP Transition Zone, defined on the west by the physiographic transition (Fenneman and Johnson, 1946) and on the east by the eastern limit of normal faulting.*

possesses some characteristics much like the extending eastern Great Basin but others much like the stable Colorado Plateau.

One of our goals here is to show just how little is known, as well as how much, about the Great Basin-Colorado Plateau transition. Our views boil down to three principal issues about continental extension and its controls. *First*, some continental blocks appear to escape deformation, while neighboring areas which lie across sometimes abrupt boundaries are highly deformed (either in extension or compression). Other examples include the Tarim-Szechuan-Ordos Basins of China. Current data on the relative roles of strength and force heterogeneity for the Great Basin-Colorado Plateau TZ, particularly density and rheology estimates, are still very ambiguous. *Second*, some process of lithospheric consumption appears to be operating at this transition. Possible mechanisms range from uniform gravitational collapse of overthickened crust, to highly non-uniform rifting with shallow-angle detachment (perhaps diffuse) separating intervals of differing extension with increased depth toward the eastern TZ. A comparable extensional example includes the Sierra Nevada overlying the downdip continuation of rifting in the Death Valley region. *Third*, there is a large possible role for magmatism in extension of the continental lithosphere here. This issue is two-fold. First, tentative evidence exists for a rift pillow under the Transition Zone, which may reflect enhanced extensional processes in the upper mantle similar to those of volcanic continent-ocean margins. Second, deep extensional processes may be affected by intense modification of portions of TZ lithosphere by previous, mid-Cenozoic, calc-alkaline plutonism. Impact of the plutonism on shallow extensional faulting is obvious in such areas.

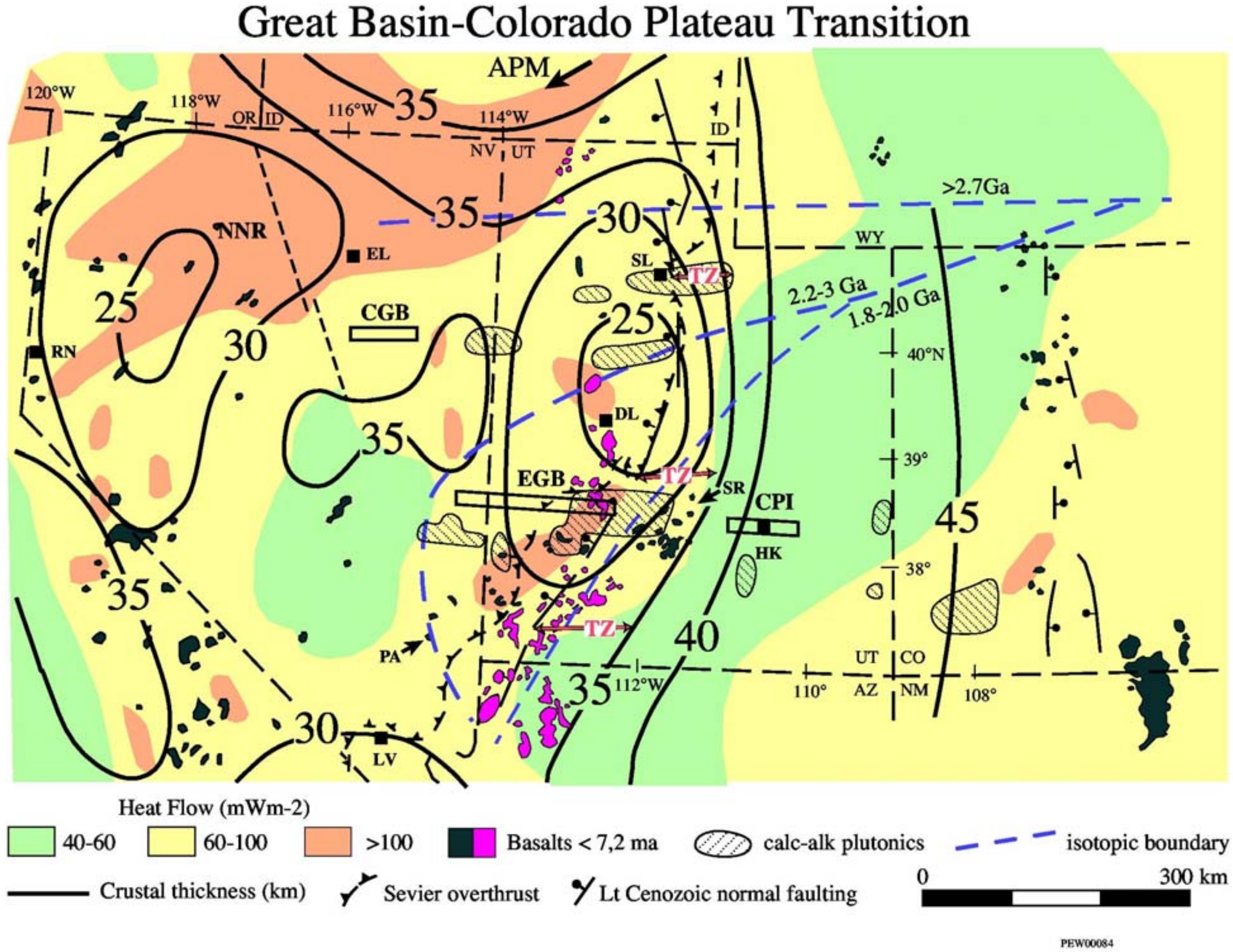

Figure 2. *Generalized geophysical location map for Great Basin and Colorado Plateau. Included also are heat flow variations (orange-green), occurence of volcanics less than 7 Ma (purple), Quaternary volcanics of N-S Sevier Desert-Grand Canyon trend (red), east-west calc-alkaline magmatic belts (grey hachures), and crustal thickness contours after Braile et al. (1974) and Wernicke (1985). Approximate extent of faulted Transition Zone denoted with TZ. MT profiling shown as open rectangles (CGB, EGB, CPI). Urban centers include Salt Lake City (SL), Elko (EL), Reno (RN), Delta (DL) and Las Vegas (LV). Modified from Wannamaker et al. (1997b), Smith et al. (1989), Blackwell et al., 1991; Ozalabey et al. (1997), and Nelson and Tingey (1997). Direction and rate of North American absolute plate motion (APM) in hot-spot frame shown with large arrow. Pahranagat-San Rafael volcanic trend is PA-SR.*

GEOLOGICAL BACKGROUND TO THE EASTERN GREAT BASIN AND COLORADO PLATEAU REGION

Distinct geologic histories for the eastern Great Basin (GB) and the Colorado Plateau (CP) exist back at least to the rifting of a continent westward away from North America in the Late Precambrian (Burchfiel et al., 1992; Karlstrom et al., 1999; Burrett and Berry, 2000). Although this long-lived contrast likely exerts control on current activity, we attempt to define a logical separation in time between this history and the profound extensional events creating the remarkable present-day transition. This occurs in the Oligocene period, so that we view the Transition Zone largely as evolution of geological/ geophysical structure since 25-30 Ma.

Pre-Great Basin Tectonic Heritage

The Phanerozoic Great Basin setting was a west-facing, Atlantic-style passive margin with a miogeoclinal sedimentary section, developing from Late Proterozoic through Devonian time, over rifted, Middle Proterozoic crystalline basement (Poole et al., 1992; Van Schmus et al., 1993). Section thickness reaches ~15 km in central Nevada (Stewart, 1980). By contrast, the Colorado Plateau to the east accumulated only a thin cratonal sedimentary cover in the Phanerozoic. This Late Proterozoic rift boundary may not be the first major structural break bordering the GB-CP transition. From Proterozoic exposures and structural trends in central Arizona, Karlstrom and Williams (1998) consider the Mojave (2.0-2.3 Ga)-Yavapai (1.7-1.85 Ga) province boundary to lie near the GB-CP transition in Utah (Figure 2). However,

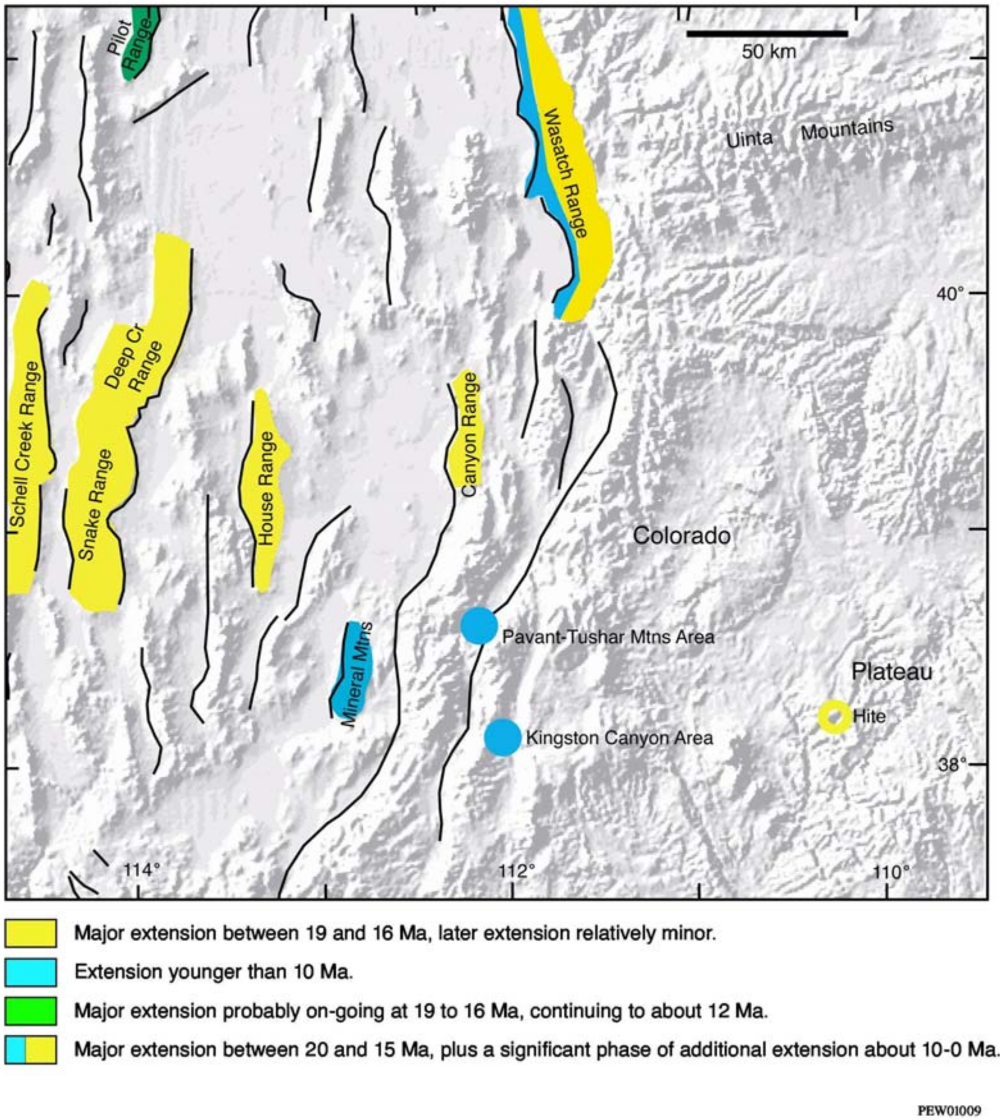

Figure 3. *Map of eastern Great Basin and Colorado Plateau provinces showing timing of major episodes of Miocene extension derived from fission-track data and other published studies (Rowley et al., 1979; 1981; Kowallis et al., 1990; Coleman et al., 1997; Miller et al., 1999; Stockli, 1999; Stockli et al., 1999, 2001; Dumitru et al., 1997, 2000; Dumitru, 2000; Armstrong et al., 2000). Note: many ranges exhibit multiple episodes of extension; FT ages here are those recorded by apatite system for cooling from ~120 to 60°C. Basemap courtesy of Joseph Colgan derived from the GLOBE database (GLOBE Task Team et al., 1999).*

based on isotopic characteristics of Precambrian basement and Miocene plutons in S.W. Utah (Mineral Mtns), others draw the Mojave-Yavapai boundary well to the west of the TZ in the Great Basin (Figure 2) (Bennett and DePaolo, 1987; Coleman and Walker, 1992).

The most apparent pre-extensional deformation affecting the region was the Late Mesozoic, Sevier fold and thrust event (Armstrong, 1968, 1982; Coney and Harms, 1984; Allmendinger, 1992). Crustal thickness in eastern Nevada and western Utah may have reached 50-60 km (Patino Douce et al., 1990), and rocks as young as mid-Mesozoic may lie underthrust as far west as 100 km of the thrust front (Wannamaker et al., 1997a). The frontal thrusts surfaced through the rapidly thinning Phanerozoic section at the east edge of the Great Basin, thus exhibiting control by the Late Precambrian margin rifting. The related but slightly lagging, thick-skinned Laramide (Late Cretaceous-Early Eocene) episode caused gentle, broad-wavelength monoclinal uplifts in the Colorado Plateau (Davis, 1978; Hamilton, 1988; Miller et al., 1992). Shallow Farallon subduction with accelerated plate convergence is the ascribed cause, and coincides with near absence of magmatism (Armstrong and Ward, 1991; Dumitru et al., 1991).

The Middle Eocene drop in North American-Pacific convergence rate foreshadowed collapse of the Sevier over-

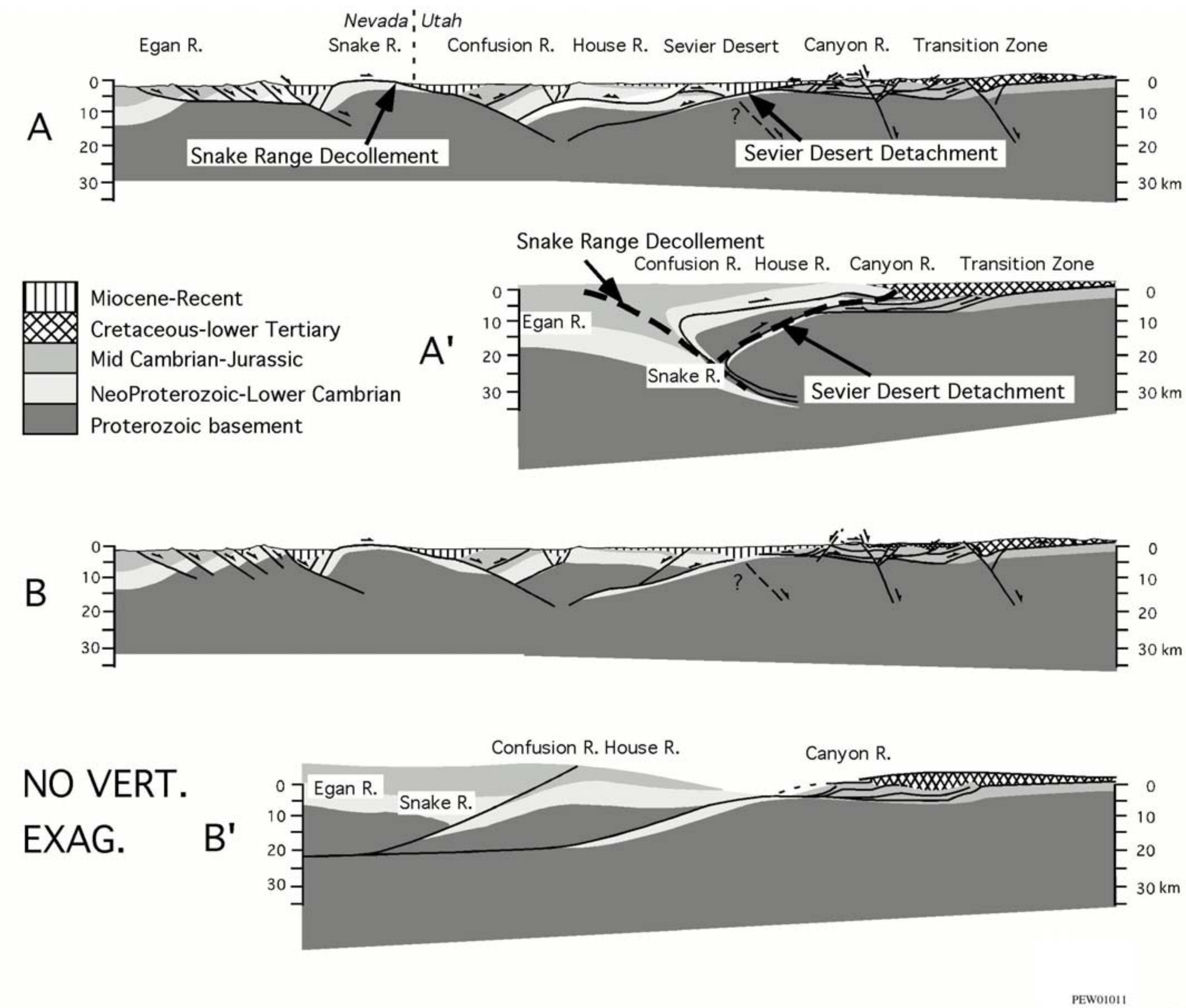

Figure 4. *Alternative cross sections and restorations along latitude N39.5°. A), Simplified, present-day cross section across western Utah and eastern Nevada at latitude N39.5°, modified from Allmendinger (1992) and Bartley and Wernicke (1984). A'), Cross section in A with Oligocene-Recent extension removed. Restoration of eastern half based on Allmendinger (1992) except extension in Transition Zone is interpreted to root eastward under the Colorado Plateau; western half based on Lewis et al. (1999) and Bartley and Wernicke (1984). B), Same as A), except that the Sevier Desert reflector is interpreted as an erosional unconformity rather than an extensional detachment (Anders and Christie-Blick, 1994; Wills and Anders, 1999) and, more generally, no low-angle normal faults other than rotated high-angle faults are present. B'), Cross section in B with Oligocene-Recent extension removed. Restoration after Allmendinger (1992) to be consistent with an erosional rather than tectonic Sevier Desert basin. Also, neo-Proterozoic strata of the Snake Range are interpreted to be near their stratigraphic depth (Miller et al., 1983; Miller et al., 1999) rather than having been tectonically buried to mid-crustal depths (cf. Bartley and Wernicke, 1984; Lewis et al., 1999).*

thrust wedge (Christiansen and Yeats, 1992; Constenius, 1996). In part coeval and spatially associated, regional calc-alkaline magmatism and metamorphic core complex development swept from the northernmost U.S. to the southern Great Basin (37-38°N lat.) by the early Miocene (Armstrong and Ward, 1991). In detail, the magmatic sweep occurs as discrete, E-W trending belts with intrusive activity spanning up to 10 Ma in each, and the spatial-temporal association with crustal extension often is rather poor (Best and Christiansen, 1991; Wernicke, 1992; Axen et al., 1993; Gans and Bohrson, 1998). In Utah and Nevada these belts are (from the north) the Late Eocene Oquirrh-Uinta, Oligocene Deep Creek-Tintic, and Late Oligocene-Early Miocene Pioche-Marysvale belts (or similar names; Stewart et al., 1977; Rowley, 1998), with intervening magmatic gaps (e.g., the "Mid-Utah magmatic gap" of Stewart et al., 1977) (Figures 1 and 2).

Nd isotopic compositions clearly indicate that most source material for the Mid-Cenozoic plutonism was not asthenospheric, but derived primarily from the Proterozoic mantle lithosphere or lower crust (e.g., Coleman and Walker, 1992; Hawkesworth et al., 1995; Riciputi et al., 1995; Askren et al., 1997; Vogel et al., 1998). The heat source for this magmatic flareup may have been disappearance of the Farallon Plate and its replacement by hot asthenosphere, perhaps to a shallower extent even than today (Perry et al., 1993; Humphreys, 1995). The only contemporary plutonism in the CP is the San Juan-LaSal-Abajo-Henry Mountains belt in southern Utah and Colorado, and crustal extension is strikingly absent (Nelson et al., 1992; Nelson and Davidson, 1993). Nevertheless, on composition and age grounds, Nelson et al. unify these fields into the San Juan-Marysvale-Reno magmatic belt, the longest of all such defined. In Oligocene time prior to the intrusive belt, south-central Utah enjoyed gentle structural relief (Hose, 1977; Rowley et al., 1978, 1979; Best and Christiansen, 1991).

Miocene to Present Extensional Regime

Middle Miocene appearance of block-style tectonism and basaltic magmas in the GB is thought to reflect cooling and embrittlement of the lithosphere since the calc-alkaline pulse and associated shallowing of the asthenosphere in the Mid-Cenozoic (e.g., Wernicke, 1992; Perry et al., 1993; Harry et al., 1993; Zoback et al., 1994). New fission track (FT) and Ar/Ar geochronology integrated with extensive geologic mapping show that Great Basin exhumation was essentially complete by Early-Middle Miocene, and later extension has been of much lesser magnitude (summarized for eastern GB in Figure 3) (Miller et al., 1999; Stockli, 1999; Stockli et al., 1999, 2001; Dumitru, 2000; Dumitru et al., 2000). Although core complex-style deformation in the Snake Range line initiated in the Late Oligocene, most unroofing here and in the Deep Creek Range occured very rapidly ~17 Ma along a 150-km-long, north-south trending fault system (Lee, 1995; Axen and Bartley, 1997; Miller et al., 1999). Slip along this low-angle decollement and exhumation of extensive footwall mylonites were coeval with major high-angle rotational faulting in the southern Snake Range and Deep Creek Range, indicating that the two styles of faulting can occur simultaneously along the length of a single normal fault system. By the Late Miocene, the most active extension of the Great Basin had shifted from being province-wide or affecting primarily the interior, to being concentrated along its eastern and western margins (Christiansen and Yeats, 1992; Wernicke, 1992; Wannamaker et al., 1997b). Reconnaissance FT data on the Colorado Plateau near Hite (Figure 3) suggest major erosional stripping ~16 Ma (Dumitru et al., 1994).

Actual geometry, kinematics, time of initiation, and current and past displacement rates of upper-crustal Cenozoic structures in the Transition Zone remain poorly known. Exhumation of the Wasatch Front rift shoulder in the Salt Lake City area north of this study is diachronous (Armstrong et al., 2000) (Figure 3), but shows events younger than typical range formation to the west. The Mineral Range on the southern profile uplifted ~9-10 Ma, even though age of the plutonic complex is 25-17 Ma (Coleman et al., 1997). A single fission track result indicates faulting between the Pavant Range and the Tushar Mountains on the southern line happened ~7 Ma, although some basin formation as early as 14 Ma is evident (Rowley et al., 1979). Stewart and Taylor (1996) believe motion on the Hurricane Fault south of the southern line occured by Late Miocene or Early Pliocene based on offsets of dated volcanics. Rowley et al. (1981) date offset rhyolites in Kingston Canyon to indicate Early Pliocene fault motion. In the Wasatch Plateau ~39.5°N, strikes of Early Miocene potassic dikes are ~N60°W, but dikes intruded 7-8 Ma strike N-S (Tingey et al., 1991). TZ deformation possibly started distinctly later than the main GB phase of extension, rather than developed gradually from west to east in response to GB extension. Further FT work here would be revealing, especially using the newer U-Th/He method sensitive to smaller uplifts (Ehlers et al., 2001).

The well-known Sevier Desert reflector (SDR) to the east of the Snake Range is widely - but not universally - interpreted as a west-dipping detachment fault (SDD) accommodating ~35 km of Cenozoic crustal extension (Allmendinger et al., 1983; von Tish et al., 1985; Planke and Smith, 1991; Otton, 1995; Allmendinger and Royse, 1995; Stockli and Linn, 1996; Coogan et al., 1995; Coogan and DeCelles, 1996; 1998). However, in an alternate viewpoint, this reflector simply represents a Tertiary-pre Mesozoic depositional contact along much or all of its length (Anders and Christie-Blick, 1994; Anders et al., 1995; Wills et al., 1996; Anders et al., 1995, 1998; Wills and Anders, 1999). We have attempted to represent these data and disparate interpretations in two, alternate reconstructions of the eastern GB and TZ near latitude 39.5°N, prior to extensional unroofing when regional relief was low (Figure 4). The upper section (A, A'), based mainly on Allmendinger (1992), Bartley and Wernicke (1984) and Lewis et al. (1999), assumes existence of the SDD and that eastern Snake Range rocks were buried originally to depths up to 30 km. The lower alternative (B, B') holds that the SDR is erosional (Anders and Christie-Blick, 1994; Wills and Anders, 1999) and that neo-Proterozoic strata of the Snake Range are interpreted to be near their stratigraphic depth in the Oligocene (Miller et al., 1983; 1999). Nevertheless, we give evidence that current eastern GB and TZ zone deformation may be largely unrelated to these SD and SR movements, regardless of the model, and reflects a different force/strength regime.

The style of Basin and Range faulting differs significantly between the calc-alkaline amagmatic (e.g., Delta latitude) versus magmatic (Milford latitude) regions. Amagmatic areas of the eastern Great Basin are characterized by relatively few large-displacement faults that define longitudinally continuous ranges and basins. By contrast, the E-W trending belts containing the majority of Cenozoic magmatic centers are characterized by a larger number of geometrically diverse, smaller-displacement faults. Aggregate displacement across such faults may be similar to that in amagmatic areas (Coleman et al., 1997), but more data and careful reconstructions of both regions are warranted. The more intricate fault pattern in the magmatic belts results in east–west belts of disorganized topography into which the well-defined basins and ranges of the amagmatic areas terminate. Any expression of the SDD or Snake Range decollement in the magmatic terrains is quite obscure (Smith and Bruhn, 1984; Miller et al., 1999). Farther south in Arizona, Falkner et al. (1995) and Axen (1998) also observe plutonic fields serving as rupture barriers to propagating faults.

Surface geology of the Transition Zone shows the same volcanic-avolcanic dichotomy as the GB proper. The relatively well-defined basins to the north (San Pete Valley, Sevier Valley) pass into more irregular morphology southward in the Marysvale field. However, overall degree of extension is quite minor (<10%) compared to the adjacent eastern Great Basin (e.g., Wernicke, 1985, 1992). Even the amagmatic basins are shallower and narrower, and basins and ranges are less continuous and well defined along strike. The TZ is located southward along strike of the active Wasatch fault, and modern seismicity, geodetic data, and young fault scarps (Lowry and Smith, 1995; Bennett et al., 1999; Thatcher et al., 1999; Wernicke et al., 2000) support active deformation in the TZ. Some faults in the magmatic terranes are whole-crustal and indicate fluid movement at such scales (Wannamaker et al., 1997a).

The first mafic lavas signalling the Basin and Range regime were highly potassic, with trace element signatures similar to island-arc basalts and low ε_{Nd} values (-5 to -7), indicating initial purging of low-melting point components from a source region mainly in ancient lithospheric mantle (Best et al., 1980; Kempton et al., 1991; Nelson and Tingey, 1997; DePaolo and Daley, 2000; Wannamaker et al., 2000) (Figure 5). Across the TZ, this magmatism concentrates in the NE-trending Pahranagut-San Rafael Belt becoming younger and more alkaline to the northeast (Figure 2). Quaternary basalts, in contrast, occur along a N-S belt at the eastern margin of the Basin and Range, from the western Grand Canyon to the Black Rock Desert in the north. The latter basalts have isotopic and geochemical characteristics similar to the Pliocene basalts of the Marysvale Volcanic Field, but further south the Quaternary basalts have more variable trace element characteristics, as well as higher ε_{Nd} values and lower $^{87}Sr/^{86}Sr$ ratios. These characteristics have been variably interpreted as reflecting upwelling asthenospheric (ocean island basalt - OIB?) mantle in magma production (Nealy et al., 1997; Nelson and Tingey, 1997; Nusbaum et al., 1997) or as evidence of chemical heterogeneities in the lithospheric mantle (Smith et al., 1999). The chemical and isotopic variations of S.W. Utah basalts in fact are similar to those in the southern Sierra (Figure 5), despite the intense Mesozoic magmatism that affected the mantle beneath the Sierra Nevada.

FUNDAMENTAL PROCESSES OF THE GREAT BASIN-COLORADO PLATEAU TRANSITION

Early extension, thick sedimentation, compressional deformation, magmatism and late extension all have seen eastward limits defined at least approximately by the present-day Great Basin to Colorado Plateau transition. Some of the ~100 km wide, GB-CP Transition Zone characteristics are intermediate between those of the two provinces, while others are uniquely anomalous. This long-lived contrast in histories, plus the struggle between the two provinces today for accommodation of their diverse states and processes, lead to our recognition of the three basic issues set out in the Introduction. While quite intriguing, previously reviewed results underscore the very coarse distribution of basic data upon which current models of the TZ are founded. We will elaborate upon this state, showing that diverse possibilities exist for processes defining the progression from active extension to stable interior across the Great Basin to Colorado Plateau transition.

Persistence of Stable Blocks like the Colorado Plateau

Deformation occurs when deviatoric stresses exceed a yield strength associated with a significant strain rate. Thus, deformation or its absence can be represented by two end-member possibilities for the Cenozoic: (1), force differences between the Colorado Plateau and Great Basin are responsible for the difference in deformation history; or (2), strength contrasts have caused the difference in deformation. In reality, we accept a role for both, although it is useful to consider individual contributions as much as possible to understand deformation processes at the TZ (Figure 6). The logical target point for reconstructing deformation would be Middle Oligocene time when structural relief was low and extension was soon to initiate (Best et al., 1989; Best and Christiansen, 1991).

Force Scenario

The present day variation of forces across the Great Basin-Colorado Plateau transition should have a large and perhaps dominant component of buoyancy forces caused by lateral variations in the density structure of the lithosphere (Figure 6a). In the absence of basal tractions, boundary forces on the sides of a section across this region must balance, which in turn makes the magnitude of boundary forces roughly uniform across the region. Estimates of the magnitude of buoyancy forces in the modern day western U.S. expressed as ΔPE, the difference in lithospheric gravitational potential energy relative to a reference column of asthenosphere, are shown in Figure 7a (from data in Jones et al., 1996). The potential energy, ΔPE, is the integral through the lithosphere of the vertical normal stress, or $\int \rho g z dz$, where ρ is density and g is gravitational acceleration. By estimating crustal densities from seismic velocity profiles and using isotasy to constrain the mantle mass surplus, this integral can be computed for the lithosphere. For a uniform rheology and vertically uniform deformation, ΔPE will be proportional to strain rate (Sonder and Jones, 1999). Somewhat surprisingly, if we plot the results of Jones et al. (1996) as ΔPE versus an estimate of overall modern day strain rate, we find a strong correlation that apparently requires little variation in the rheology of the lithosphere (Figure 7b). While we would not infer a total absence of strength variation from this, at face value it does indicate that this force hypothesis is plausible.

Unfortunately, the permissibility of absence of force in the Colorado Plateau and across the Transition Zone is largely the result of ignorance of density structure through the lithosphere of the two provinces. This was noted by Jones et al. (1996), and in particular we warn against assuming a clear difference in PE between the entral and eastern Great Basin (Figure 7a) (also see Flesch et al., 1999). The best source of deep density estimates is well-resolved seismic models, which are grossly lacking across the TZ. Figure 8 illustrates the problem. In short, there are no reversed seismic refraction profiles to provide intrinsic P-wave velocity estimates with depth and position in the region. Crustal thickness contours and velocity in this figure are based mainly on results from the early 1970's (reviewed by Smith et al., 1989; and Pakiser, 1989). Additional interpretation of two-station earthquake travel times (also unreversed) by Loeb and Pechman (1986) (cited in Smith et al., 1989, and Pakiser, 1989) lead to a picture of very thin crust (to 23 km) of $V_P < 7$ km/s under the little-extended TZ, in turn underlain by a substantial rift "pillow" of 7.4-7.5 km/s P velocity (Figure 8). This model was criticised by Pakiser (1989) who stated that individual station pairs yielded P_n estimates scattered from 7 to 10 km/s and attributed this to substantial Moho topography. Note that the crustal thickness under the GB-CP transition is unknown in Pakiser's model (Figure 8); it could in fact be even thicker than in the CP interior due to Sevier era thickening if subsequent extension is represented solely by surface faulting (cf. Wernicke, 1985). On the other hand, Lynch (1999), from travel-time tomography using local earthquakes, modeled such a high-V_p medium in the lower crust beneath the Great Salt Lake and Wasatch Front (latitude 41°N). Without remedying this enormous uncertainty in structure, the distribution of forces through the GB-CP transition cannot be resolved. Similar uncertainties exist for upper mantle structure.

In addition to density, modeling body force variations back through time requires paleoelevations to completely bracket the magnitude of the buoyancy forces involved (Jones et al, 1998). Paleobotany analysis suggests uplift to 3-4 km, higher even than today, was achieved in the eastern CP and eastern GB in Eocene time (Wolfe et al., 1997; Gregory-Wodzicki, 1997; Wolfe et al., 1998; Forest et al., 1999). The GB-CP Transition Zone actually represents a large hole in paleobotany studies in the western U.S. but its potential to reveal uplift or subsidence history appears promisng.

While basal traction forces in lithospheric deformation usually are obscure, there is intriguing evidence that they may have a role in deformation of the TZ (Figure 6b). The role would arise if a difference in lithospheric thickness does exist between the Great Basin and Colorado Plateau (irrespective of total strength differences). By Late Oligocene time, it was suggested that a shallow, enhanced asthenosphere may have existed under much of the present GB relative to the CP. To the extent an asthenosphere represents a zone of mechanical decoupling between the overlying plate and middle mantle levels below (Turcotte and Schubert, 1982; Karato and Wu, 1993), this decoupling may tend to occur at shallower levels under the GB than the CP. Motion of the North American plate carrying both provinces southwestward in a hot-spot frame (Figure 2)

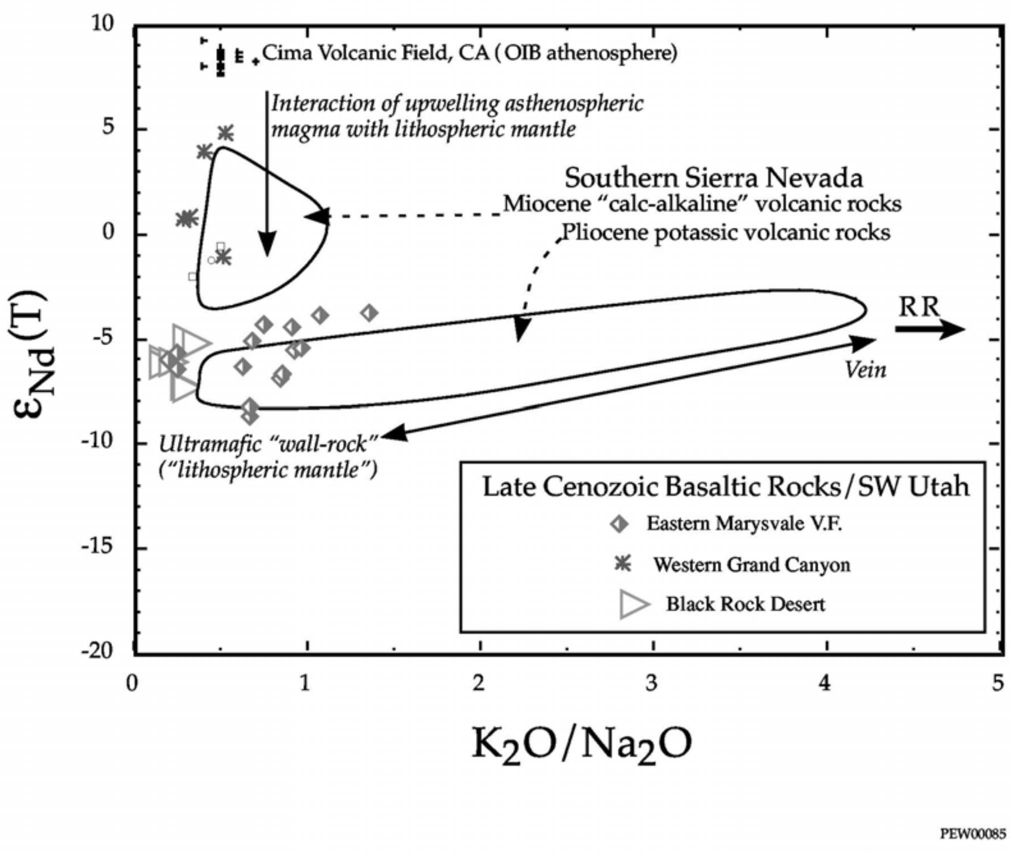

***Figure 5**. ε_{Nd} vs. K_2O/Na_2O for basaltic rocks from SW Utah (Mattox, 1997; Nelson and Tingey, 1997). RR denotes Robbers Roost lamproite (Wannamaker et al., 2000). Chemical and isotopic compositions of basaltic rocks interpreted to be lithospheric mantle derived (Pliocene basaltic rocks from the eastern Marysvale V.F. and Quaternary basalts from the Black Rock Desert) are similar to those of Pliocene volcanic rocks at the western margin of the Basin and Range in the southern Sierra Nevada (Farmer et al., in prep). Miocene basalts in the Sierra and Quaternary basalts from the western Grand Canyon area show a near vertical trend in this diagram and are interpreted to represent interaction between asthenosphere derived magmas and the same low ε_{Nd} peridotitic wall-rock involved in the Pliocene magmatism.*

thus may induce a certain "bottlenecking" of asthenospheric flow at the CP. Most of this conflict could be resolved by asthenospheric flow around the CP (out-of-plane motion for section views).

The shear wave (SKS) splitting measurements for the southwestern U.S. by Sheehan et al. (1997) and Savage and Sheehan (2000) could be reflecting this (Figure 9). Depth of origin of observed anisotropy is hard to pinpoint, but much must come from depths >100 km. Anisotropy exhibits pronounced, N-S fast axes in the eastern GB and the Transition Zone while null measurements are typical of the more quiescent central GB and north-central CP. Gao et al. (1997, 1999) hold that a rift-normal fast direction corresponds to mantle crystallographic alignment with spreading, while a rift-parallel fast direction could represent a broad zone of melt dike intrusion. Vauchez et al. (1999, 2000) challenged this, arguing that rift-parallel fast directions indicate solid-state flow parallel to the rift axis (i.e. N-S along the TZ as in Figure 9), and that melt pockets tend to align parallel to flow and may augment anisotropy. Vauchez et al. (2000) call upon transtensional rifting to create this parallel flow. However, in our study area, modern geodetic data show rift-normal motion to occur at the surface (Thatcher et al., 1999; Wernicke et al., 2000). SKS patterns hence suggest possible deep N-S flow of eastern GB mantle in response to CP keel impact (see Fouch et al., 2000). Deflected asthenospheric flow at the GB-CP interface could impart a dynamic stress which may be indicated in the rather high effective elastic thickness (T_e) estimates for the Transition Zone (discussed shortly). Such force should also tend to reduce extensional stress within the CP, and may contribute to a favorable stress regime for low-angle detachment under the CP (lithospheric scale application of Westaway, 1999). However, resolution of splitting is only comparable to station spacing (~100 km), making it difficult to assign fast directions to zones of mantle flow or melting.

Plate boundary forces influencing Miocene to present deformation in the eastern Great Basin usually have been discounted due to difficulty in propagating stresses from Pacific-North American plate interaction this far eastward assuming likely rheology (Sonder and Jones, 1999). However, temporal variations in basaltic magmatism in the eastern Great Basin suggest some influence of plate motion. Best et al. (1980) describe three temporal peaks in eastern GB basaltic volcanism at 14-11 Ma, 9-5 Ma and 2-0 Ma. Correspondingly, recent geologic and plate motion reconstructions confirm a westerly motion of the Sierran-Great Valley block relative to the Colorado

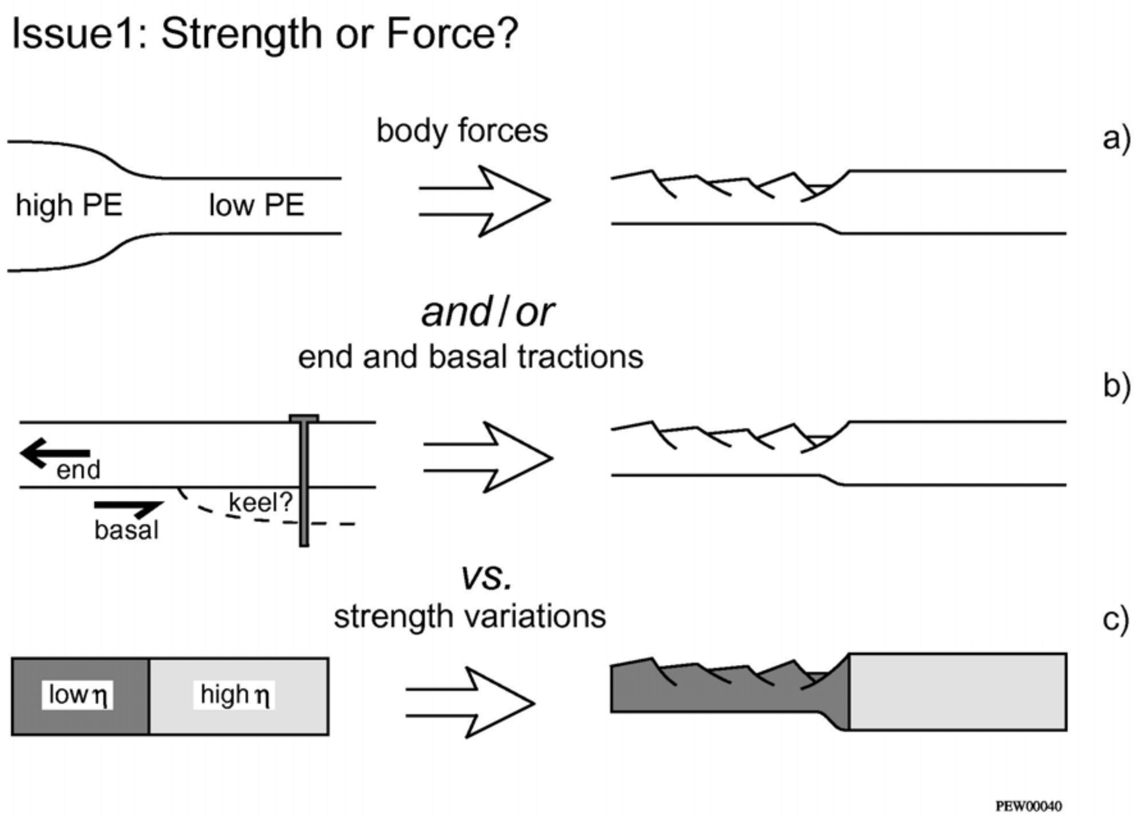

Figure 6. *Conceptual depiction of principal factors affecting deformation in the Great Basin-Colorado Plateau region. Gravitational potential energy (GPE) differences are due mainly to density profile and elevations variations. Distant plate boundary forces are believed to impose a fairly uniform stress gradient across the study area. Basal traction forces may be small for purely 1-D geometries, but perhaps more substantial with lithospheric basal "topography". Strength (viscosity) denoted by* η.

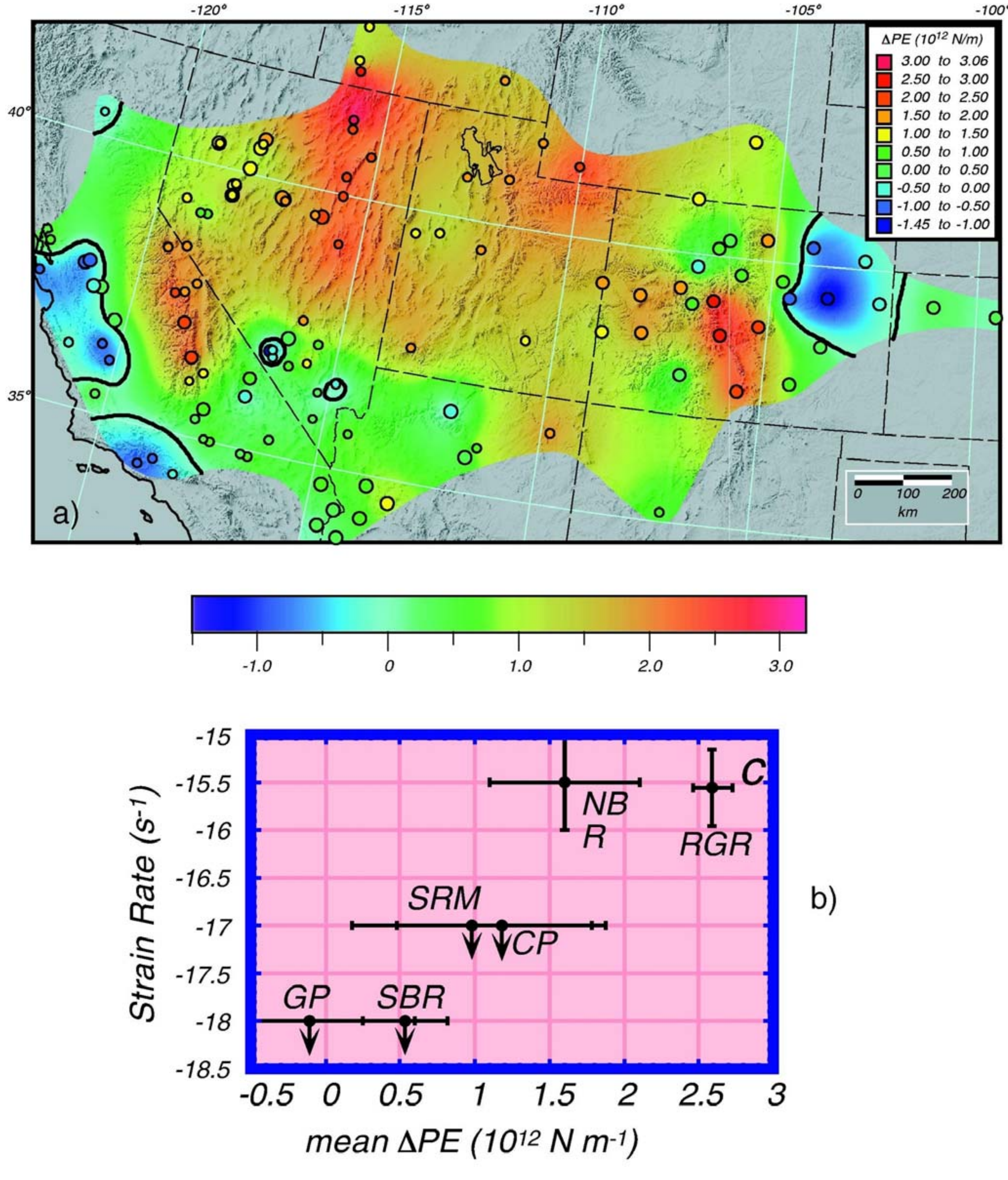

Figure 7. *Top (a): Gravitational potential energy (ΔPE) distribution in the western U.S. (from data in Jones et al., 1996). Open circles indicate locations of seismic profiles used to constrain crustal density columns; dot sizes reflect relative confidence in techniques used in the seismic interpretations. Note sparsity and lower confidence in Utah coverage. (b): Plot of ΔPE versus strain rate for Great Plains (GP), Southern Basin and Range (SBR), Southern Rocky Mountains (SRM), Colorado Plateau (CP), Northern Basin and Range (or Great Basin, NBR), and Rio Grande Rift (RGR).*

Plateau in the Miocene, but with an increase in rate ~12 Ma, and a change to a NW-NNW direction since ~8 Ma (Wernicke and Snow, 1998; Atwater and Stock, 1998). Directions of spreading and least principal stress varied from ENE-WSW to ESE-WNW after about ~8 Ma (Wernicke, 1992; Zoback et al., 1994; Wernicke and Snow, 1998). The earlier two basaltic pulses coincide with the aforesaid plate reorganizations and changes in deformation, while the most recent correlates to increased spreading rate of the Gulf of California (DeMets and Dixon, 1999; Dixon et al., 2000). Alternatively, changes in deformation correlated with plate motions may represent changes in accommodation space or in basal traction forces.

Strength Scenario

As with the force scenario, we consider the idea that strength is the controlling factor in deformation from the Colorado Plateau to the Great Basin (Figure 6c). The intrinsic strength of the lithosphere is a function of its rheology, which in turn is controlled by a number of factors including rock composition, temperature, grain size, and fluid content (Poirier, 1985; Tullis, 1990; Tullis and Yund, 1991; Kohlstedt et al., 1995; Tullis et al., 1996). Regarding composition, Farmer and DePaolo (1983, 1984) observe a discontinuity in the $^{87}Sr/^{86}Sr$ (at fixed ε_{Nd}) of Cenozoic granitic rocks between the NV-UT border and the Wasatch Front. They interpreted this to suggest that low Rb/Sr felsic lower crust underlies the Wasatch, but thinned (i.e., rifted) Precambrian crust without low Rb lower crust occurs to the west in the Great Basin. Based on seismic velocities (near Arizona border) and pluton petrogenetic modeling, Nelson and Tingey (1997) correlated thicker mafic lower crust of the CP with its unextended nature. Lee et al. (2001) infer from mantle xenoliths that Basin and Range extended crust generally overlies Mojavia province mantle lithosphere, which is less refractory and more deformable than Yavapai lithosphere such as under the CP. This underscores the uncertain state of the Mohavia-Yavapai boundary in the eastern GB, where highly extended lithosphere contains at least some Yavapai-aged rocks.

A standard model for GB-CP structural differentiation is collapse of the thickened pre-existing crust (Wernicke, 1992; Burchfiel et al., 1992), but this works less well for extension in the TZ itself. Although Paleozoic margin isopachs and Sevier thrust sheet toes approximately parallel the TZ, in detail they swing abruptly westward from central west to southwest Utah (Stewart, 1980; Allmendinger, 1992; Figure 2). This is mimicked by the inferred Cretaceous foredeep (Royse, 1993). In the Transition Zone in central Utah (~latitude 39.5°N), Phanerozoic sediments including the foredeep can reach ~8 km thickness. Just south of the volcanic covered Pioche-Marysvale field

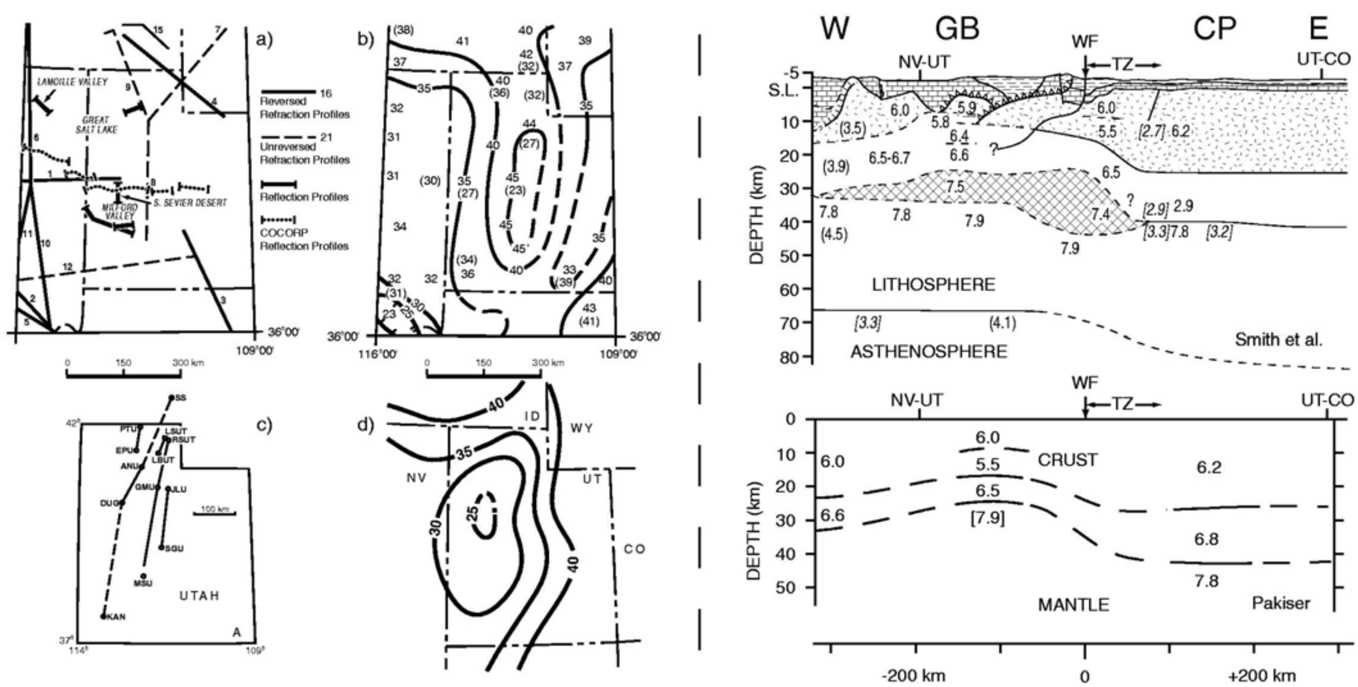

__Figure 8.__ Left: coverage of seismic profiling near Great Basin-Colorado Plateau transition, after Smith et al. (1989) and Pakiser (1989): a), active source, reflection and refraction profiles in western Utah and eastern Nevada; b), crustal thickness contours in km fgrom Smith et al. considering concept of "double-Moho" with thick, high V_p rift pillow lying between 25 and 45 km depth beneath Transition Zone; c), earthquake station pairs considered by Pakiser to demonstrate Moho roughness in V_p estimates; d), alternate crustal thickness contours in km from Pakiser. Right: alternate crustal cross sections through west-central Utah redrawn from Smith et al. (1989) (top) and Pakiser (1989) (bottom). P-wave velocities (km/s) are in bold, S-wave velocities in parentheses, and densities (gm/cc) are in italics.

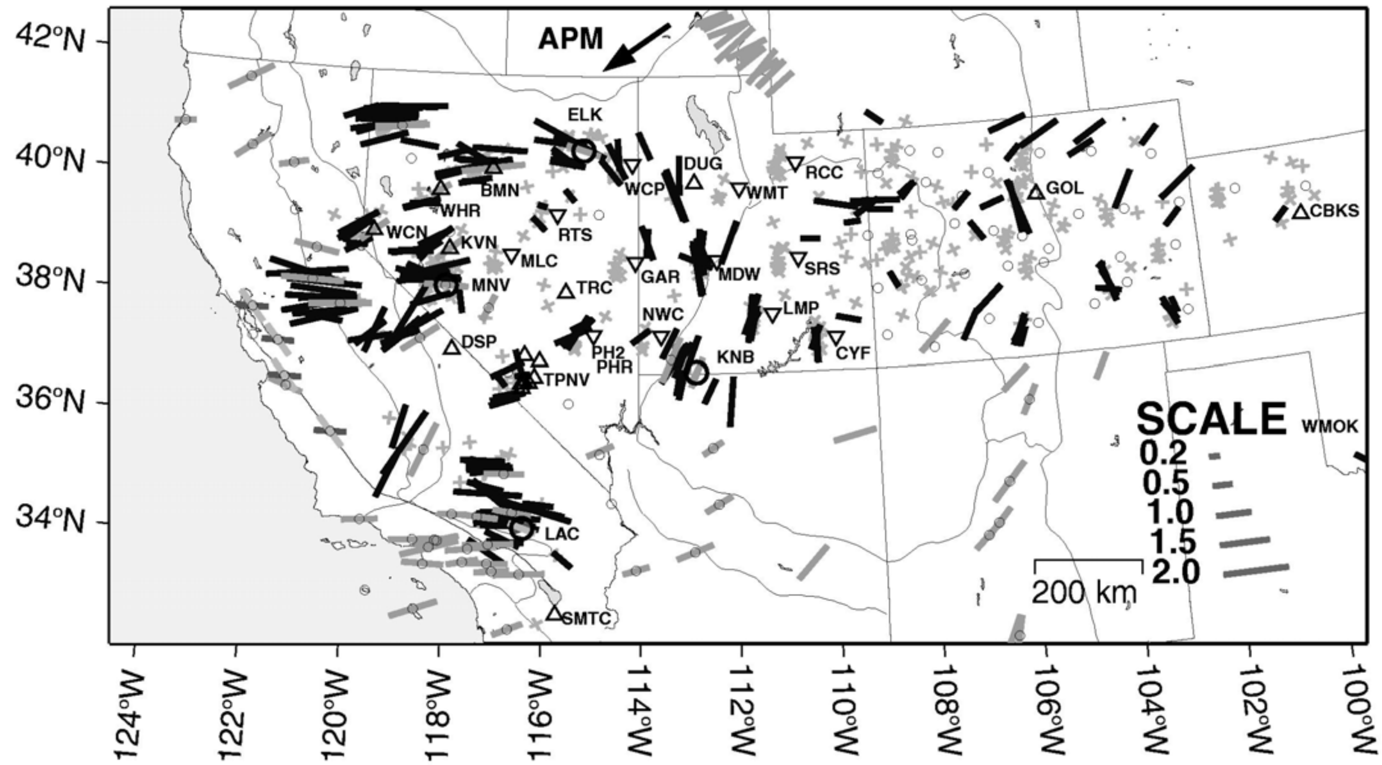

Figure 9. Compilation of shear-wave splitting measurements in the S.W. U.S (modified from Sheehan et al., 1997; Savage and Sheehan, 2000). Each measurement is represented by one or two lines. A positive measurement is denoted by a single black or grey line oriented parallel to fast direction of anisotropy, with length proportional to the amount of splitting in seconds. Null data are plotted with two small grey lines with directions equal to the two allowed directions of anisotropy. Measurements are plotted at the 220 km depth projection of the split ray, except for positive measurements shown as grey lines which are station averages from other experiments. Station locations are indicated by large open circles (CPGB experiment) or small open circles (all other stations). Direction and rate of North American absolute plate motion (APM) in hot-spot frame shown with large arrow.

(~latitude 38°N), these sediments in the TZ are only 2-3 km thick (Hintze, 1980, 1988).

The lesser age of the subducted Farallon plate under the Great Basin (Humphreys, 1995), may have provided a hotter and thicker asthenosphere there relative to the CP. However, Mid-Cenozoic heating of the GB by plutonism was highly heterogeneous, and long-standing magmatic gaps like the Mid-Utah should have left the lithosphere cooler and stronger (cf. Perry et al., 1993). Nelson and Davidson (1993) include the Marysvale field of the southerly GB-CP Transition Zone in little-extended or competent, Colorado Plateau domain crust of that time. As we describe further, this seems unlike today's state of the TZ crust or mantle.

Heat flow of the TZ (Figure 10) remains high and characteristic of the GB (~90 mWm^{-2}) to a distance of 80-100 km east of the physiographic province boundary for both transects. This compares to an average ~60 mWm^{-2} for the CP interior, with subregions as low as 51 mWm^{-2} (Bodell and Chapman, 1982; Powell, 1997). An enhanced deep thermal state of the transition is interpreted from Curie Depth models also (Shuey et al., 1973; 1977). Seismic focal depth estimates in the TZ (Figure 10) suffer from coarse and narrow station coverage (Lowry and Smith, 1995), but hypocenters extend to ~20 km depth. This is consistent with an elevated geotherm of 30°C/km and a plagioclase crystalline rheology (Tullis, 1990) appropriate to the likely Precambrian subcrop (Karlstrom and Williams, 1998). Although these depths are hard to relate to the eastern GB because of its extremely scarce seismicity (Smith et al., 1989), those of the CP interior can extend through the entire crust (Wong and Humphries, 1989; Wong and Chapman, 1990).

A direct view of variations in strength between provinces is provided by gravity/topography coherence analysis. Isostatic analysis indicates a significant change in the effective elastic thickness (T_e) from <10 km in the GB to >20 km in the CP (Lowry and Smith, 1995; Lowry et al., 2000; Figure 11a). T_e estimates in the Colorado Plateau and Transition Zone likely reflect a modern (even if transient) rather than a fossil state because the TZ and CP are, if anything, undergoing at least some warming. T_e of the eastern GB is small, although perhaps underestimated slightly because of possible cooling since Early Miocene time. Curiously though, T_e estimates from both coherence analysis (Lowry and Smith, 1995) and modeling of Wasatch footwall uplift (Zandt and Owens, 1980) are relatively large in the GB-CP transition itself, at odds with heat flow, conductivity (presented later), seismicity, and Curie Depth. However, T_e estimates integrate all vertical shear stresses that contribute to isostatic support, so that anomalously high T_e in the TZ may simply indicate presence of dynamic flow stress, e.g., the lateral flow that might be induced by deflection of plate shear flow in the asthenosphere in advance of a CP keel. Resolving the potential contribution of dynamic flow to T_e requires high quality seismic and thermal regime estimates.

Rheological parameterization of T_e (Figure 11b; Lowry et al., 2000) indicates that some kind of material heterogeneity (i.e., composition, fluid content, or grain size) enhances strength of the CP lower crust and/or mantle relative to the GB. They suggest that tectonism focuses where lithosphere with negligible mantle viscosity abuts lithosphere with significant uppermost mantle viscosity, but exactly where in the TZ this occurs is difficult to define. Anisotropy also is inferred in the study area, with lower T_e in the east-west direction on the Great Basin side of the transition zone, rotating to a north-south axis of lower T_e in the Colorado Plateau (Lowry and Smith, 1995). T_e by itself does not suffice to distinguish whether the crust or the mantle contributes more to the strength difference, and the needed seismic and thermal estimates are limited. Early results of Julian (1970) show a CP low-velocity zone (LVZ) at 125 km depth versus results from Archambeau et al. (1969) with an LVZ at only 80 km depth. Estimates for the CP by Beghoul et al. (1993), Zandt et al. (1995) and Lastowka et al. (2001) are in the range 62-90 km depth but tend to emphasize the western portion. Teleseismic velocity models of the Great Basin (Parsons, 1995; Zandt et al., 1995; Lastowka et al., 2001) show a depth to LVZ of 55-70 km, only moderately less than CP estimates. However, the complex orientation of SKS splits over most of the western U.S. lead Schutt and Humphreys (2001) to suggest that North American motion at depth is taken up by strain at depths greater than typically associated with the asthenosphere.

Sheehan et al. (1997), Libarkin and Sussman (2000), and Lastowka et al. (2001) indicate that some sort of mantle support is required for CP elevations. Lowry et al. (2000), McQuarrie and Chase (2000), and Chase (2000), in contrast, argue for important if not dominant contributions from the crust. The hydrous metasomatism suffered by the CP lithosphere during prior subduction may greatly reduce its solid-state viscosity relative to refractory peridotite at equal temperatures (Karato and Jung, 1998; Boutelier and Keen, 1999). Despite higher temperatures, Cenozoic magmatism of the Great Basin has purged the mantle lithosphere of hydrous phases to some degree, so some relative strengthening of the GB mantle lid seems possible. Humphreys and Dueker (1994) call on a more refractory upper mantle under the Great Basin as at least some of the mechanism for its high elevation. These uncertainties are compounded by the ambiguous relationship between velocity and strength, so that relative differences may be all that are reliable.

Extensional Consumption of Stable Lithosphere

While we have previously examined the causes of deformation, or its lack, our second topical issue considers how the transition from active extension to stable interior can proceed (Figure 12). Perhaps the simplest model of this is uniform gravitational collapse of thickened crust as it is warmed by recent, adjoining rifting (Hopper and Buck, 1998) (Figure 12a). However, we give evidence that the extension encroaching today at the GB-CP transition is highly non-uniform with depth, as has been inferred at other exhumed rift orogens (Brun and Beslier, 1996; Froitzheim and Rubatto, 1998). Specifically, geophysics suggests deep extension is most vigorous under the TZ itself, whereas surface extension is most evident in the eastern GB and western TZ. Axen et al. (1998) review evidence from field areas worldwide showing not only large degrees of lower crustal deformation and lateral movement are possible, but that this deformation could be cryptic in terms of contemporary, surface geological activity. Fortunately, from indicators reviewed previously and additional evidence discussed below, the GB-CP transition in much of Utah shows a clear strike, so we can interpret much in terms of cross-sectional geometries.

The Case for Non-Uniform Extension Under the GB-CP Transition

The Transition Zone as we have described it ressembles the "discrepant zone" of Wernicke (1985) in this region. The offset regimes of upper and lower crustal extension suggested by seismic crustal thickness models (Figure 1) gave rise to a famous model of whole-lithosphere, simple normal shear (Figure 12c). The concept has eluded clear field verification, since for example the associated rolling hinges are difficult to identify (Axen and Bartley, 1997). It also predates the wide consideration now given to lateral flow in a "fluidized" lower crust, which was called on more recently to remove deep crust from under the Colorado Plateau in Arizona (Wernicke, 1992) (Figure 12b). Nevertheless, regardless of mechanism, an offset between upper and lower crustal extension remains a prescription for some form of simple shear detachment (perhaps diffuse) to accommodate differential extension (e.g., Hopper and Buck, 1998). Tectonic models for thinning through the TZ remain controversial because both geological and geophysical data are so limited in sampling and reliability.

Govers and Wortel (1993) showed numerically that a lithospheric scale normal detachment dipping beneath a region of thickened crust might form shortly after relaxation of compressive stress. Although their model originally was pertinent for collapse of compressional belts, its sense perhaps can be reversed at the thickened GB-CP transition for Late Cenozoic activity. In their review, Harry and Bowling (1998) indicate that whole-lithosphere, simple normal shear deformation can be produced in rheologically heterogenous lithosphere. However, for a layered rheology with small random perturbations in layer thickness and isotherm depth, lithospheric-scale dipping normal detachments appear unlikely to develop (Govers and Wortel, 1995), suggesting that strength heterogeneity is essential for non-uniform extension to take place.

Most intriguing in our study area, presently known geologic relations (e.g., Standlee, 1982; Willis, 1986; Witkind, 1981) suggest that the east-dipping normal fault that defines the western boundary of the Wasatch Plateau may be the master fault of an extensional shear system which roots under the Colorado Plateau and which formed the basins and

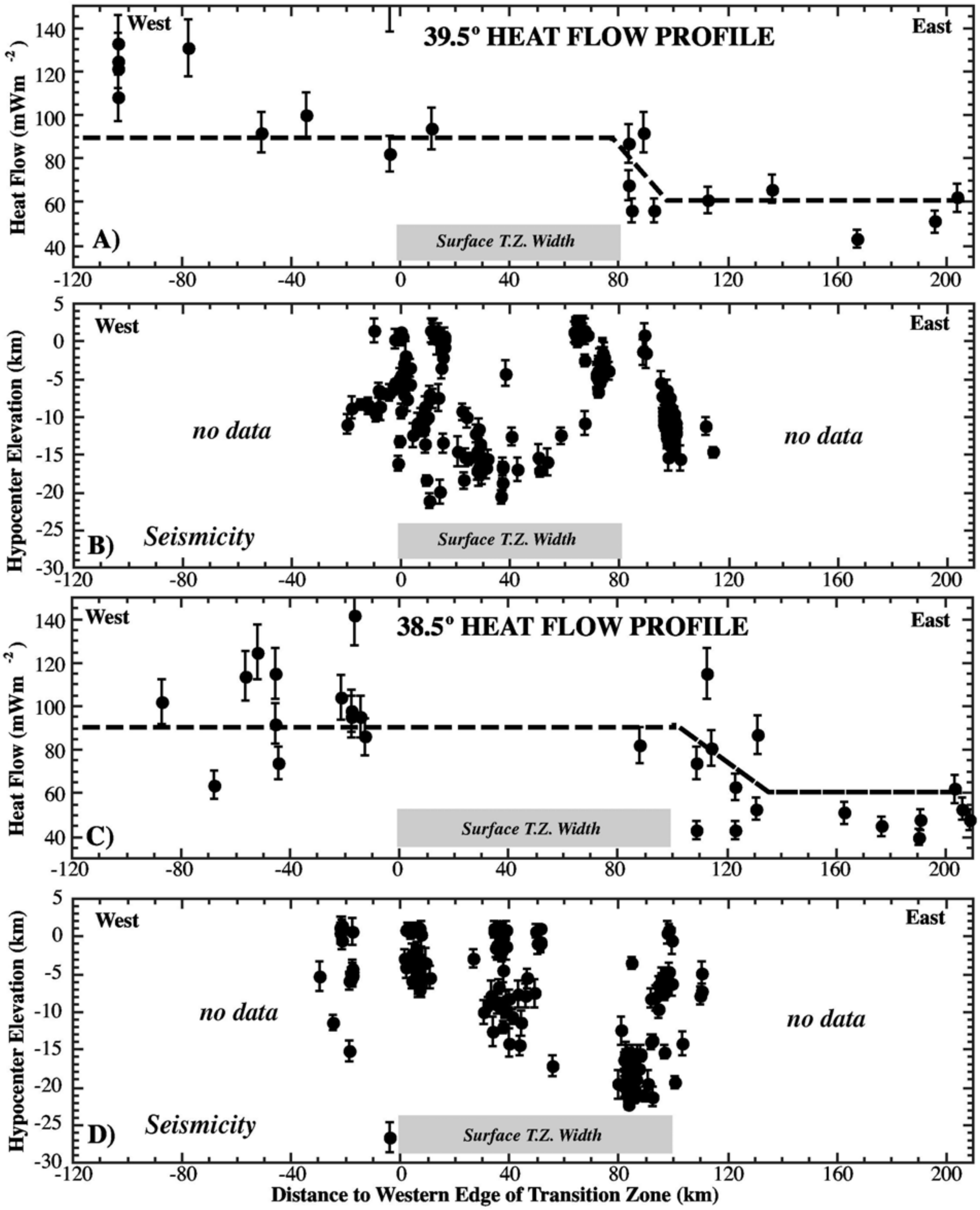

Figure 10. *Heat flow and seismicity within 0.5° latitude for two E-W transects, 39.5°N (northern, A and B) and 38.5°N (southern, C and D). Heat flow data (Pollack et al., 1993) are shown with 10% uncertainties. Heat flow is high (~90 mWm-2) in the G.B. and 80 km (A) to ~100 km (C) east of the transition zone into the CP. Heat flow is lower (~60 mWm-2) in the CP interior. Typical GB and CP heat flow is indicated with the dashed line. Seismicity along the northern (B) and southern (D) transects was compiled from the U. Utah Seismograph Station network (1962-present). Selection criteria for hypocenters includes: a seismograph station located within 1 focal depth or 5 km, maximum error in vertical location of 2 km, maximum azimuthal gap of 180 degrees, RMS error less than 0.5 s. Seismicity along the northern transect (B) is typically ~5 km deeper beneath the surface expression of the Transition Zone (TZ) and 7+ km shallower within the CP interior, relative to the southern transect. Nearly 90 new heat-flow determinations are being worked up to help define the Transition Zone thermal regime (Henrikson, 2000).*

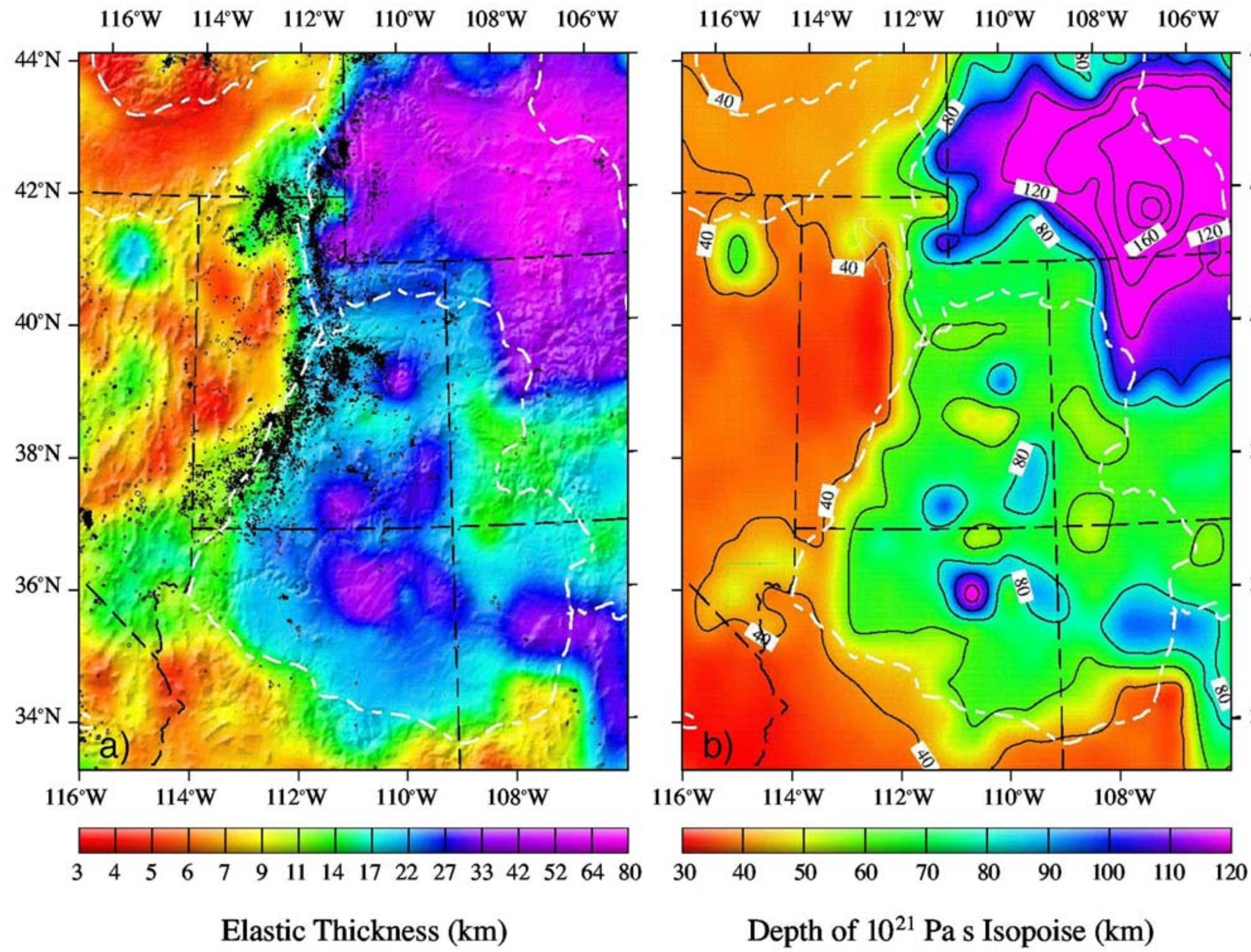

Figure 11. *Elastic thickness and lithospheric rheology, modified from Lowry et al. (2000). a, Elastic thickness of eastern Great Basin and Colorado Plateau illuminated by topographic relief. Background seismicity focuses at lateral gradients in lithospheric strength; b), Depth to the 10^{-21} Pa-s isopoise of viscosity, assuming granitic crustal yield strength using surface heat flow and elastic thickness. Effective viscosity is more sensitive to elastic thickness than to geotherm or other parameters assumed in the yield-strength envelope.*

ranges of the central Utah Transition Zone (Bartley and Walker, 1999). Moreover, recent unpublished mapping by USGS personnel (R. E. Anderson, pers. comm. to Bartley, 1997) reveals a highly attenuated, middle Tertiary stratigraphic section overlying Jurassic strata along a tectonic contact in the Sevier Valley area (Figure 1) (also see Anderson and Barnhard, 1987; Diehl et al., 1997). This contact descends eastward stratigraphically relative to an upper plate which is progressively less disrupted eastward. The structural geometry is consistent with an E-dipping detachment under the High Plateaus, and may be just the uppermost member of a family of such. If this evidence is representative, then deformation of the TZ is not directly linked to mapped structures in the adjacent Great Basin (e.g., the Sevier Desert detachment).

There is further evidence that the SDD may be little-involved in deep extensional processes today. Coogan and DeCelles (1996) maintain that normal faulting in the upper plate of the SDD soles into the detachment with no deeper expression in the footwall. However, their Sevier sub-basin (western and central components combined) can be followed southward nearly 100 km based on steep gravity gradients over the basin bounding faults (Zoback and Anderson, 1983; Figure 13). Zoback and Anderson observe that the Quaternary basalts of the Sevier desert lie along one or the other of these bounding faults. This localized tectonism must represent a major break in the detachment and indicates that lower crustal extension can take place which is largely unrelated to expression of the SDD (op. cit.). This narrow recent magmatism resembles that of the northern Ethiopian Rift, where initial lower-angle detachments also have been abandoned in favor of magmatic segmentation as the system approaches complete continental breakup (Ebinger and Casey, 2001). Nevertheless, Harry and Bowling (1998) conclude for lithospheric-scale simple shear rifting that associated lateral pressure gradients may deflect ascending magma, so that surface extrusion can be significantly offset from mantle source regions. Hence, the deep-seated structure feeding the SD magmatism may be a third candidate for a regional, shear detachment. This possibility, following Harry and Sawyer (1992), is intriguing because it suggests that lavas of the Sevier Desert region could be linked to extensional magmatic processes under the TZ. However, inclined magma pathways are not favored by Nelson and Tingey (1997).

A fourth possibility for surface manifestation of deep, simple shear processes is the Snake Range decollement (SRD) (Figure 4). This and related structures to the north were considered in the classic model of simple shear by Wernicke (1985) and would suggest such a deformational style could have started in the Middle Miocene. Establishing connections between shallow and deep processes is crucial for restoring structure to pre-extensional Oligocene time and bracketing force and strength distributions in history. Unfortunately, the COCORP reflection data through the Wasatch High Plateaus near 39.5°N latitude did not resolve coherent reflectors in the deep crust which might denote such detachments (Allmendinger et al., 1986). Based on experience in the southern Sierra Nevada and New Zealand Southern Alps (Jones and Phinney, 1998; Stern et al., 2000), P-S passive seismic conversions and active source reflectivity could be help define which, if any, of the prior candidate regions may connect with deep, active processes to the east. Large differences in total extension characterize the two sections of Figure 4 (~90 km in upper versus ~50 km in lower), as well as in implied amounts of lower crustal inflow/ additions to the eastern (upper sections) versus the western (lower sections) portions of the transect. Expression of both SDD and SRD in the magmatic terrains further south is unclear at present (Coleman et al., 1997; Miller et al., 1999).

The high heat flow and shallow Curie Depth associated with the Transition Zone are further evidence of enhanced extension at depth. That high heat flow may exist over the amagmatic region of central-west Utah as well indicates that it is not just associated with the volcanic belt of the southern TZ (Figures 10a and c). For uniform stretching, the degree of surface extension is far from sufficient to generate Great Basin-like heat flow either with instantaneous (McKenzie, 1978) or diachronous (Lachenbruch and Sass, 1978) extensional models. Within current heat flow sampling, there does not appear to be a steady decline in heat flow eastward from the physiographic boundary, as should be the case for conductive diffusion of heat from the Great Basin. We note, however, that a step-like, eastern thermal edge also is not predicted by a mere E-dipping simple shear under the TZ, surfacing near to the west of the main physiographic transition (cf., Buck et al., 1988; Ruppel, 1995). Presumably, additional convective processes such as ductile crustal thinning, magmatism or high-T fluids must contribute beneath the GB-CP transition. Crustal thickness estimates in Figure 2 suggest thinning under the transition which is vastly greater than implied by surface faulting in relation to original thickness assumptions. Nearly 90 new heat flow values for the TZ are available now from bottom-hole temperature analysis (Henrikson, 2000), and these should focus considerably the view of TZ thermal state.

We note that thermal modeling and geochronology in the TZ should incorporate effects of hydrology, dynamic topography and mass wasting on heat flow (Stuwe et al., 1994; van der Beek et al., 1995; Ketcham, 1996; Ehlers et al., 2001). At a scale critical to the interpretation of thermochronologic data, the thermal regime surrounding an active normal fault is complicated by "hot" material uplifted on the footwall, and "cold" material advected into the subsurface with the hanging wall (Figure 14). Above a threshold displacement rate, isotherms are curved and depth to closure temperature can vary up to 40% with similar consequences for exhumation rate estimates. Hence, interpretation of fission track data in tectonically active regions necessitates temporal, 2-D or 3-D thermal modeling, especially for low-T thermochronometers (U-Th/ He).

Mantle geophysical models also suffer from sparse sampling but still reveal some very interesting features. Tomographic images by Humphreys and Dueker (1994) under the Transition Zone relied predominantly on the Intermountain Seismic Network (ISN) stations, which lie in a narrow, N-S swath. These results compiled with other regional velocity estimates are presented in Figure 15 (K. Dueker, pers. comm.; Savage and Sheehan, 2000). The lowest upper mantle velocities, plus pronounced P and S delays, appear not under the eastern Great Basin *per se*, but at least overlap into the TZ. Again, lateral resolution is only ~100 km in the tomography. However, Rayleigh wave inversions by Lastowka et al. (2001) also show upper mantle S velocity is lower under the TZ than either the eastern GB or western CP.

Magnetotelluric Geophysical Evidence for Non-Uniform Extension

Deep electrical conductivity, a further indicator of thermal and magmatic processes at depth (Wannamaker and Hohmann, 1991), adds compelling evidence for mantle involvement under the TZ. The early conductivity models of Porath et al. (1970) and Porath (1971) indicated that the strongest, N-S trending conductive axis also lies under the southern continuation of the Wasatch Front, and not the most extended Sevier Desert (Figure 16). The method of Porath et al. which utilized only the magnetic field time variations had fair lateral resolution, but almost no depth resolution because it did not include the electric field as in magnetotellurics (Vozoff, 1991). This resulted in alternate models explaining the data by shallow asthenospheric structure, or by upwelling from the 410 km mantle phase transition (Figure 16c,d). The true structure likely varies through both levels. The geomagnetic anomalies implying this axis, and another beneath slightly rifted western Colorado, strike clearly N-S and strengthen northward (Figure 16a, b; Gough, 1989). They do not trend NE-SW (cf. Humphries and Dueker, 1994) and do not appear affected by possible NE magmatic alignments such as the interpreted Pahranagat-San Rafael trend (Nelson and Tingey, 1997). They instead parallel the Quaternary mafic eruptions of the Black Rock-Grand Canyon trend (op. cit.) and the general strike of the TZ. They continue only weakly into NW Arizona and stop short of the southern Basin and Range.

Independent evidence from magnetotelluric (MT) profiling supports the concept of significant offset of upper crustal and upper mantle extension across the TZ (locations in Figure 2). Lower crustal electrical conductivity under the active eastern GB appears high, much more so than that of the relatively quiescent central GB of eastern Nevada (Wannamaker et al., 1997b; Figure 17a). The eastern GB structure is compatible with the average extension rate over the past 10 Ma being accommodated mainly by basaltic underplating. This in turn should release sufficient hypersaline brines to maintain a porosity with lower bound near 0.3-0.4 vol. % in the face of upward fluid egress (e.g., Frost and Bucher, 1994; Wannamaker, 2000). Although range exhumation occurred primarily in the Middle

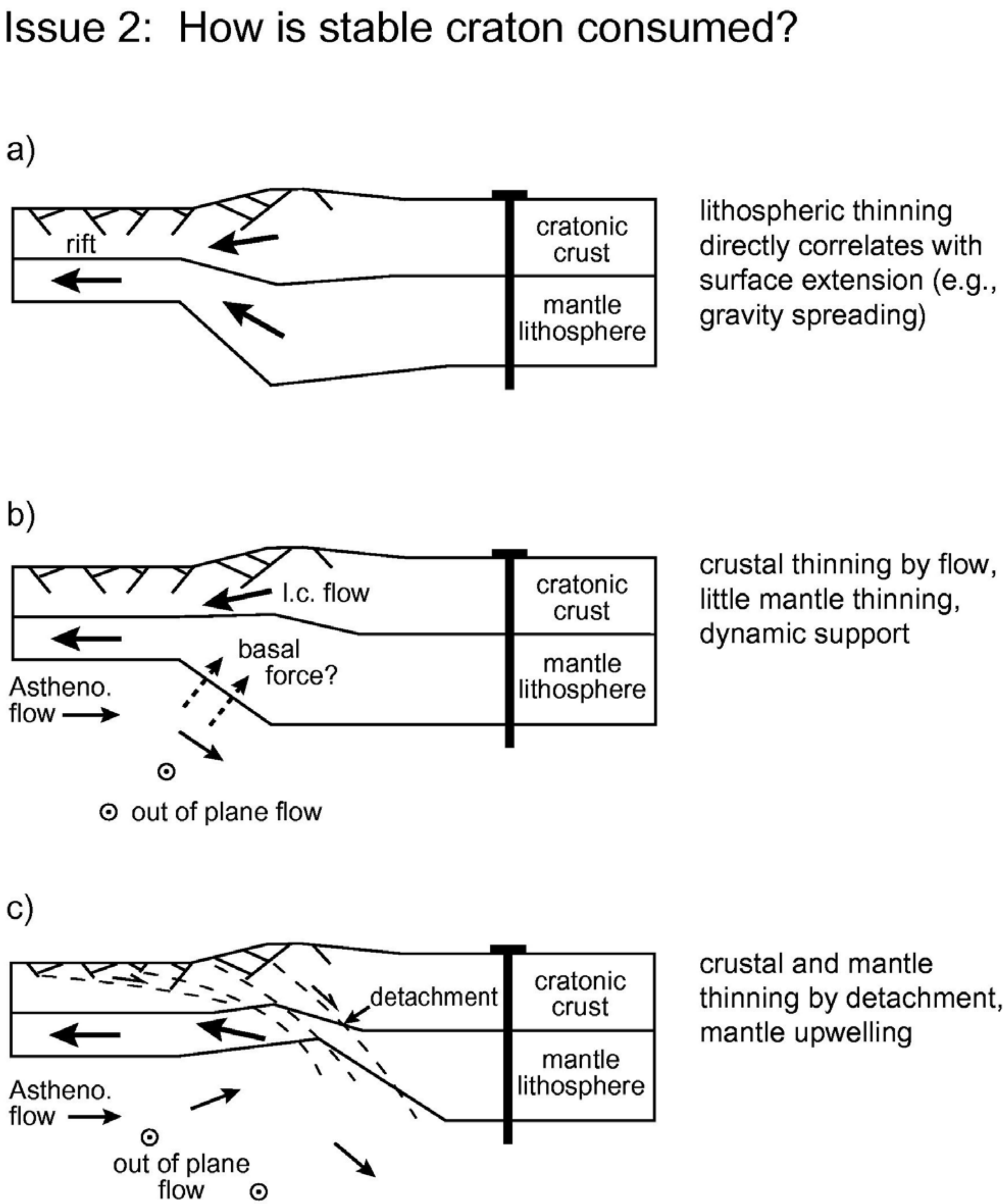

Figure 12. Possible manifestations of extensional consumption of formerly stable craton: a), Extension is fairly uniform with depth, occuring primarily by gravitational collapse of lithosphere which is thick compared to nearby extended regime; b), extension is in response to flow of lower crust towards extended terrane. There is relatively little mantle involvement, but mantle support of high elevation is required. Part of GB mantle deflection due to CP lithospheric keel impact may be out-of-plane; c), Upper crustal extension is a symptom of whole-lithosphere, simple-normal shear involving both lower crust and upper mantle, but with laterally offset domains of activity versus depth.

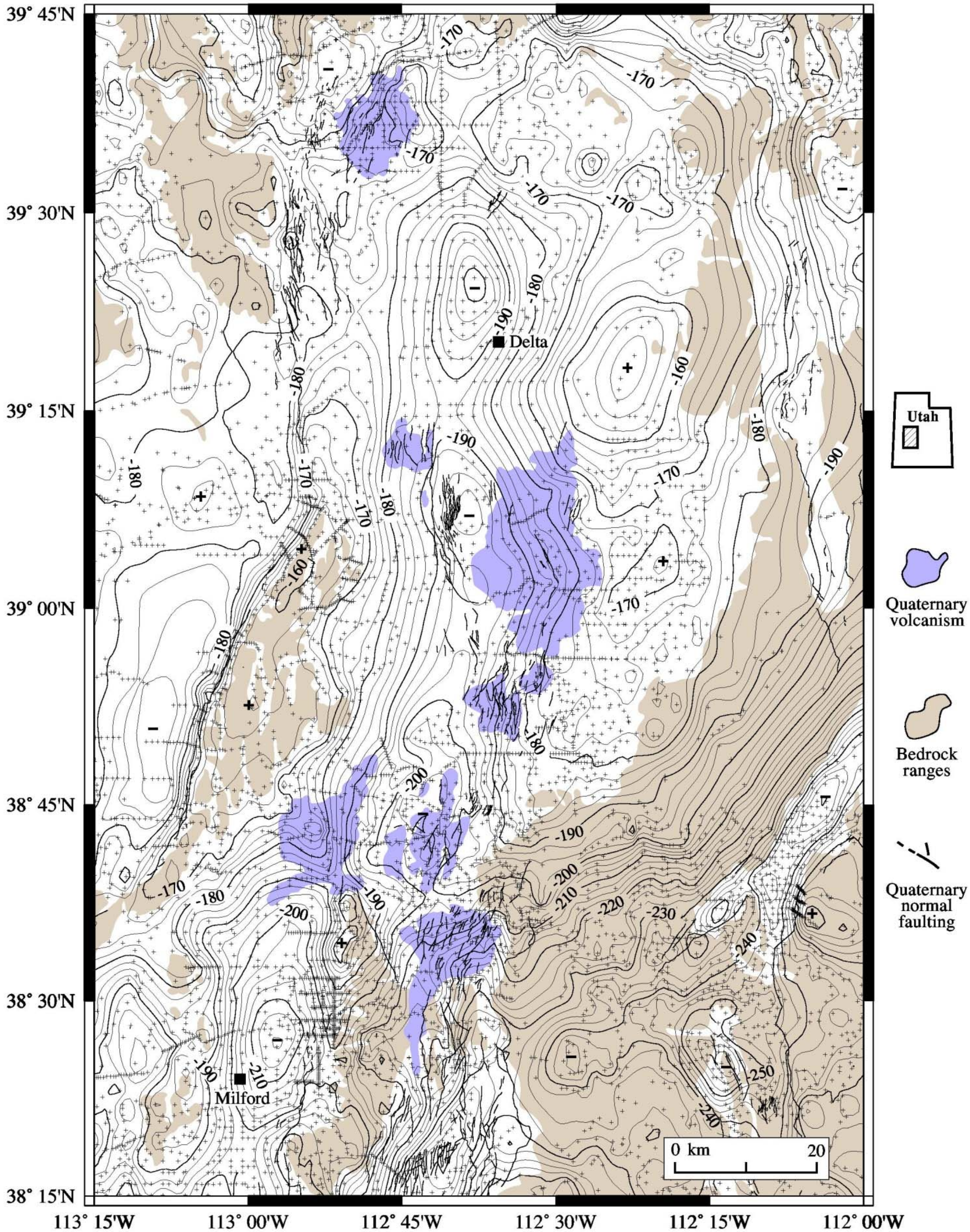

Figure 13. Complete Bouguer gravity map of the Sevier Desert-Black Rock Desert region of west-central Utah. Units in mGal. The pronounced north-south gravity low near the center of the map corresponds to the west-central Sevier sub-basin of Coogan and DeCelles (1996), and gradients to the sides define the graben-bounding fault zones. Note control of sites of Quaternary basaltic volcanism (purple) by graben bounding faults. Gravity data from U.S.G.S files and Quaternary faulting from Hecker (1993). Modeled after Zoback and Anderson(1983).

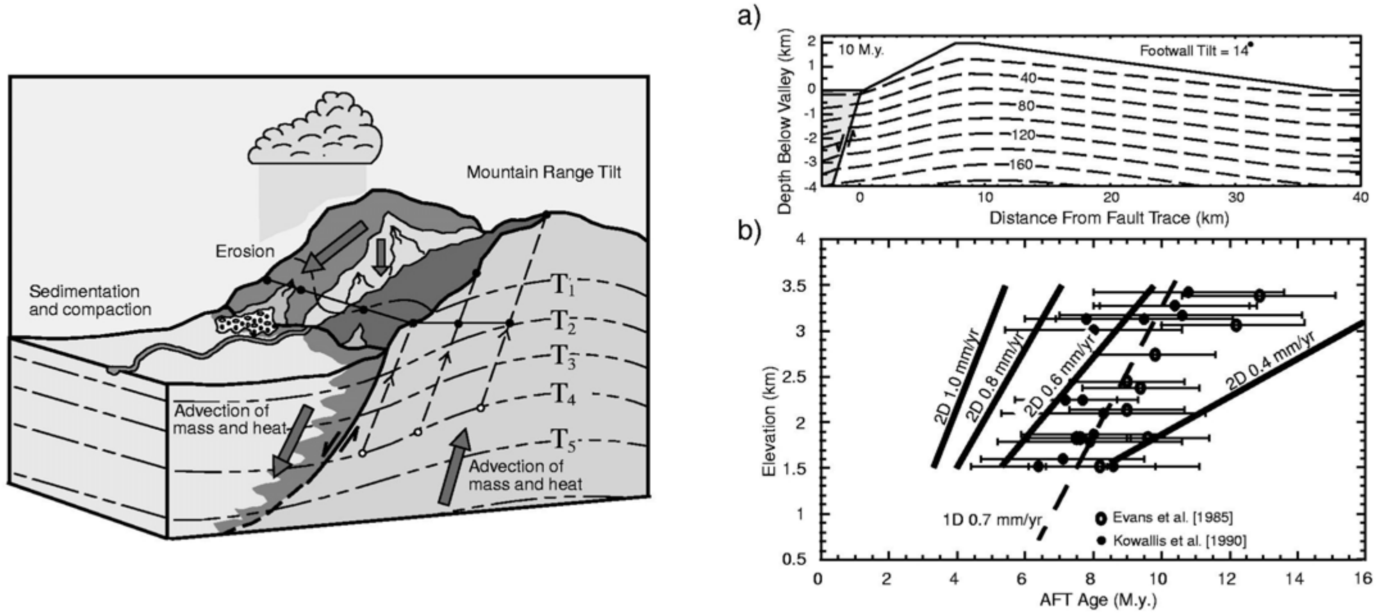

Figure 14. *Left: Generalized 3-D block diagram showing some principal factors that affect subsurface temperatures. Thin dashed lines represent surfaces (isotherms) of constant temperature (T_i). Connecting lines with arrows show relative particle motions of rocks containing thermochronometers (open circles) that are later sampled at the surface (closed circles). Large arrows show the direction of mass movement from tectonic and erosional forces. Right: (a) 2-D thermal model isotherms after 10 myr of exhumation at 1.0 mm/yr. (b) sample elevation versus apatite fission track ages for samples collected up Lone Peak near Salt Lake City. Dashed line represents previous interpretations of the footwall exhumation rate using 1-D thermal model assumptions. Solid lines represent 2-D thermal model predicted fission track ages for exhumation rates between 0.4 and 1.0 mm/yr. More realistic 2-D models suggest an exhumation rate of about 0.5 mm/yr compared to the 1-D thermal model predicted exhumation rate of 0.7 mm/yr. Modified from Ehlers et al. (2001).*

Miocene, Quaternary basalts and high heat flow imply continued activity relative to the GB interior. Today's central GB structure is compatible with an extension rate originally like that of the eastern area, but which has decreased since 5-10 Ma and has a fluid content lower bound diminished to 0.1-0.2 vol. % (Wannamaker et al., 1997b).

Reconnaissance long-period MT recordings by the University of Utah extend the GB data to span the period range T = 0.01 s to 10,000 s (Figure 17b). As with the earlier GB MT data, the LPMT results imply remarkable uniformity along each profile in the upper mantle at the resolving scales of MT, so that characteristic 1-D resistivity profiles referenced to the dense U. Utah profiling could be extended from a simple impedance data integral (following Wannamaker et al., 1997b). These are tabulated as layered models in Figure 17b and show that enhanced conductivity of the eastern GB relative to the GB interior extends well into the upper mantle. Note the low resistivity ~100 km depth in the eastern GB, interpreted to represent broad-scale asthenospheric upwelling in the region. Fluid/melt based mechanisms for conductivity in both the lower crust and upper mantle are strongly favored over graphite in this project area because ambient f(O_2) likely is well above the graphite stability limit, given the relatively high temperatures and long history of subduction (Carmichael, 1991; Balhaus, 1993; Brandon and Draper, 1996; Parkinson and Arculus, 1999).

If basaltic underplating currently is focussing under the TZ in non-uniform extension, lower crust and upper mantle should be even more conductive there than below the eastern GB. Preliminary 2-D inversion of new, similarly broadband MT data from the CP suggests this well may be the case (Figure 18). This inversion was constructed using the forward problem and jacobians described by Wannamaker et al. (1987) and DeLugao and Wannamaker (1996), with parameter step equations of Tarantola (1987). The parameter step is estimated by Gauss-Newton minimization of a weighted sum of data-model misfit plus model departure from an a-priori structure. The data are quite precise, having e.g. phase errors within 0.5^o commonly. For data x-coordinate (geoelectric strike) parallel to the GB-CP transition (~N15^{o}E), impedance skews almost invariably are <0.1. Also, normalized vertical magnetic field anomalies (K_{zy} in Figure 18b) associated with the TZ to the west approach 0.4 in amplitude, but values reflecting along-strike changes reach only a few percent. This 2-D nature of deep structure is in keeping with the ~N-S trends in the early geomagnetic variation studies (Figure 16). To ensure no conflicts from near-surface 3-D structures, just the components of data corresponding to current flow across strike (ρ_{yx} and ϕ_{yx}) plus along-strike components ϕ_{xy} and K_{zy} for period T > 20 s are included in the inversion, as these show more 2-D behavior in most cases (Wannamaker, 1999).

In the inversion of Figure 18a, smoothed 1-D models for the GB and CP data sets from the integrated impedances were input as the a-priori model. Next, the eastward extent of the conductive eastern GB profile was tested well to the east near 112^{o}W, only 20 km west of the CP profile, and actually well within proper CP physiography. Not even this test position is quite compatible with the data; the inversion places addi-

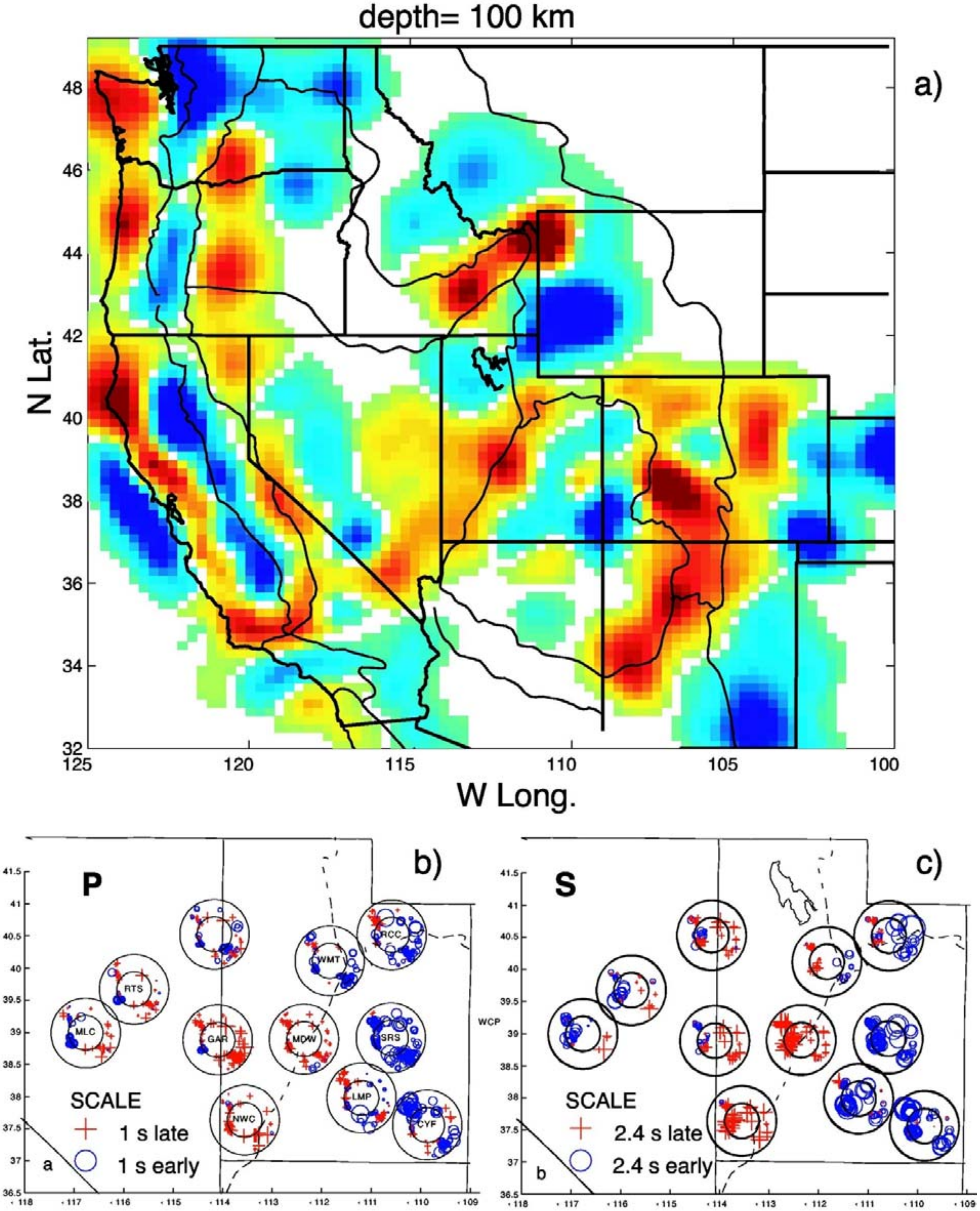

Figure 15. *Top (a): composite P-wave image of upper mantle seismic structure at 100 km depth beneath North America (K. Dueker, pers. comm.; Savage and Sheehan, 2000). Blue is high velocity mantle and red is low velocity mantle. The total range in P velocity is about 8%. Resolving scale is about 100 km. Using standard scaling relations, red regions are partially molten and blue regions are subsolidus. Bottom: Map view of travel time residual measurements at the CPGB PASSCAL portable array, plotted at the back-azimuths of the events, with radial position dependent on angle of incidence. Outer circle represents earthquakes located 30º away, inner circle represents earthquakes 90º away. Blue circles represent arrivals that are early with respect to the earthquake average. Red crosses represent late arrivals. State borders are shown as solid line, boundary between Colorado Plateau and Great Basin shown by dashed line. Left is P residuals (b), right is S residuals (c). S residuals are rescaled by factor of 2.4 to make sizes directly comparable to the P delays.*

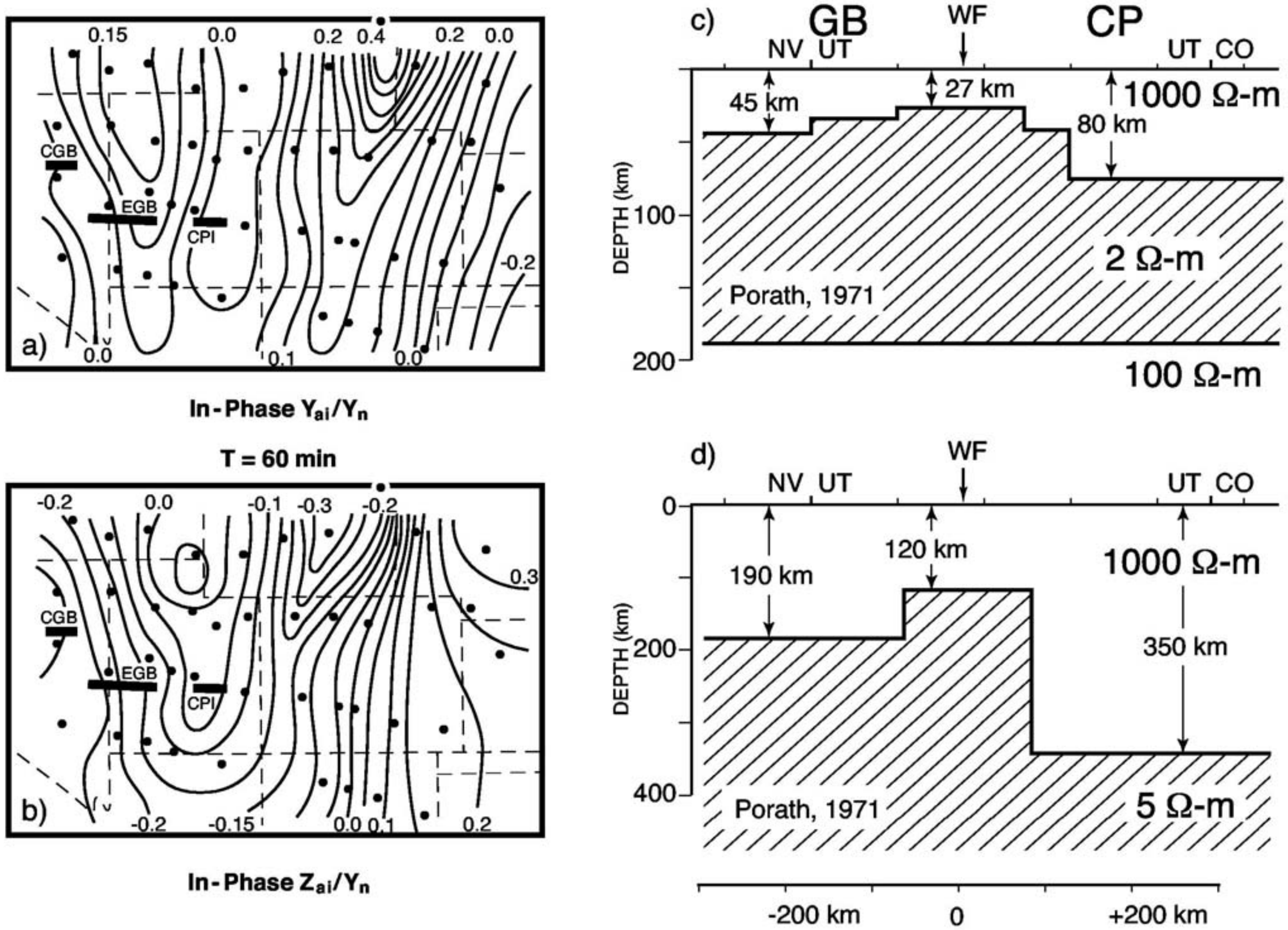

Figure 16. *Anomaly maps of horizontal (Y) and vertical (Z) magnetic field variations at a period of 60 minutes for Great Basin and Colorado Plateau region, together with alternate 2-D models of electrical resistivity (inverse of conductivity) structure providing equally good fit to observations (from Porath et al., 1970; Porath, 1971).* Y_{ia}/Y_n *refers to in-phase anomalous horizontal field normalized by external inducing field, while* Z_{ia}/Y_n *refers to normalized in-phase anomalous vertical field. Model section views derived from station profile nearest to southern proposed transect. Horizontal black rectangles are locations of modern, dense MT profiles of Wannamaker et al. (1997b) in the central and eastern Great Basin (CGB and EGB), plus easterly profile in the CP interior (CPI) discussed herein. Wasatch Front is WF.*

tional low resistivity at the easternmost limit of the very conductive region to achieve a sufficient fit, especially to K_{zy} and ϕ_{xy}. Hence, these newly analysed Colorado Plateau MT data show that very high GB conductivity persists far eastward of the southern Wasatch Front, and also that the highest conductivity may be under the easternmost Transition Zone. A gradient in upper mantle conductivity in fact persists across this line right to its eastern end (Figure 18a). The gradient is not an artifact of inversion smoothing, but follows from the gradient in both modes of the MT impedance across the profile (Figure 18b and c).

That higher conductivity may exist under the TZ even compared to the eastern GB suggests enhanced magmatic extensional activity in the former. First, a simple diffusive decay of heat eastward from the physiographic boundary (Bodell and Chapman, 1982) should induce a similarly diffuse transition in conductivity, since such gradational heating should exsolve diminishing quantities of fluids eastward under fading prograde metamorphic conditions (e.g., Wannamaker, 2000). Second, if the limited TZ normal faulting at the surface represented pure shear at depth, lower lithospheric fluid release should not occur to increase conductivity. This is because such extension *per se* actually cools the crust by bringing material surfaceward in the solid state (e.g., McKenzie, 1978). In a closed system, cooling of this sort should absorb fluids in retrogression. A deep, simple shear detachment may also induce some upper plate, prograde metamorphism as hotter footwall material is brought up into contact. However, in the absence of induced melting, this too should fade or dip eastward in our case as the detachment goes to greater depth. We are left with the likelihood that new conductive fluids are being introduced by exsolution from crystallizing magmas near the base of the crust of the TZ. Several volume percent of H_2O-CO_2, highly saline brines typically are released in this process (Frost et al., 1989; Wannamaker et al., 1997b), providing abundant conductive fluid. The domains of greatest fluid concentration will be those most prone to diffusion creep rheology mechanisms (Kohlstedt et al., 1995; Tullis et al., 1996), and thus high conductivity also provides an image of zones with very weak rheology under the eastern GB and TZ.

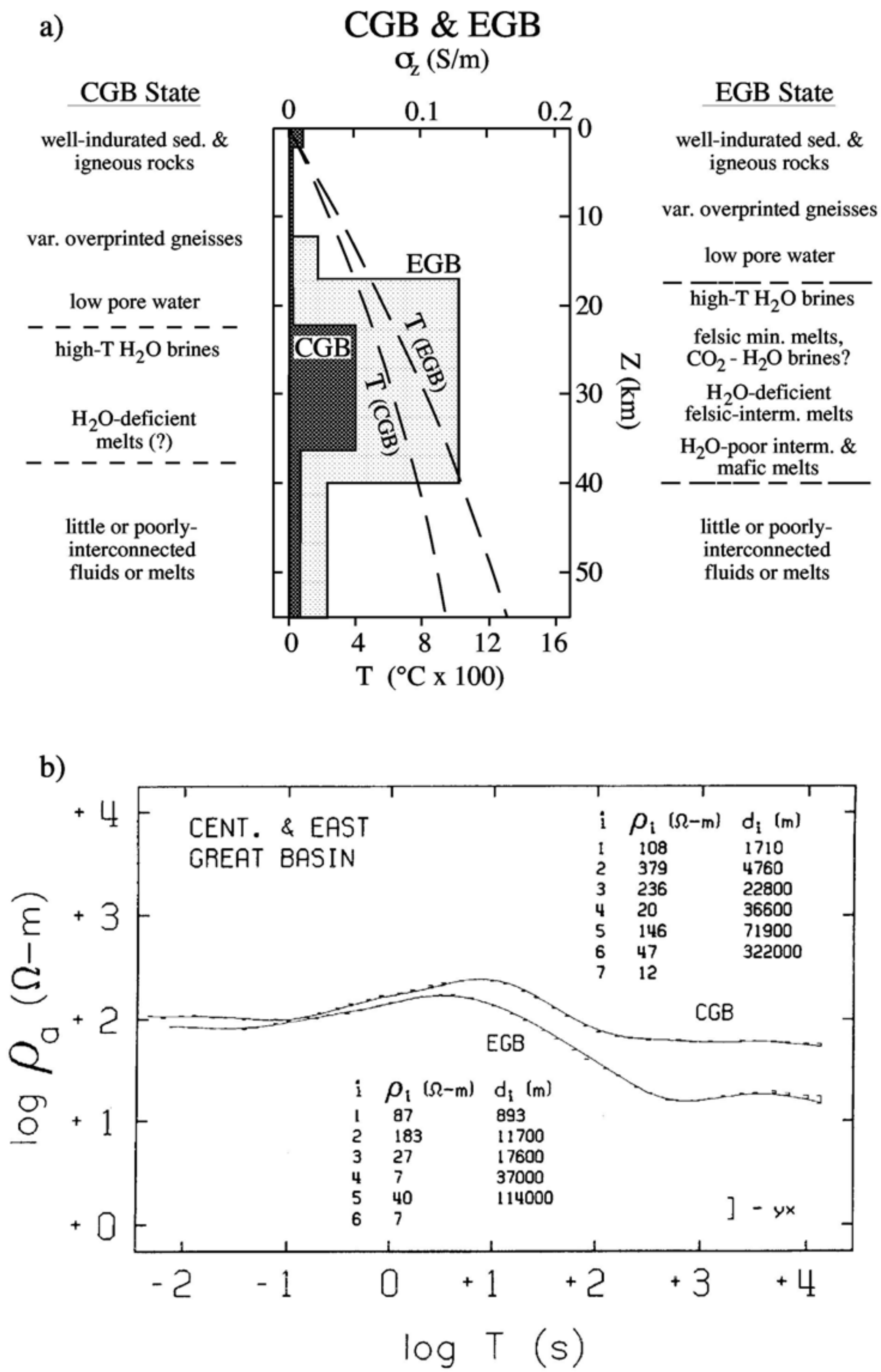

Figure 17. *Top (a): Layered conductivity models for central and eastern Great Basin compared to geotherms and likely physico-chemical states in the subprovinces. Fluid states assume intermediate composition, high-grade metaigneous rocks on average in the lower crust. Electrical properties are plotted as conductivity to emphasize differences in lower crust of the two regions, and because conductivity is more linearly proportional to content of conductive grain-boundary phases (modified from Wannamaker et al., 1997b). Bottom (b): Integrated impedance, characteristic apparent resistivity sounding curves for central and eastern GB of Wannamaker et al., augmented with 5 reconnaissance long-period (to 13,000 s) sites on each profile. Error intervals are 2 s.d. Layered earth models for each region are tabulated in plot. Note diagnostically lower apparent and model resistivity values of lower crust and upper mantle in active eastern Great Basin.*

The resistivity at mantle depths even under the CP is not very high (Figure 18a), especially below 80-100 km, and probably indicates some interconnected melt fraction (Faul et al., 1994; Roberts and Tyburczy, 1999). The gradient in resistivity with depth below ~100 km also appears low, much less than is compatible with solid-state conduction and a conductive geotherm even with hydrous defects (e.g., Karato, 1990). This suggests presence of an asthenosphere although future specific sensitivity tests on structural bounds are warranted (e.g., Jones, 1999). Furthermore, a properly resolved conductivity structure in the Transition Zone and in the CP requires complete data coverage through the TZ as well as further to the east. We expect this can shed light on support mechanisms for the elevated CP by imaging regions of contrasting thermal activity (Sheehan et al., 1997; Lowry et al., 2000).

Comparison with the S. Sierra and Arizona TZ. - A similar condition of non-uniform deformation may exist under the Sierra Nevada (Jones et al., 1992; Wernicke et al., 1996; Jones and Phinney, 1998). The absence of substantial variation in Moho depth across a profound upper crustal boundary in Cenozoic stretching, combined with little evidence for magmatic additions to the crust, seems to demand vertical variations in strain. In the Sierra Nevada, high elevations and virtually undeformed batholithic upper crust mask active processes below (Jones, 1987; Wernicke, 1992). These include temporal changes in xenoliths and basalt geochemistry, low mantle seismic velocities, high heat flow, low electrical resistivities, a nearly flat Moho ~35 km b.s.l., and seismically inferred strain markers (Wernicke et al., 1996; Ducea and Saleeby, 1998; Fliedner et al., 1996; Saltus and Lachenbruch, 1991; Ruppert et al., 1998; Park et al., 1996; Jones and Phinney, 1998).

Thinning of the lower crust under the Sierra might represent flow of Sierran lower crust towards the Great Basin (Wernicke, 1992). Jones and Phinney (1998) have resolved shallowly W-dipping, P-S seismic conversion interfaces in the lower crust here. One of these indicates a thin, low-velocity zone with anisotropic characteristics, tentatively correlated with a primary shear interface over possibly fluidized lower crust. On the other hand, the presence of the Sierra in the hanging wall of the west-dipping faults of this region also allows thinning by mere mechanical (non-flow) means as extension shifts westward with depth. It is difficult to separate primary Late Cenozoic structures in these images from possibly reactivated Mesozoic fabric, much of which would have had the same dip previously, thereby clouding interpretations of stress regime. However, unlike the Sierran side of the Great Basin, the Utah TZ lies in the footwall of the familar, west-dipping extensional systems immediately adjacent to it (Wernicke, 1992). East-dipping anisotropic shear zones resolved seismically would more likely be primary Late Cenozoic structures, less obscured by pre-existing structures than in the Sierran region.

The nearest high quality refraction study elsewhere across the Colorado Plateau - Basin and Range transition is the USGS PACE survey in west-central Arizona 500 km south of our Utah study region (Goodwin and McCarthy, 1990; McCarthy et al., 1991; Parsons et al., 1992). The southern Basin and Range-Colorado Plateau transition here is not a direct analogue to the GB-CP TZ because in west-central Arizona the highly extended mid-Tertiary metamorphic core complex belt is immediately adjacent to the Colorado Plateau (Coney and Harms, 1984; Spencer and Reynolds, 1990). As a result, this transition zone is underlain by low-angle detachment structures which exhumed the core complexes from beneath the margin of the Colorado Plateau (McCarthy and Parsons, 1994; Reynolds and Spencer, 1985). PACE refraction profiles indicate an extended low-velocity crust (~30 km) which gradually thickens to ~40 km within the southwest Colorado Plateau (McCarthy and Parsons, 1994). Limited wide-angle and teleseismic data profiling there suggest thinning of the lithosphere at the TZ of Arizona (Benz and McCarthy, 1994; Zandt et al., 1995). Seismic reflectivity patterns suggest processes of mafic inflation plus lateral ductile flow were present to accommodate core complex exhumation (Hauser et al., 1987; Henyey et al., 1987; McCarthy and Parsons, 1994). Furthermore, PACE results were taken in the relatively inactive Southern Basin and Range. Unlike the active process of west-central Utah, there is limited seismicity, lower heat flow, lack of strong N-S conductivity trends, little geodetically-inferred motion, and evidently no rift pillow. In contrast to the southwest CP, about 200 km of modern extension associated with the northern Basin and Range separates the GB-CP Transition Zone in central Utah from the metamorphic core complex belt (i.e., Snake Range).

Role of Magmatism in Lithospheric Thinning

There are two main facets to the magmatic process which pertain to the Great Basin-Colorado Plateau transition (Figure 19). The first is the ongoing basaltic or bimodal volcanism, including its thermal, compositional and geodynamic controls (Figure 19a). The second is the prodigious plutonism of south-central and southwestern Utah, which undoubtedly has changed the deep crust here profoundly and forms its own natural laboratory for distinguishing rifting processes in magmatic versus amagmatic regimes (Figure 19b).

Present-Day Magmatic Regime

The classic and familiar mechanism for magmatism during extension is that of decompression melting of upwelling mantle. The amount of magmatism produced in a 1-D model of this process is determined by two variables: mantle temperature and lithospheric thinning (e.g., Langmuir et al., 1992; McKenzie and Bickle, 1988). Thickness of intruded igneous layers (e.g., "underplate") directly relates to extension and mantle temperature increases.

The concept of mafic underplating in continental extension zones has become well entrenched in the literature since its introduction in the early 1970's (e.g., Meissner, 1973). So-called "rift pillows," with V_p = 7.2-7.6 km/s and thicknesses up to 15-20 km, were interpreted beneath a number of continental rift zones (e.g., summaries in Mooney et al., 1983; Catchings and Mooney, 1988; Thompson et al., 1989). Similarly, the most recent model for crustal structure across the GB-CP transition in Utah (Loeb and Pechmann, 1986; reprinted in Smith et al., 1989), shows a 15-km-thick layer of velocity 7.5 km/s overlying mantle of 7.9 km/s (Figure 13). The "anomalous 7.5 km/s layer" (Smith et al., 1989) may represent either mis-interpreted upper mantle on unreversed profiles (Pakiser, quoted in Smith et al., 1989) or a thick mafic underplate, the interpretation favored by Smith et al.

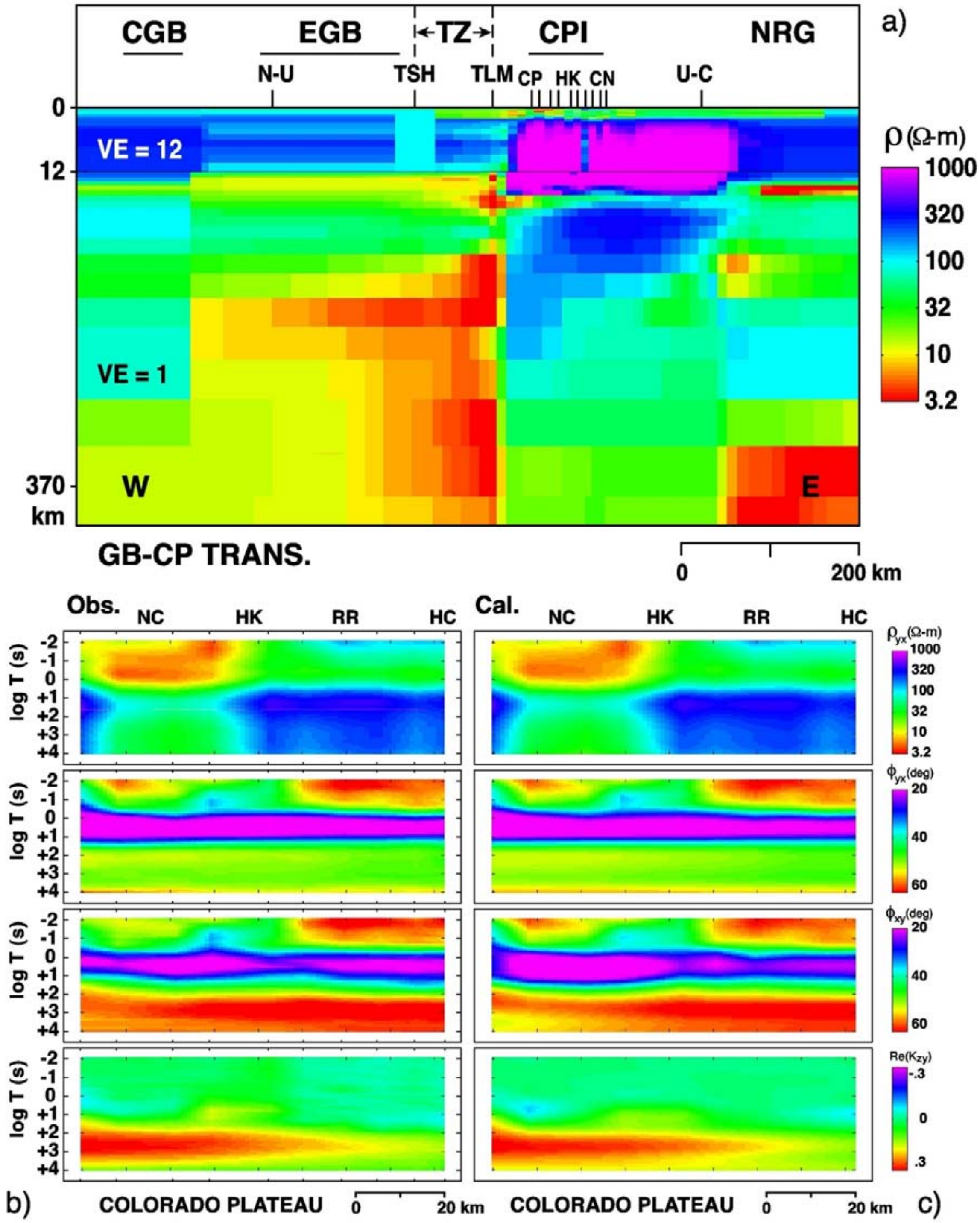

__Figure 18.__ Top (a): Inversion model of MT data to 10,000 s period from the CP. Projection of data profiles in central and eastern Great Basin (CGB and EGB) and Colorado Plateau interior (CPI) drawn overhead. Note pronounced low resistivities under eastern Transition Zone (TZ), and upper mantle gradient under CP. Vertical exaggeration (VE) in upper 12 km is 6:1. Resistivity values in ohm-m. Geographic features include Tushar Rangefront (TSH, S. extension of Wasatch Front), Thousand Lake Mountain (E. limit of TZ normal faulting), Capital Reef (CP), Hanksville (HK), and Canyonlands (CN). NRG is northern Rio Grande Rift. Bottom: pseudosections of observed (left, a) and computed (right, b) MT results for profile of 10 stations in the Colorado Plateau. In pseudosections, log period (T, in s) serves as ordinate and horizontal distance as abscissa for contour plots of MT response amplitude (Vozoff, 1991). Quantities shown include cross-strike (TM mode) apparent resistivity and impedance phase (ρ_{yx} and ϕ_{yx}, top two panels), along-strike (TE mode) impedance phase (ϕ_{xy}, third panel), and the real part of the normalized vertical field due to along-strike current flow (Re(K_{zy}), fourth panel). For computed responses of the latter two quantities, only results for T > 20 s are relevant. Geographic locations include North Caineville Mesa (NC), Robbers Roost (RR) and Horseshoe Canyon (HC).

Several observations and inferences raise doubts about the general existence and/or significance of "rift pillows", however. First, thick, high-velocity underplated layers are not observed beneath rifted continental crust on volcanic rifted margins. Instead, the high-velocity lower crust (HVLC) commonly observed on volcanic rifted margins does not "underplate" existing continental crust, but rather occurs seaward of continental crust, beneath thick extrusive basalts erupted during and immediately after complete continental rupture -- that is, the HVLC on rifted margins is properly seen as a thickened oceanic "layer 3" (e.g., Holbrook et al., 1994a,b; Lizzaralde and Holbrook, 1997). The absence of significant mafic underplating beneath continental crust on volcanic margins, where rifting proceeded to completion and mantle temperatures were certainly anomalously high, begs the question of how such underplated layers could exist beneath continental rift zones, where implied factors of extension (β) from surface indications rarely exceed β = 2. Second, some models of melting beneath rift zones and mid-ocean ridges show that underplated layers of 15 km thickness are virtually impossible for low extension factors (Langmuir et al., 1992; McKenzie and Bickle, 1988). For β = 2, generation of only 10 km of underplate would require a mantle temperature anomaly of at least 200°C (White and McKenzie, 1989). Generation of 15 km of underplate in oceanic regimes requires both high temperature (150-200°C anomaly) and relatively high extension factors (β = 4-5).

Finally, in many continental rift zones where rift pillows were interpreted, later, higher-quality seismic data resulted in seismic velocity models with thinner or absent HVLC. For example, the Rhinegraben, once considered a classic rift pillow locale (e.g., Edel et al., 1975), is now believed to lack a significant HVLC (e.g., Gajewski et al., 1987; Zucca, 1984). A similar evolution occurred in studies of the East African rift, where recent data suggest little or no lower crust with velocities >7.0 km/s and addition of only 3-4 km of magmatic material to the lower crust (e.g., Mechie et al., 1997). In the Rio Grande Rift, Prodehl and Lipman (1989) note that the only refraction survey implying a rift pillow is unreversed, and other surveys do not require its presence.

On the other hand, use of MORB-like source compositions to model extensional magmatism is not appropriate for the S.W. United States because of the evident involvement of volatile-metasomatized lithospheric mantle in the source (Kempton et al., 1991; Hawkesworth et al., 1995). Large amounts of melt can be generated during adiabatic extension because of the lower-T, steeper solidus curve of the metasomatic material (Harry and Leeman, 1995). This model is applied to the Mid-Cenozoic plutonic rocks as well as the Late Cenozoic basalts, which are considered to differ primarily in regional thermal regime and state of stress (e.g., Wernicke, 1992). It predicts ~50% extension before deeper, asthenospheric melts form and, coincidentally, the generation of lithospheric melts has completed. The model is consistent with a 35-50 km depth range estimated for the Quaternary lithospheric basalts of the Black Rock (Sevier) Desert area, and greater (50-70 km) depths estimated for the asthenospheric basalts more common toward southernmost Utah (Nelson and Tingey, 1997; N-S trend of Qb in Figure 2). Mantle electrical resistivities below S. W. Utah decrease strongly at the 50-100 km depth range, consistent with melting (Figure 18a).

Further complications arise from multi-dimensional effects. Melt- and temperature-dependent viscosity and buoyancy may concentrate the amount and degree of basaltic production into a zone only a few tens of km wide (Buck and Su, 1989; Cordery and Phipps Morgan, 1993; Kincaid et al., 1996; Boutelier and Keen, 1999). This seems pertinent for rift pillows because V_p of 7.2-7.6 km/s is high even for gabbro and would imply some olivine component with high degrees of melting (Kelemen and Holbrook, 1995). Harry and Bowling (1998) conclude for lithospheric-scale simple shear rifting that a), attendant mantle thinning can be high and b), subsequent mantle melting can be large without high potential temperatures. Melting occurs only in a narrow zone in the last stages of rifting before oceanic crust is produced if extension rate is moderately high (>1.5 cm/yr) and there is only minor crustal lateral heterogeneity (Harry and Bowling, 1999).

We therefore are left with a dilemma for the Great Basin-Colorado Plateau Transition Zone. From better quality seismic surveys worldwide, we may hypothesize that underplated layers during continental rifting ("rift pillows") are rare and, when present, are thin (<5 km). That is, they are much less significant than previously supposed. Conversely, the Utah TZ appears to be a stubborn holdout against this suggestion. Various data sets show potential evidence for a HVLC beneath the Wasatch Front (e.g., refraction data of Braile et al., 1974; Smith et al., 1989, and regional network data of Loeb and Pechmann, 1986). Furthermore, Sheehan et al. (1997) observe a strong, mid-crustal receiver function conversion below the Wasatch Front suggestive of the top of such a pillow from their two receiver function sites along the physiographic GB-CP transition (MDW and WMT) (Figure 20), also implying considerable strike extent. Theoretical simulations so far range broadly enough to allow a high-V_P zone under the Transition Zone, albeit perhaps narrower than others have drawn (Figure 8) due to focused, non-uniform extension.

Influence of Pre-Existing Magmatic Belt

Shallow plutonic rocks dominate the geology of the southern Utah TZ ~38.5°N latitude (Figure 2). Major Oligocene-Miocene magmatic centers include the Beaver Lake-Mineral Mountains plutonic complex (Nielson et al., 1986; Coleman and Walker, 1994; Coleman et al., 1997) at the border between the Basin and Range and the Transition Zone, and the Marysvale volcanic field (Rowley, 1998; Rowley et al., 1998) in the Transition Zone itself. That the main source region for these plutons lies in the upper mantle means their intrusion is a net addition of material to the crust (Figure 19b). Although the surface manifestations of intrusives in the Mineral Mountains-Marysvale field are copious (Rowley, 1998; Rowley et al., 1998), petrogenetic modeling of thermal budget, fractional crystallization and possible crustal assimilation indicate the silicic endmember intrusives may be only a fraction of the total magma input to or reconstituted in the crust (e.g., Grunder, 1995). The bulk of hidden plutonic material includes source basalt, and a large volume of remelted, hybridized earlier crust. While details of emplacement mechanisms have been debated (Gastil, 1979; Paterson and Fowler, 1993), displacement or thorough annealing of pre-existing lower crust to form a massive, more mafic plutonic root zone seems inescapable.

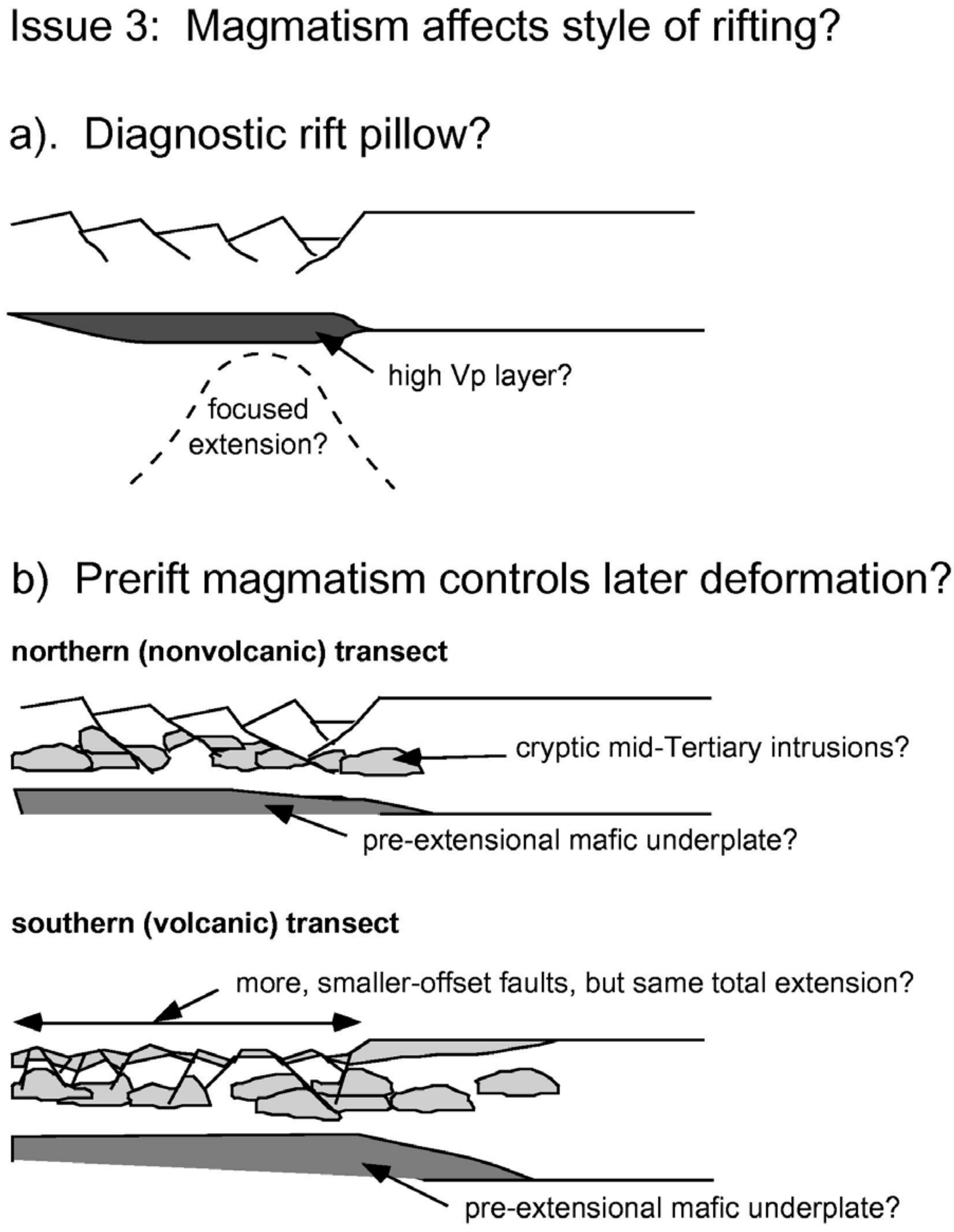

Figure 19. *Possible effects of rifting on current magmatism, and of prior magmatism on style of present rifting: a), High-V_p rift pillow forms in response to focused extension in uppermost mantle. Lack of such feature constrains extension rate and domain; b), Pre-extensional magmatism affects style of surface deformation and modifies strength of lithosphere. However, plutonism may be cryptic in amagmatic areas obscuring its role unless uncovered geophysically.*

A more siliceous composition such as possibly represents unintruded Great Basin crust of central Utah could, with its weaker rheology, facilitate lower crustal flow westward from beneath the Transition Zone (Wernicke, 1992) (Figure 12b), and perhaps serve as a control on the TZ eastern limits. However, the modification of prior lower crust under south-central Utah to a recrystallized and more mafic composition should represent a fundamental strength modification. At a specific temperature, pyroxene-rich lithologies are considerably stronger than the siliceous compositions often used to model crustal rheologies (Kohlstedt et al., 1995). Furthermore, enhanced lithospheric strength in south-central Utah may have a mantle component too; the source regions would now be more refractory (Karato and Jung, 1998).

To the contrary, compositional strengthening could be counteracted by any residual thermal anomaly. Although Great Basin-wide plutonism in the Middle Cenozoic was interpreted to cause even higher temperatures at depth than in the somewhat embrittled province today (Harry et al., 1993), substantial magmatic gaps such as under central Utah may have largely escaped this heating (Perry et al., 1993). It has profound implications for defining the strength contrast across the TZ in the early Miocene, such that a CP lithospheric keel perhaps was not even created here until the Late Miocene. That the current, N-S regime could suddenly develop across such a strong, thermally-related, E-W strength grain would imply that strength heterogeneity in this case is unimportant, or that a yet stronger pre-existing heterogeneity running N-S was in control. Moreover, the simple fact that the southern Utah TZ crust has been thickened by magmatism could make it more deformable than that northward.

If cumulative upper crustal extension in the southern Utah TZ compares with that in central Utah, it could support the notion of a detachment such as the SDD under both magmatic and amagmatic areas (Coleman et al., 1997). Early reflection data did not image such a structure distinctly in southern Utah (Smith and Bruhn, 1984). On the other hand, perhaps substantial plutonism occurred under the amagmatic mid-Utah region as well, but is cryptic to the surface. If so, both regions may have received similar thermal and magmatic input and suffered similar strength modifications over most of the lithosphere. Differences in deformational style in this case would be basically superficial, the surface products would not be diagnostic of deep processes, and the style of deep deformational structure would be similar for both regions. Improved seismic velocity models could resolve these possiblities, since distinct velocities are expected for original Great Basin lower crust (V_p = 6.6-6.8 km/s), basaltic plutonic roots (V_p = 6.9-7.2 km/s), and possible mafic-ultramafic rift pillow volumes (V_p = 7.2-7.6 km/s) (e.g., Rudnick and Fountain, 1996; Christensen and Mooney, 1995).

SUMMARY AND CONCLUSIONS

Despite its excellent exposures and numerous investigations, there remain wide-ranging possibilities for the geological/geophysical evolution of the GB-CP Transition Zone of central and southern Utah. Based on previously reviewed evidence, we suggest the possibilities may fall between two disparate scenarios (Figure 21). These two models may become distinguishable with improved space-time data on geologic structure, thermochronology and geochemistry, and geophysical data. Neither model invokes Pacific plate boundary forces, thought not to propagate this far to the interior of the U.S (Sonder and Jones, 1999). Restoration possibilities appear more clear-cut from the present to the pre-extensional Middle Cenozoic (Oligocene), since this is a time of low relief in the region. The population density of Utah along the TZ is increasing dramatically, and more study is warranted to improve thermal regime estimates, future earthquake locations, depth range of seismicity, and the framework of driving stresses.

In one end-member model of the Transition Zone (Figure 21a), extension is limited to the eastern GB where the greatest lithospheric thinning underlies the visible rifting at the surface. This fundamental boundary was established in the Early-Middle Miocene and the limited extension of the TZ is a mere weak continuation reflecting incipient gravitational collapse (Hopper and Buck, 1998). The subdued expression may be the result of stronger CP-like crust under the TZ, or aided by pre-existing weaknesses some of which may date from the Precambrian. The TZ may have formed initially in the Miocene also, or developed continuously eastward since then enabled by thermal diffusion from the west. Alternately, the TZ may have started abruptly in the Latest Miocene-Early Pliocene time in response to slowing of North American absolute motion to the WSW and partial reduction of compressive stress associated with basal traction forces. If TZ uplift is contemporaneous, these three timing possibilities should be visible in the paleobotany records. There would be a lack of E-dipping shear structures under the TZ, either structural or seismic, and lithosphere would not be thinned beyond that characteristic of the eastern GB. Teleseismic delay, electrical conductivity and heat flow should be intermediate between the GB and the CP. However, while the sparse heat flow data may be questioned, the conductivity at least in south-central Utah appears high under the TZ, not intermediate. Gravity/topography coherence analysis allows the possibility of an elastic thickness more like that of the CP. Support of the high TZ is essentially isostatic due to original thickened crust, if high crustal thickness exists for the TZ Pakiser (1989).

In the other endmember model for the Transition Zone (Figure 21b), non-uniform extension involves both lower crust and upper mantle under the Transition Zone with significant Late Cenozoic magmatism at depth (rift pillow). A great diversity of structures is likely here, even disregarding Mid-Cenozoic plutonism (not drawn). This scenario directly relates to the possible high-V_P (7.4-7.5 km/s) medium discussed with Figures 8 and 20, and the role of magmatism. Its position at the Moho implies large lithospheric thinning whose eastern limit may be bounded by more mafic CP crust. The locus of current melting may be significantly offset from surface extrusions, and low resistivity images and receiver functions under the TZ are in keeping with this. The prospectively thinned lower crust should be a first-order seismic feature, as potentially are shear zones in the lower crust dipping eastward at probably low angles. If a discrete simple shear component exists, the most thinned crust could lie under the Transition Zone, or at least its western border. A zone of lithospheric melting or asthenospheric upwelling presumably accompanies any rift pillow and could reside beneath it or yet further east (Figure 21b). In this regard, the eastern GB and TZ may resemble models of Driscoll and Karner (1998), Gartrell (2000) and Ebinger and Casey (2001) for continental crust approaching complete breakup. If upwelling is driven by convergence of the CP lithosphere with the GB asthenosphere, then it may date from the earliest Miocene, or it may be evident in the latest Miocene record as the result of plate motion change. CP keel impact upon GB asthenosphere may be inducing rift-parallel (out-of-plane) flow, consistent with seismic anisotropy. High heat flow should occur from latent heat contributions during melt crystallization, and an abrupt heat flow transition to the CP interior would indicate shallowness of magmatic/hydrothermal penetration. Any time difference between crustal thinning and mantle support of the TZ could be evident in the paleobotany records.

Wernicke (1992) suggests there is flow of lower crust from under the Transition Zone westward but little thinning of the mantle as a model for the older, inactive TZ in Arizona. This may be a response to weaker rheology if Transition Zone lower crust was more like that of the GB than the CP. However, without lithospheric thinning to provide buoyancy, some mantle force such as basal traction (e.g., CP keel impact) must make up the need for elevation support in the central Utah TZ. Mantle doming forces can cause crustal flow of opposing directions versus depth especially in the later stages of rifting (Huismans et al., 2001).

Southward in the Mineral Range-Marysvale magmatic field, large additions of mafic material to the lower crust of Figure 21b seem likely. Especially if the lower crust here was previously more siliceous (GB-like?), mafic intrusion would inhibit Late Cenozoic extension via compositional hardening and thermal annealing of previous weaknesses. On the other hand, if mantle upwelling and melting is driven by sub-lithospheric

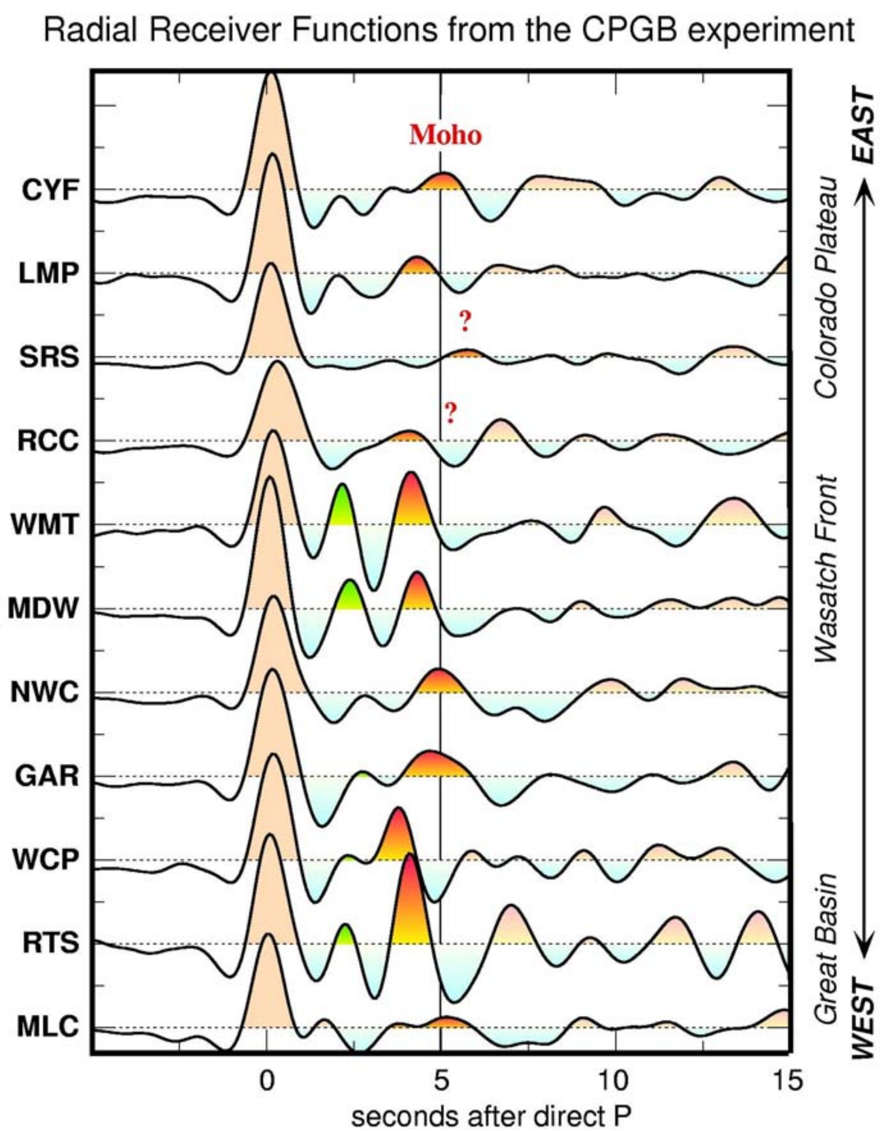

Figure 20. *Receiver functions for the CPGB stations, arrayed roughly on an east-west profile, showing the change in midcrustal converted energy along the Wasatch Front (green, both sites WMT and MDW) and the decrease in Moho amplitude (orange) going into the Colorado Plateau. The latter suggests lower crust-mantle contrast (i.e., higher crustal velocity) for CP. After Sheehan et al. (1997).*

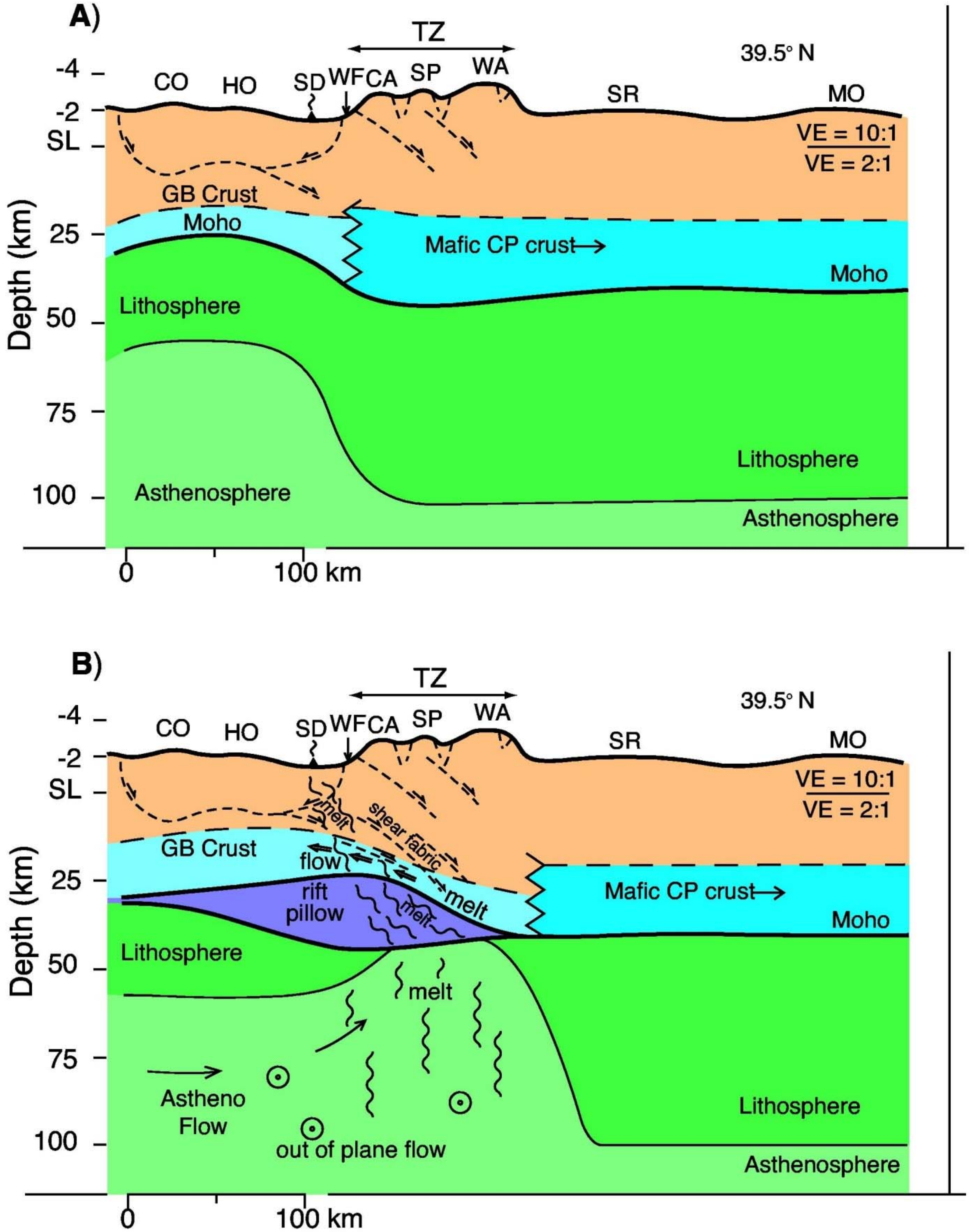

Figure 21. *Schematic depiction of two possible lithospheric cross-sections representing differing geodynamic processes responsible for presently observed characteristics of the GB-CP Transition Zone, and surrounding regions. Geographic features include House Range (HO), Confusion Range (CO), Sevier Desert (SD), Wasatch Front (WF), Valley Mountains (VA), San Pitch Mountains (SP), Wasatch Plateau (WA), San Raphael Swell (SR), and Monument Uplift (MO). Out-of-plane flow suggested to be due to impact of Colorado Plateau lithospheric keel upon Great Basin asthenosphere. Note vertical exaggeration change at sea level (SL).*

processes, it is the strength of the mantle which may control their vigor, and a distinction here with the northerly amagmatic area is less certain. Isotopic evolution of the Mineral Range-Marysvale field may reflect long-term modification of crust and mantle. Beyond this plutonism, however, the Late Precambrian margin rifting and subsequent overthrust thickening and fore-deep sediment blanketing swung well westward of the TZ here suggesting a stronger TZ crust in the south prior to Middle-Late Cenozoic events. A rift pillow or east-dipping shear fabric overprinting the deep plutonic rocks would prove its youth and confirm continuity of process along strike.

ACKNOWLEDGMENTS

The synthesis of investigations in this paper was supported under U.S. National Science Foundation grants EAR95-96068 and EAR96-16450 and Dept. of Energy contract DE-FG07-00ID13891. Wideband MT data in the Colorado Plateau were collected by Jeffery Johnston and Louise Pellerin while students at the University of Utah. Helpful reviews were provided by Peter Malin and Erika Gasperikova. Doug Jensen produced or finalized many of the illustrations. We thank volume editor Mel Erskine for his substantial efforts and patience.

This paper is dedicated to the memory of Arthur C. Ekker, late owner of the Robbers Roost ranch and icon of the American cowboy.

REFERENCES

Allmendinger, R. W., 1992, Fold and thrust tectonics of the western United States exclusive of the accreted terranes, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., Boulder, CO, p. 583-607.

Allmendinger, R. W., Sharp, J. W., Von Tish, D., Serpa, L., Brown, L., Kaufman, S., Oliver, J. E., and Smith, R. B., 1983, Cenozoic and Mesozoic structure of the eastern Basin and Range province, Utah, from COCORP seismic-reflection data: Geology, v. 11, p. 532-536.

Allmendinger, R. W., Farmer, H., Hauser, E., Sharp, J., von Tish, D., Oliver, J., and Kaufman, S., 1986, Phanerozoic tectonics of the Basin and Range-Colorado Plateau transition from COCORP data and geologic data: a review, in Barazangi, M., and Brown, L., editors, Reflection seismology: the continental crust, AGU Geodynamics Series, v. 14, p. 257-268.

Allmendinger, R. W., and Royse, F., 1995, Jr., Is the Sevier Desert detachment of west-central Utah a normal fault?, Comment: Geology, v. 23, p. 669.

Anders, M. H. and Christie-Blick, N., 1994, Is the Sevier Desert reflection of west-central Utah a low-angle normal fault?: Geology, v. 22, p. 771-774.

Anders, M. H., Christie-Blick, N., and Wills, S., 1995, Is the Sevier Desert detachment of west-central Utah a normal fault?, Reply: Geology, v. 23, p. 670.

Anders, M. H., N. Christie-Blick, and S. Wills, 1998, Extensional collapse along the Sevier Desert reflection, northern Sevier Desert basin, western United States, Comment: Geology, v. 26, p. 474.

Anderson, R. E., and Barnhard, T. P., 1987, Neotectonic framework of the central Sevier Valley area, Utah, and its relationship to seismicity, in Gori, P. L., and Hays, W. W., editors, Assessment of regional earthquake hazards and risk along the Wasatch Front, Utah, Reston, Virginia, U. S. Geol. Surv., p. F1-F134.

Archambeau, C. B., Flinn, E. A., and Lambert, O. G., 1969, Fine structure of the upper mantle: J. Geophys. Res., v. 74, p. 5825-5865.

Armstrong, P. A., Kamp, P. J., Farley, K. A., Chapman, D. S., Ehlers, T. E., and Constenius, K. N., 2000, Middle Miocene to Recent unroofing of the Central Wasatch Mountains: the transition from the Basin and Range to the Colorado Plateau: Geol. Soc. Amer. Abstr. Prog., Annual meeting, p. A-44.

Armstrong, R. L., 1968, Sevier orogenic belt in Nevada and Utah: Geol. Soc. Amer. Bull., v. 78, p. 429-458.

Armstrong, R. L., 1982, Cordilleran metamorphic core complexes: Ann. Rev. Earth Planet. Sci., v. 10, p. 129-154.

Armstrong, R. L., and Ward, P., 1991, Evolving geographic patterns of Cenozoic magmatism in the North American Cordillera: the temporal and spatial association of magmatism and metamorphic core complexes: J. Geophys. Res., v. 96, p. 13,201-13,224.

Atwater, T., and Stock, J., 1998, Pacific-North American plate tectonics of the Neogene southwestern United States: an update: Internat. Geol. Rev., v. 40, p. 375-402.

Askren, D. R. Roden, M. F., and Whitney, J. A., 1997, Petrogenesis of Tertiary andesite lava flows interlayered with large-volume feksic ash-flow tuffs of the western U.S.A.: J. Petrol., v. 38, p. 1021-1046.

Axen, G. J., 1998, The Caliente-Enterprise zone, southeastern Nevada and southwestern Utah, in Faulds, J. E., and Stewart, J. H., editors, Accommodation zones and transfer zones; the regional segmentation of the Basin and Range Province, Geol. Soc. Amer. Spec. Pap. 323, p. 181-194.

Axen, G. J., and Bartley, J. M., 1997, Field tests of rolling hinges: existence, mechanical types, and implications for extensional tectonics: J. Geophys. Res., v. 102, p. 20,515-20,537.

Axen, G. J., Taylor, W. J., and Bartley, J. M., 1993, Space-time patterns and tectonic controls of Tertiary extension and magmatism in the Great Basin of the western United States: Geol. Soc. Amer. Bull., v. 105, p. 56-76.

Axen, G. J., Selverstone, J., Byrne, T., and Fletcher, J. M., 1998, If the strong crust leads, will the weak crust follow?: GSA Today, v. 8, no. 12, p. 1-8.

Balhaus, C., 1993, Redox states of lithospheric and asthenospheric upper mantle: Contr. Miner. Petr., v. 114, p. 331-348.

Bartley, J. M., and Walker, J. D., 1999, Structure of Sevier Valley and its role in the Colorado Plateau-Basin & Range transition: Geol. Soc. Amer. Abstr. Prog., v. 31, no. 4, p. A-2.

Bartley, J. M., and Wernicke, B. P., 1984, The Snake Range decollement interpreted as a major extensional shear zone: Tectonics, v. 3, p. 647-657.

Beghoul, N., Barazangi, M., and Isacks, B. L., 1993, Lithospheric structure of Tibet and Western North America: mechanisms of uplift and a comparative study: J. Geophys. Res., v. 98, p. 1997-2016.

Bennett, R. A., Davis, J. L., and Wernicke, B. P., 1999, Present-day pattern of Cordilleran deformation in the western United States: Geology, v. 27, p. 371-374.

Bennett, V. D., and DePaolo, D. J., 1987, Proterozoic crustal history of the western United States as determined by isotopic mapping: Geol. Soc. Amer. Bull., v. 99, p. 674-685.

Benz, H. M., and McCarthy, J., 1994, Evidence for an upper mantle low velocity zone beneath the southern Basin and Range-Colorado Plateau transition zone: Geophys. Res. Lett., v. 21, p. 509-512.

Best, M. G., and Christiansen, E. H., 1991, Limited extension during peak Tertiary volcanism, Great Basin of Nevada and Utah: J. Geophys. Res., v. 96, p. 13,509-13,528.

Best, M. G., McKee, E. H., and Damon, P. E., 1980, Space-time-composition patterns of Late Cenozoic mafic volcanism, southwestern Utah and adjoining areas: Amer. J. Sci., v. 280, p. 1035-1050.

Best, M. G., Christiansen, E. H., and Blank, H. R., Jr., 1989, Oligocene caldera complex and calc-alkaline tuffs and lavas of the Indian Peak volcanic field, Nevada and Utah: Geol. Soc. Amer. Bull., v. 101, p. 1076-1090.

Blackwell, D. D., Steele, J. L., and Carter, L. S., 1991, Heat flow patterns of the North American continent: A discussion of the geothermal map of North America, in Slemmons, D. B., Engdahl, E. R., Zoback, M. D., and Blackwell, D. D., editors, Neotectonics of North America, p. 423-436, Geol. Soc. Amer., Boulder.

Bodell, J. M., and Chapman, D. S., 1982, Heat flow in the north-central Colorado Plateau: J. Geophys. Res., v. 87, p. 2869-2884.

Boutelier, R. R., and Keen, C. E., 1999, Small-scale convection and divergent plate boundaries: J. Geophys. Res., v. 104, p. 7389-7404.

Braile, L. W., Smith, R. B., Keller, G. R., and Welch, R. M., 1974, Crustal structure across the Wasatch front from detailed seismic refraction studies: J. Geophys. Res., v. 79, p. 2669-2677.

Brandon, A. D., and Draper, D. S., 1996, Constraints on the origin of the oxidation state of mantle overlying subduction zones: an example from Simcoe, Washington, USA: Geochim. Cosmochim. Acta., v. 60, p. 1739-1749.

Brun, J. P., and Beslie, M. O., 1996, Mantle exhumation at passive margins: Earth Planet. Sci. Lett., v. 142, p. 161-173.

Buck, W. R., 1991, Modes of continental lithospheric extension: J. Geophys. Res., v. 96, p. 20,161-20,178.

Buck, R. and Su, W., 1989, Focused mantle upwelling below mid-ocean ridges due to feedback between viscosity and melting: Geophys. Res. Lett., v. 16, p. 641-644.

Buck, W. R., Martinez, F., Steckler, M. S., and Cochran, J. R., 1988, Thermal consequences of lithospheric shear: pure and simple: Tectonics, v. 7, p. 213-234.

Burchfiel, B. C., Cowan, D. S., and Davis, G. A., 1992, Tectonic overview of the Cordilleran orogen in the western United States, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., Boulder, p. 407-480.

Burrett, C., and Berry, R., 2000, Proterozoic Australia-Western United States (AUSWUS) fit between Laurentia and Australia: Geology, v. 28, p. 103-106.

Carmichael, I. S. E., 1991, The redox states of basic and silicic magmas: a reflection of their source regions?: Contr. Mineral. Petr., v. 106, p. 129-141.

Catchings, R. D., and Mooney, W. D., 1988, Crustal structure of the Columbia Plateau: Evidence for continental rifting: J. Geophys. Res., v. 93, p. 459-474.

Chase, C. G., 2000, Geoid/topography data suggest shallow support of Colorado Plateau elevation: Geol. Soc. Amer. Abstr. Prog., Annual meeting, p. A-42.

Christiansen, N. I., and Mooney, W. D., 1995, Seismic velocity structure and composition of the continental crust: a global view: J. Geophys. Res., v. 100, p. 9761-9788.

Christiansen, R. L., and Yeats, R. S., 1992, Post-Laramide geology of the U. S. Cordilleran region, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., p. 261-406.

Coleman, D. S., and Walker, J. D., 1992, Evidence for the generation of juvenile granitic crust during continental extension, Mineral Mountains batholith, Utah: J. Geophys. Res., v. 97, p. 11,011-11,024.

Coleman, D. S. and J. D. Walker, Modes of tilting during extensional core complex development: Science, 263, 215-218, 1994.

Coleman, D. S., Bartley, J. M., Walker, J. D., Price, D. E., and Friedrich, A. M., 1997, Extensional Faulting, Footwall Deformation and Plutonism in the Mineral Mountains, Southern Sevier Desert, in Link, P., and Kowallis, B., editors, Mesozoic to Recent geology of Utah (Field Trip Guidebook, 1997 GSA Annual Meeting), Brigham Young Univ. Geol. Studies, v. 42, pt. 2, p. 203-233.

Coney, P. J., and Harms, T. A., 1984, Cordilleran metamorphic core complexes: Cenozoic extensional relics of Mesozoic compression: Geology, v. 12, p. 550-554.

Coogan, J. C., and DeCelles, P. G., 1996, Extensional collapse along the Sevier Desert reflection, northern Sevier Desert basin, western United States: Geology, v. 24, p. 933-936.

Coogan, J. C., DeCelles, P. G., Mitra, G., Sussman, A. J., 1995, New regional balanced cross section across the Sevier Desert region and the central Utah thrust belt: Geol. Soc. Amer. Abstr. Progs., v. 27, no. 4, p. 7.

Coogan, J. C., and DeCelles, P. G., 1998, Extensional collapse along the Sevier Desert reflection, northern Sevier Desert basin, western United States: Reply, Geology, v. 26, p. 475.

Cordery, M. J., and Phipps Morgan, J., 1993, Convection and melting at mid-ocean ridges: J. Geophys. Res., v. 98, p. 19-477-19,503.

Constenius, K. N., 1996, Late Paleogene extensional collapse of the Cordilleran foreland fold and thrust belt, Geol. Soc. Amer. Bull., v. 108, p. 20-39.

Davis, G. H., 1978, Monocline fold pattern of the Colorado Plateau, Mem. 151, Geol. Soc. Amer., 215-233.

DeLugao, P. P., and Wannamaker, P. E., 1996, Calculating the two-dimensional magnetotelluric Jacobian in finite elements using reciprocity: Geophys. J. Int., v. 127, p. 806-810.

DeMets, C., and Dixon, T. H., 1999, New kinematic models for Pacific-North American motion from 3 Ma to present, I: evidence for steady motion and biases in the NUVEL-1A model: Geophys. Res. Lett., v. 26, p. 1921-1924.

DePaolo, D. J., and Daley, E. E., 2000, Neodymium isotopes in basalts of the southwest basin and range and lithospheric thinning during continental extension: Chem. Geol., v. 169, p. 157-185.

Diehl, S. F., Anderson, R. E., Humphrey, J. D., and Barnhard, T. P., 1997, Hydrothermal vein deposits associated with a Tertiary low-angle normal fault, Salina, Utah: Amer. Assoc. Petr. Geol. Bull., v. 81, p. 1222.

Dixon, T., Farina, F., DeMets, C., Suarez-Vidal, F., Fletcher, J., Marquez-Azua, B., Miller, M., Sanchez, O., and Umhoefer, P., 2000, New kinematic models for Pacific-North American motion from 3 Ma to Present, II: evidence for a "Baja California shear zone": Geophys. Res. Lett., v. 27, p. 3961-3964.

Driscoll, N. W., and Karner, G. D., 1998, Lower crustal extension across the Northern Carnarvon basin, Australia: evidence for an eastward dipping detachment: J. Geophys. Res., v. 103, p. 4975-4991.

Ducea, M. N., and Saleeby, J. B., 1998, The age and origin of a thick mafic-ultramafic keel from beneath the Sierra Nevada batholith: Contr. Miner. Petr., v. 133, p. 169-185.

Dumitru, T. A., Gans, P. B., Foster, D. A., and Miller, E. L., 1991, Refrigeration of the western Cordeilleran lithosphere during Laramide shallow-angle subduction: Geology, v. 19, p. 1145-1148.

Dumitru, T. A., Duddy, I. R., and Green, P. F., 1994, Mesozoic-Cenozoic burial, uplift, and erosion history of the west-central Colorado Plateau: Geology, v. 22, p. 499-502.

Dumitru, T. A., Miller, E. L., Stockli, D. F., and Surpless, B. E., 1997, Fission track constraints on time-space patterns of Miocene extension in the northern Basin and Range Province, Geol. Soc. Amer. Abstr. Prog., v. 29, p. A-232.

Dumitru, T. A., 2000, Fission-track geochronology, in Noller, J. S., J. M. Sowers, and W. R. Lettis, editors, Quaternary Geochronology: Applications in Quaternary Geology and Paleoseismology, Amer. Geophys. Union Geophys. Monogr., in press.

Dumitru, T. A., Miller, E. L., Surpless, B. E., Martinez, C. M., Egger, A. E., and Stockli, D., 2000, Large structural domains of synchronous Miocene extension in the northern Basin and Range Province: Geol. Soc. Amer. Abstr. Prog., Annual meeting, p. A-43.

Ebinger, C. J., and Casey, M., 2001, Continental breakup in magmatic provinces: an Ethiopian example: Geology, v. 29, p. 527-530.

Edel, J. B., Fuchs, K., Gelbke, C., and Prodehl, C., 1975, Deep structure of the southern Rhinegraben area from seismic refraction investigations: J. Geophys., v. 41, p. 333-356.

Ehlers, T. A., Armstrong, P. A., and Chapman, D. S., 2001, Normal fault thermal regimes and the interpretation of low-temperature thermochronometers: Phys. Earth Planet. Inter., in press.

Falkner, C. M., Miller, C. F., Wooden, J. L., and Heizler, M. T., 1995, Petrogenesis and tectonic significance of the calc-alkaline, bimodal Aztec Wash pluton, Eldorado Mountains, Colorado River extensional corridor: J. Geophys. Res., v. 100, p. 10,453-10,476.

Farmer, G. L., and DePaolo, D. J., 1983, Origin of Mesozoic and Tertiary granites in the western U.S. and implications for pre-Mesozoic crustal structure, 1. Nd and Sr isotopic studies in the geocline of the northern Great Basin: J. Geophys. Res., v. 88, p. 3379-3401.

Farmer, G. L., and DePaolo, D. J., 1984, Origin of Mesozoic and Tertiary granites in the western U.S. and implications for pre-Mesozoic crustal structure, 2. Nd and Sr isotopic studies of unmineralized and Cu- and Mo-mineralized granites in the Precambrian craton: J. Geophys. Res., v. 89, p. 10,141-10,160.

Faul, U. H., Toomey, D. R., and Waff, H. S., 1994, Intergranular basaltic melt is distributed in thin, elongated inclusions: Geophys. Res. Lett., v. 21, p. 29-32.

Fenneman, N. M., and Johnson, D. W., 1946, Physical division of the United States, U. S. Geol. Surv., scale 1:7,000,000.

Fitton, J. G., James, D., and Leeman, W. P., 1991, Basic magmatism associated with Late Cenozoic extension in the western United States: compositional variations in space and time: J. Geophys. Res., v. 96, p. 13,693-13,711.

Flesch, L. M., Holt, W. E., Haines, A. J., and Shen-Tu, B., 1999, Dynamics of the Pacific-North American plate boundary in the western United States: Science, v. 287, p. 834-836.

Fliedner, M. M., Ruppert, S., and SSCD Working Group, 1996, Three-dimensional crustal structure of the southern Sierra Nevada from seismic fan profiles and gravity modeling: Geology, v. 24, p. 367-370.

Forest, C. E., Wolfe, J. A., Molnar, P., and Emanuel, K. A., 1999, Paleoaltimetry incorporating atmospheric physics and botanical estimates of paleoclimate: Geol. Soc. Amer. Bull., v. 111, p. 497-511.

Fouch, M. J., Fischer, K. M., Parmentier, E. M, Wysession, M. E., and Clarke, T. J., 2000, Shear wave splitting, continental keels, and patterns of mantle flow: J. Geophys. Res., v. 105, p. 6255-6275.

Froitzheim, N., and Rubatto, D., 1998, Continental breakup by detachment faulting: field evidence and geochronological constraints (Tasna nappe, Switzerland): Terra Nova, v. 10, p. 171-176.

Frost, B. R., and Bucher, K., 1994, Is water responsible for geophysical anomalies in the deep continental crust? A petrological perspective: Tectonophys., v. 231, p. 293-309.

Frost, B. R., Frost, C. D., and Touret, J. L. R., 1989, Magmas as a source of heat and fluids in granulite metamorphism, in Bridgwater, D., editor, Fluid movements - elemental transport and the composition of the deep crust, NATO ASI Series C, v. 281, p. 1-18.

Gajewski, D., Holbrook, W. S., and Prodehl, C., 1987, A three-dimensional crustal model of southwest Germany derived from seismic refraction data: Tectonophysics, v. 142, p. 49-70.

Gans, P. B., and Bohrson, W. A., 1998, Suppression of volcanism during rapid extension in the Basin and Range province, United States: Science, v. 279, p. 66-68.

Gao, S., Davis, P. M., Liu, H., Slack, P. D., Rigor, A. W., Zorin, Y. A., Mordvinova, V. V., Kozhevnikov, V. M., and Logatchev, N. A., 1997, SKS splitting beneath continental rift zones: J. Geophys. Res., v. 102, p. 22,781-22,797.

Gao, S., Davis, P. M., Liu, H., Slack, P. D., Rigor, A. W., Zorin, Y. A., Mordvinova, V. V., Kozhevnikov, V. M., and Logatchev, N. A., 1999, Reply to Vauchez et al.: J. Geophys. Res., v. 104, p. 10,791-10,794.

Gartrell, A. P., 2000, Rheological controls on extensional styles and the structural evolution of the Northern Carnarvon Basin, North West Shelf, Australia: Austral. J. Earth Sci., v. 47, p. 231-244.

Gastil, R. G., 1979, A conceptual hypothesis for the relation of differing tectonic terranes to plutonic emplacement: Geology, v. 7, p. 542-544.

GLOBE Task Team et al. (Hastings, D. A., Dunbar, P. K., Elphingstone, G. M., Bootz, M., Murakami, H., Maruyama, H., Masaharu, H., Holland, P., Payne, J., Bryant, N. A., Logan, T. L., Muller, J.-P., Schreier, G., and MacDonald, J. S.), eds., 1999. The Global Land One-kilometer Base Elevation (GLOBE) digital elevation model, version 1.0., National Oceanic and Atmospheric Administration, National Geophysical Data Center, Boulder, Colorado.

Goodwin, E. B., and McCarthy, J., 1990, Composition of the lower crust in west-central Arizona from three-component seismic data: J. Geophys. Res., v. 95, p. 20,097-20,109.

Gough, D. I., 1989, Magnetometer array studies, earth structure, and tectonic processes: Rev. Geophys. V. 27, p. 141-157.

Govers, R., and Wortel, M. J. R., 1993, Initiation of asymmetric extension in continental lithosphere: Tectonophys., v. 223, p. 75-96.

Govers, R., and Wortel, M. J. R., 1995, Extension of stable continental lithosphere and the initiation of lithospheric scale faults: Tectonics, v. 14, p. 1041-1055.

Gregory-Wodzicki, K. M., 1997, The late Eocene House Range Flora, Sevier Desert, Utah: Paleoclimate and paleoelevation: Pallaios, v. 12, p. 552-567.

Grunder, A. L., 1995, Material and thermal roles of basalt in crustal magmatism: case study from eastern Nevada: Geology, v. 23, p. 952-956.

Hall, D. B., 2001, Deep physical state of the western Colorado Plateau and its transition to the Great Basin from magnetotelluric resistivity data: M. S. thesis, University of Utah, 89 pp.

Hamilton, W. B., 1988, Laramide crustal shortening, in Schmidt, C. J., and Perry, W. J., Jr., editors, Interaction of the Rocky Mountain foreland and the Cordilleran thrust belt, Mem. 171, Geol. Soc. Amer., p. 27-39.

Harry, D. L., and Bowling, J., 1998, Extension and volcanism: insights from dynamic modeling studies, in Erzinger, J., editor, Proc. Intern. Lith. Prog. Worksh. "Volcanic Margins", Geoforschungszentrum, Potsdam, 53 pp.

Harry, D. L., and Bowling, J. C., 1999, Inhibiting magmatism on nonvolcanic rifted margins: Geology, v. 27, p. 895-898.

Harry, D. L., andLeeman, W. P., 1995, Partial melting of melt metasomatized subcontinental mantle and the magma source potential of the lower lithosphere: J. Geophys. Res., v. 100, p. 10,255-10,269.

Harry, D. L. and Sawyer, D. S., 1992, Basaltic volcanism, mantle plumes and the mechanics of rifting: the Parana flood basalt province of South America: Geology, v. 20, p. 207-210.

Harry, D. L., Sawyer, D. S., and Leeman, W. P., 1993, The mechanics of continental extension in western North America: implications for the magmatic and structural evolution of the Great Basin ; Earth Planet. Sci. Lett., v. 117, p. 59-71.

Hauser, E. C., Gephart, J., Latham, T., Brown, L., Kaufman, S., Oliver, J., and Lucchitta, I., 1987, COCORP Arizona transect: strong crustal reflections and offset Moho beneath the transition zone: Geology, v. 15, p. 1103-1106.

Hawkesworth, C., Turner, S., Gallagher, K., Hunter, A., Bradshaw, T., and Rogers, N., 1995, Calc-alkaline magmatism, lithospheric thinning and extension in the Basin and Range: J. Geophys. Res., v. 100, p. 10,271-10,286.

Hecker, S., 1993, Quaternary tectonics of Utah with emphasis on earthquake-hazard characterization, Utah Geol. Surv. Bull.

Henrikson, A., 2000, New heat flow determinations from oil and gas wells in the Colorado Plateau and Basin and Range of Utah: M.S. thesis, University of Utah.

Henyey, T. L., Okaya, D. A., Frost, E. G., and McEvilly, T. V., 1987, CALCRUST (1985) seismic reflection survey, Whipple Mountains detachment terrane, California: an overview: Geophys. J. R. Astr. Soc, v. 89, p. 111-118.

Hintze, L. F., 1980, Geologic map of Utah, Utah Geol. Mineral. Surv., scale 1:500,000.

Hintze, L. F., 1988, Geologic history of Utah, Brigham Young Univ. Geol. Studies, Spec. Pub. 7, 202 p.

Holbrook, W. S., Purdy, G. M., Sheridan, R. E., Glover, L., III, Talwani, M., Ewing, J., and Hutchinson, D., 1994a, Seismic structure of the U.S. Mid-Atlantic continental margin: J. Geophys. Res., v. 99, p. 17,871-17,891.

Holbrook, W. S., Reiter, E. C., Purdy, G. M., Sawyer, D., Stoffa, P. L., Austin, J. A., Jr., Oh, J., and Makris, J., 1994b, Deep structure of the U. S. Atlantic continental margin, offshore South Carolina, from coincident ocean bottom and multichannel seismic data: J. Geophys. Res., v.99, p. 9155-9178.

Hopper, J. R., and Buck, W. R., 1998, Styles of extensional decoupling: Geology, v. 26, p. 699-702.

Hose, R. K., 1977, Structural geology of the Confusion Range, west-central Utah: U.S.Geol.Surv. Prof. Paper 971, 9 p.

Huismans, R. S., Podlakchikov, Y. Y., and Cloetingh, S., 2001, Transition from passive to active rifting: relative importance of asthenospheric doming and passive extension of the lithosphere: J. Geophys. Res., v. 106, p. 11,271-11,291.

Humphreys, E. D., 1995, Post-Laramide removal of the Farallon slab, western United States: Geology, v. 23, p. 987-990.

Humphreys, E. D., and Dueker, K. G., 1994, Physical state of the western U. S. mantle: J. Geophys. Res., v. 99, p. 9635-9650.

Jones, A. G., 1999, Imaging the continental upper mantle using electromagnetic methods: Lithosv. 48, p. 570-580.

Jones, C. H., 1987, A geophysical and geological investigation of extensional structures, Great Basin, western United

States: Ph. D. thesis, Mass. Inst. Technol., Cambridge, 226 pp.

Jones, C. H., and Phinney, R. A., 1998, Seismic structure of the lithosphere from teleseismic converted arrivals observed at small arrays in the southern Sierra Nevada and vicinity, California: J. Geophys. Res., v. 103, p. 10,065-10,090.

Jones, C. H., Wernicke, B. P., Farmer, G. L., Walker, J. D., and Coleman, D. S., 1992, Variations across and along a major continental rift: an interdisciplinary study of the Basin and Range province, western U.S.A.: Tectonophys., v. 213, p. 57-96.

Jones, C. H., Unruh, J. R., and Sonder, L. J., 1996, The role of gravitational potential energy in active deformation in the southwestern United States: Nature, v. 381, p. 37-41.

Jones, C. H., Sonder, L. J., and Unruh, J. R., 1998, Lithospheric gravitational potential energy and past orogenesis: implications for conditions of initial Basin and Range and Laramide deformation: Geology, v. 26, p. 639-642.

Julian, B. R., 1970, Regional variations in upper mantle structure beneath North America: Ph.D. thesis, Calif. Inst. Tech., 208 pp.

Karato, S., 1990, The role of hydrogen in the electical conductivity of the upper mantle: Nature, v. 347, p. 272-273.

Karato, S.I., 1998, A dislocation model of seismic wave attenuation and micro-creep in the Earth: Pure Appl. Geoph., v. 153, p. 239-256.

Karato, S., and H. Jung, 1998, Water, partial melting and the origin of the seismic low velocity and high attenuation zone in the upper mantle: Earth Planet. Sci. Lett., v. 157, p. 193-207.

Karato, S., and P. Wu, 1993, Rheology of the upper mantle: a synthesis: Science, v. 260, p. 771-778.

Karlstrom, K. E., and Williams, M. L., 1998, Hetergeneity of the middle crust: implications for strength of continental lithosphere: Geology, v. 26, p. 815-819.

Karlstrom, K. E., Harlan, S. S., Williams, M. L., McLelland, J., Geissman, J. W., and Ahall, K.-I., 1999, Redefining Rodinia: geologic evidence for the Australia-western U.S. connection in the Proterozoic: GSA Today, v. 9, no. 10, p. 1-7.

Kelemen, P. B., and Holbrook, W. S., 1995, Origin of thick, high-velocity igneous crust along the U.S. east coast margin: J. Geophys. Res., v. 100, p. 10,077-10,094.

Kempton, P. D., Fitton, J. G., Hawkesworth, C. J., and Ormerod, D. S., 1991, Isotopic and trace element constraints on the composition and evolution of the lithosphere beneath the southwestern United States: J. Geophys. Res., v. 96, p. 13,713-13,735.

Ketcham, R. A., 1996, Thermal models of core-complex evolution in Arizona and New Guinea: Implications for ancient cooling paths and present-day heat flow: Tectonics, v. 15, p. 933-951.

Kincaid, C., Sparks, D. W., and Detrick, R., 1996, The relative importance of plate-driven and buoyancy-driven flow at mid-ocean ridges: J. Geophys. Res., v. 101, p. 16,177-16,193.

Kohlstedt, D. L., Evans, B., and Mackwell, S. J., 1995, Strength of the lithosphere: constraints imposed by laboratory experiments: J. Geophys. Res., v. 100, p. 17,587-17,602.

Kowallis, B. J., Ferguson, J., and Jorgensen, G. J., 1990, Uplift along the Salt Lake segment of the Wasatch fault from apatite and zircon fission track dating in the Little Cottonwood stock: Nuclear Tracks, v. 17, p. 325-329.

Lachenbruch, A. H., and Sass, J. H., 1978, Models of an extending lithosphere and heat flow in the Basin and Range province, in Smith, R. B., and Eaton, G. P., editors, Cenozoic tectonic and regional geophysics of the western Cordillera, Mem. 152, Geol. Soc. Amer., p. 209-250.

Langmuir, C. H., Klein, E. M., and Plank, T., 1992, Petrological constraints on melt formation and migration beneath mid-ocean ridges, in Phipps Morgan, J., Blackman, D. and Sinton, J. L., editors, Mantle Flow and Melt Generation at Mid-Ocean Ridges, Geophysical Monograph 71, Amer. Geophys. Union, p. 183-280.

Lastowka, L. A., Sheehan, A. F., and Schneider, J. N., 2001, Seismic evidence for partial lithospheric delamination model of Colorado Plateau uplift: Geophys. Res. Lett., v. 28, p. 1319-1322.

Lee, J., 1995, Rapid uplift and rotation of mylonitic rocks from beneath a detachment fault: Insights from potassium feldspar $^{40}Ar/^{39}Ar$ thermochronology, northern Snake range, Nevada: Tectonics, v. 14, p. 54-77.

Lewis, C. J., Wernicke, B. P., Selverstone, J., and Bartley, J. M., 1999, Deep burial of the footwall of the northern Snake Range decollement, Nevada: Geol. Soc. Amer. Bull., v. 111, p. 39-51.

Libarkin, J. C., and Sussman, A. J., 2000, Evaluation of proposed mechanisms for Colorado Plateau uplift: constraints on pre-orogenic crust and mantle lithospheric thickness: Geol. Soc. Amer. Abstr. Prog., Annual meeting, p. A-42.

Lizarralde, D., and Holbrook, W. S., 1997, U. S. mid-Atlantic margin structure and early thermal evolution: J. Geophys. Res., v. 102, p. 22,855-22,875.

Loeb, D. T., and Pechmann, J. C., 1986, The P-wave velocity structure of the crust-mantle boundary beneath Utah from network travel-time measurements (abstract): Earthquake Notes, v. 57, p. 10-11.

Lowry, A. R., and Smith, R. B., 1995, Strength and rheology of the western U. S. Cordillera: J. Geophys. Res., v. 100, p. 17,947-17,963.

Lowry, A. R., Ribe, N. M., and Smith, R. B., 2000, Dynamic elevation of the Cordillera, western United States: J. Geophys. Res., v. 105, p. 23,371-23,390.

Lynch, D. P., 1999, Three-dimensional finite difference tomography of the Basin and Range-Colorado Plateau-Rocky Mountain transition using earthquake and controlled source data: M.S. thesis, University of Utah, 155 p.

Mattox, S. R., 1997, Geochemistry, source characteristics, and tectonic implications of Upper Cenozoic basalts of the Basin and Range-Colorado Plateau Transition Zone, Utah: U. S. Geol. Surv. Bull. 2153-J.

McCarthy, J., and Parsons, T., 1994, Insights into the kinematic Cenozoic evolution of the Basin and Range-Colorado Plateau transition from coincident seismic refraction and reflection data: Geol. Soc. Amer. Bull., v. 106, p. 747-759.

McCarthy, J., Larkin, S. P., Fuis, G. S., Simpson, R. W., and Howard, K. A., 1991, Anatomy of a metamorphic core complex: Seismic refraction/wide-angle reflection profiling in southeastern California and western Arizona: J. Geophys. Res., v. 96, p. 12,259-12,291.

McKenzie, D. P., 1978, Some remarks on the development of sedimentary basins: Earth Planet. Sci. Lett., v. 40, p. 25-32.

McKenzie, D., and Bickle, M. J., 1988, The volume and composition of melt generated by extension of the lithosphere: J. Petrology, v. 29, p. 625-679.

McQuarrie, N., and Chase, C., 2000, Raising the Colorado Plateau: Geology, v. 28, p. 91-94.

Mechie, J., Keller, G. R., Prodehl, C., Khan, M. A., and Gaciri, S. J., 1997, A model for the structure, composition and evolution of the Kenya Rift: Tectonophys., v. 278, p. 95-119.

Meissner, R., 1973, The 'Moho' as a transition zone: Geophys. Surv., v. 1, p. 195-216.

Miller, E. L., Gans, P. B., and Garing, J., 1983, The Snake Range decollement: An exhumed mid-Tertiary brittle--ductile transition: Tectonics, v. 2, p. 239-263.

Miller, D. M., Nilsen, T. H., and Bilodeau, W. L., 1992, Late Cretaceous to early Eocene geologic evolution of the U. S. Cordillera, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., Boulder, p. 205-260.

Miller, E. L., Dumitru, T. A., Brown, R. W., and Gans, P. B., 1999, Rapid Miocene slip on the Snake Range-Deep Creek Range fault system, east-central Nevada: Geol. Soc. Amer. Bull., v. 111, p. 886-905.

Mooney, W. D., Andrews, M. C., Ginzburg, A., Peters, D. A., and Hamilton, R. M., 1983, Crustal structure of the northern Mississippi embayment and a comparison with other continental rift zones: Tectonophys., v. 94, p. 327-348.

Nealy, L. D., Budahn, J. R., Maldonado, F., and Unruh, D. M., 1997, Geochemistry and petrogenesis of Quaternary basaltic rocks from the Red Hills and western Markagunt Plateau, southwestern Utah: U. S. Geol. Surv. Bull., 2153-I.

Nelson, S. T., and Davidson, J. P., 1993, Interactions between mantle-derived magmas and mafic crust, Henry Mountains, Utah: J. Geophys. Res., v. 98, p. 1837-1852.

Nelson, S. T., Davidson, J. P., and Sullivan, K. R., 1992, New age determinations of central Colorado Plateau laccoliths, Utah: recognizing disturbed K-Ar systematics and re-evaluating tectonomagmatic relationships: Geol. Soc. Amer. Bull., v. 104, p. 1547-1560.

Nelson, S. T., and Tingey, D. G., 1997, Time-transgressive and extension-related basaltic volcanism in southwest Utah and vicinity: Geol. Soc. Amer. Bull., v. 109, p. 1249-1265.

Nielson, D. L., Evans, S. H., Jr. and Sibbett, B. S., 1986, Magmatic, structural, and hydrothermal evolution of the Mineral Mountains intrusive complex, Utah: Geol. Soc. Amer. Bull., v. 97, p. 756-777.

Nusbaum, R. L., Unruh, D. M., and Millings, V. E., III, 1997, The role of lithosphere and asthenosphere in the genesis of Late Cenozoic volcanism at Diamond Valley and Vey volcano, southwestern Utah: U. S. Geol. Surv. Bull. 2153-K.

Otton, J. K., 1995, Western frontal fault of the Canyon Range: Is it the breakaway zone of the Sevier Desert detachment?: Geology, v.23, p. 547-550.

Ozalaybey, S., Savage, M. K., Sheehan, A. F., Louie, J. N., and Brune, J. N., 1997, Shear-wave velocity structure in the northern Basin and range province from the combined analysis of receiver functions and surface waves: Bull. Seis. Soc. Amer., v. 87, p. 183-199.

Pakiser, L. C., 1989, Geophysics of the intermontane system, in Pakiser, L. C., and Mooney, W. D., editors, Geophysical framework of the continental United States, Mem. 172, Geol. Soc. Amer., p. 235-247.

Park, S. K., Hirasuna, B., Jiracek, G. R., and Kinn, C., 1996, Magnetotelluric evidence of lithospheric mantle thinning beneath the southern Sierra Nevada: J. Geophys. Res., v. 101, p. 16,241-16,255.

Parkinson, I. J., and R. J. Arculus, The redox state of subduction zones: insights from arc-peridotites, Chem. Geol., 160, 409-423, 1999.

Parsons, T., 1995, The Basin and Range Province, in Olsen, K. H., editor, Continental rifts: evolution, structure, tectonics, Dev. Geotect., v. 25, Elsevier, p. 277-324.

Parsons, T., Howie, J. M., and Thompson, G. A., 1992, Seismic constraints on the nature of lower crustal reflectors beneath the extending southern Transition Zone of the Colorado Plateau, Arizona: J. Geophys. Res., v. 97, p. 12,391-12,407.

Paterson, S. R., and Fowler, T. K., Jr., 1993, Extensional pluton-emplacement models: do they work for large plutonic complexes: Geology, v.21, p. 781-784.

Patino Douce, A. E., Humphyreys, E. D., and Johnson, A. D., 1990, Anatexis and metamorphism in tectonically thickened continental crust exemplified by the Sevier hinterland, western North America: Earth Planet. Sci. Lett., v. 97, p. 290-315.

Perry, F. V., DePaolo, D. J., and Baldridge, W. S, 1993., Neodymium isotopic evidence for decreasing crustal contributions to Cenozoic ignimbrites of the western United States: implications for the thermal evolution of the Cordilleran crust: Geol. Soc. Amer. Bull., v. 105, p. 872-882.

Planke, S., and Smith, R. B., 1991, Cenozoic extension and evolution of the Sevier Desert basin, Utah, from seismic reflection, gravity, and well log data: Tectonics, v. 10, p. 345-365.

Poirier, J. P., 1985, Creep of crystals: High-temperature deformation processes in metals, ceramics and minerals, Cambridge University Press, New York.

Pollack, H. N., Hurter, S. J., and Johnson, J., 1993, Heat flow from the Earth's interior: analysis of the global data set: Rev. Geophys., v. 31, p. 267-280.

Poole, F. G., Stewart, J. H., Palmer, A. R., Sandberg, C. A., Madrid, R. J., Ross, R. J., Jr., Hintze, L. F., Miller, M. M., and Wrucke, C. T., 1992, Latest Precambrian to latest Devonian time; development of a continental margin, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., p. 9-56.

Porath, H., 1971, Magnetic variation anomalies and seismic low-velocity zone in the western United States: J. Geophys. Res., v. 76, p. 2643-2648.

Porath, H., Oldenburg, D. W., and Gough, D. I., 1970, Separation of magnetic variation fields and conductive structures in the western United States: Geophys. J. Roy. Astr. Soc., v. 19, p. 237-260.

Powell, W. G., 1997, Thermal state of the lithosphere in the Colorado Plateau-Basin and Range transition zone, Utah: Ph.D. thesis, University of Utah, 232 pp.

Prodehl, C., and Lipman, P. W., 1989, Crustal structure of the Rocky Mountain region, in Pakiser, L. C., and Mooney, W. D., editors, Geophysical framework of the continental United States, Mem. 172, Geol. Soc. Amer., p. 249-284.

Reynolds, S., and Spencer, J., 1985, Evidence for large-scale transport on the Bullard detachment fault, west-central Arizona: Geology, v. 13, p. 353-356.

Riciputi, L. R., Johnson, C. M., Sawyer, D. A., and Lipman, P. W., 1995, Crustal and magmatic evolution in a large multicyclic caldera complex: isotopic evidence from the central San Juan volcanic field: J. Volc. Geoth. Res., v. 67, p. 1-28.

Roberts, J. J., and Tyburczy, J. A., 1999, Partial-melt electrical conductivity: influence of melt composition: J. Geophys. Res., v. 104, p. 7055-7065.

Rowley, P. D., 1998, Cenozoic transverse zones and igneous belts in the Great Basin, Western United States; their tectonic and economic implications, in Faulds, J. E., and Stewart, J. H., editors, Accommodation zones and transfer zones; the regional segmentation of the Basin and Range Province, Spec. Pap. 323, Geol. Soc. Amer., p. 195-228.

Rowley, P. D., Anderson, J. J., Williams, P. L., and Fleck, R. J., 1978, Age of structural differentiation between the Colorado Plateaus and Basin and Range provinces, southwestern Utah: Geology, v. 6, p. 51-55.

Rowley, P. D., Steven, T. A., Anderson, J. J., and Cunningham, C. G., 1979, Cenozoic stratigraphic and structural framework of southwestern Utah: U. S. Geol. Surv. Prof. Pap. 1149, 22 pp.

Rowley, P. D., Steven, T. A., and Mehnert, H. H., 1981, Origin and structural implications of upper Miocene rhyolites in Kingston Canyon, Piute County, Utah: Geol. Soc. Amer., v. 92, p. 590-602.

Rowley, P. D., Cunningham, C. G., Steven, T. A., Mehnert, H. H., and Naeser, C. W., 1998, Cenozoic igneous and tectonic setting of the Marysvale volcanic field and its relation to other igneous centers in Utah and Nevada, in Friedman, J. D. and Huffman, A. C. J., editors, Laccolith complexes of southeastern Utah; time of emplacement and tectonic setting; workshop proceedings: U. S. Geol. Surv. Bull. 2158, 167-201.

Royse, F., Jr., 1993, Case of the phantom foredeep: Early Cretaceous in west-central Utah: Geology, v. 21, p. 133-136.

Rudnick, R. L., and Fountain, D. M., 1995, Nature and composition of the continental crust: a lower crustal perspective: Rev. Geophys., v. 33, p. 267-309.

Ruppel, C., 1995, Extensional processes in continental lithosphere: J. Geophys. Res., v. 100, p. 24,187-24,215.

Ruppert, S., Fliedner, M. M., and Zandt, G., 1998, Thin crust and active upper mantle beneath Southern Sierra Nevada: Tectonophys., v. 286, p. 237-252.

Saltus, R. W., and Lachenbruch, A. H., 1991, Thermal evolution of the Sierra Nevada: Tectonic implications of new heat flow data: Tectonics, v. 10, p. 325-344.

Savage, M. K., and Sheehan, A. F., 2000, Seismic anisotropy and mantle flow from the Great Basin to the Great Plains, western United States: J. Geophys. Res., v. 105, p. 13,715-13,734.

Schutt, D. L., and Humphreys, E. D., 2001, Evidence for a deep asthenosphere beneath North America from western United States SKS splits: Geology, v. 29, p. 291-294.

Sheehan, A. F., Jones, C. H., Savage, M. K., Ozalaybey, S., and Schneider, J. M., 1997, Contrasting lithospheric structure between the Colorado Plateau and Great Basin: initial results from Colorado Plateau-Great Basin PASSCAL experiment: Geophys. Res. Lett., v. 24, p. 2609-2612.

Shuey, R. T., Schellinger, D. K., Johnson, E. H., and Alley, L. B., 1973, Aeromagnetics and the transition between the Colorado Plateaus and Basin and Range provinces: Geology, v. 1, p. 107-110.

Shuey, R. T., Schellinger, D. K., Tripp, A. C., and Alley, L. B., 1977, Curie depth determination from aeromagnetic spectra: Geophys. J. Roy. Astr. Soc., v. 50, p. 75-101.

Smith, E. I., Sanchez, A., Walker, J. D., and Wang, K., 1999, Geochemistry of mafic magmas in the Hurrican Volcanic Field, Utah: implications for small- and large-scale chemical variability of the lithospheric mantle: J. Geol., v. 107, p. 433-448.

Smith, R. B., and Bruhn, R. L., 1984, Intraplate extensional tectonics of the eastern Basin-Range: Inferences on structural style from seismic reflection data, regional tectonics, and thermal-mechanical models of brittle-ductile deformation: J. Geophys. Res., v. 89, p. 5733-5762.

Smith, R. B., Nagy, W. C., Julander, K. A. S., Viveiros, J. J., Barker, C. A., and Gants, D. J., 1989, Geophysical and tectonic framework of the eastern Basin and Range-Colorado Plateau-Rocky Mountain transition, in Pakiser, L. C., and Mooney, W. D., editors, Geophysical framework of the continental United States, Mem. 172, Geol. Soc. Amer., p. 205-233.

Sonder, L. J., andJones, C. H., 1999, Western United States: how the west was widened: Annu. Rev. Earth Planet. Sci., v. 27, p. 417-462.

Spencer, J., and Reynolds, S., 1990, Relationship between Mesozoic and Cenozoic tectonic features in west-central Arizona and adjacent southeastern California: J. Geophys. Res., v. 95, p. 539-555.

Standlee, L. A., 1982, Structure and stratigraphy of Jurassic rocks in central Utah - their influence on tectonic development of the Cordilleran foreland thrust belt, in Powers, R. B., editor, Geologic studies of the Cordilleran thrust belt, Rocky Mtn. Assoc. Geol., v. 1, p. 357-382.

Stern, T., Molnar, P., Okaya, D., and Eberhart-Phillips, D., 2000, Teleseismic P wave delays and modes of shortening the mantle lithosphere beneath South Island, New Zealand: J. Geophys. Res., v. 105, p. 21,615-21,631.

Stewart, J. H., 1980, Geology of Nevada, Nevada Bur. Mines Geol., Spec. Publ. 4, 136 pp.

Stewart, J. H., Moore, W. J., and Zeitz, I., 1977, East-west patterns of Cenozoic igneous rocks, aeromagnetic anomalies, and mineral deposits, Nevada and Utah: Geol. Soc. Amer. Bull., v. 88, p. 67-77.

Stewart, M. E., and Taylor, W. J., 1996, Structural analysis and fault segment boundary identification along the Hurricane fault in southwestern Utah: J. Struct. Geol., v. 18, p. 1017-1029.

Stockli, D. F., 1999, Regional timing and spatial distribution of Miocene extension in the northern Basin and Range province: Ph.D. thesis, Stanford University, 239 p.

Stockli, D. F., Dumitru, T. A., Miller, E. L., and Surpless, B. E., 1999, Low-temperature thermochronological constraints on Neogene and younger tectonics in the northern Basin and Range province: Geol. Soc. Am. Abstr. Prog., v. 31, p. A-57.

Stockli, D.F., Linn, J.K., Walker, J.D., and Dumitru, T.A., 2001, Miocene unroofing of the Canyon Range during extension along the Sevier Desert Detachment, west-central Utah: Tectonics, in press.

Stuwe, K., White, L., and Brown, R., 1994, The influence of eroding topography on steady-state isotherms: Applications to fission track analysis: Earth Planet. Sci. Lett., v. 124, p. 63-74.

Tarantola, A., 1987, Inverse problem theory, Elsevier, New York, 671 pp.

Thatcher, W., Foulger, G. R., Julian, B. R., Svarc, J., Quilty, E., and Bawden, G. W., 1999, Present-day deformation across the Basin and Range province, western United States: Science, v. 283, p. 1714-1718.

Thompson, G. A., Catchings, R., Goodwin, E., Holbrook, S., Jarchow, C., Mann, C., McCarthy, J., and Okaya, D., 1989, Geophysics of the western Basin and Range province, in Pakiser, L. C., and Mooney, W. D., editors, Geophysical framework of the continental United States, Mem. 172, Geol. Soc. Amer., p. 177-203.

Tingey, D. G., Christiansen, E. H., Best, M. G., Ruiz, J., and Lux, D. R., 1991, Tertiary minette and melanepheline dikes, Wasatch Plateau, Utah: records of mantle heterogeneities and changing tectonics: J. Geophys. Res., v. 96, p. 13,529-13,544.

Tullis, J., 1990, Experimental studies of deformation mechanisms and microstructures in quartzo-feldspathic rocks, in Deformation processes in minerals, ceramics and rocks, Barker, D. J. and Meredith, P. G., editors, Unwin Hyman, London, p. 190-227.

Tullis, J., and Yund, R. A., 1991, Diffusion creep in feldspar aggregates: experimental evidence: J. Struct. Geol., v. 13, p. 987-1000.

Tullis, J., Yund, R. A., and Farver, J., 1996, Deformation-enhanced fluid distribution in feldspar aggregates and implications for ductile shear zones: Geology, v. 24, p. 63-66.

Turcotte, D. L., and Schubert, G., 1982, Geodynamics: applications of continuum physics to geological problems: John Wiley & Sons, New York, 450 pp.

van der Beek, P., Andriessen, P., and Cloetingh, S., 1995, Morphotectonic evolution of rifted continental margins: Inferences from a coupled tectonic-surface processes model and fission track thermochronology: Tectonics, v. 14, p. 406-421.

Van Schmus, W. R., Bickford, M. E., Anderson, J. L., Bender, E. E., Anderson, R. R., Bauer, P. W., Robertson, J. M., Bowring, S. A., Condie, K. C., Denison, R. E., Gilbert, M. C., Grambling, J. A., Mawer, C. K., Shearer, C. K., Hinze, W. J., Karstrom, K. E., Kisvarsanyi, E. B., Lidiak, E. G., Reed, J. C., Jr., Sims, P. K., Tweto, O., Silver, L. T., Treves, S. B., Williams, M. L., and Wooden, J. L., 1993, Transcontinental Proterozoic provinces, in Reed, J. C., Jr., Bickford, M. E., Houston, R. S., Link, P. K., Rankin, D. W., Sims, P. K., and Van Schmus, W. R., editors, Precambrian: Coterminous U. S., The Geology of North America, v. C-2, Geol. Soc. Amer., p. 171-334.

Vauchez, A., Barruol, G., and Nicolas, A., 1999, Comment on "SKS splitting beneath continental rifts zones" by Gao et al.: J. Geophys. Res., v. 104, p. 10,787-10,789.

Vauchez, A., Tommasi, A., Barruol, G., and Maumus, J., 2000, Upper mantle deformation and seismic anisotropy in continental rifts: Phys. Chem. Earth, v. 25, p. 111-117.

Vogel, T. A., Cambray, F. W., Feher, L. A., Constenius, K. N., and the WIB Research Team, 1998, Petrochemistry and emplacement history of the Wasatch Igneous Belt, Utah, in John, D. A., and Ballantyne, G. H., editors, Geology and ore deposits of the Oquirrh and Wasatch Mountains, Utah: Soc. Econ. Geol. Guideb. 29, p. 35-46.

Von Tish, D. B., Allmendinger, R. W., and Sharp, J. W., 1985, History of Cenozoic extension in central Sevier Desert, west-central Utah, from COCORP seismic reflection data: Amer. Assoc. Petr. Geol. Bull., v. 69, p. 1077-1087.

Vozoff, K., 1991, The magnetotelluric method, in M. N. Nabighian, editor, Electromagnetic methods in applied geophysics, v. 2B, Soc. Explor. Geophys., Tulsa, p. 641-711.

Wannamaker, P. E., 1999, Affordable magnetotellurics: interpretation in natural environments, in Oristaglio, M., and Spies, B., editors, Three-dimensional electromagnetics, Geophys. Devel. Ser., no. 7, Soc. Explor. Geophys., Tulsa, Okla., p. 349-374.

Wannamaker, P. E., 2000, Comment on "The petrologic case for a dry lower crust", by B. D. Yardley and J. W. Valley: J. Geophys. Res., v. 105, p. 6057-6064.

Wannamaker, P. E. and Hohmann, G. W., 1991, Electromagnetic induction studies, U.S. National Rept to the IUGG: Rev. Geophys., Supplement, p. 405-415.

Wannamaker, P. E., Stodt, J. A., and Rijo, L., 1987, A stable finite element solution for two-dimensional magnetotelluric modeling: Geophy. J. Roy. Astr. Soc., v. 88, p. 277-296.

Wannamaker, P. E., Johnston, J. M., Stodt, J. A. and Booker, J. R., 1997a, Anatomy of the Southern Cordilleran Hingeline, Utah and Nevada, from deep resistivity profiling: Geophysics, v. 62, p. 1069-1086.

Wannamaker, P. E., Doerner, W. M., Stodt, J. A., and Johnston, J. M., 1997b, Subdued state of tectonism of the Great Basin interior relative to its eastern margin based on deep resistivity structure: Earth Planet. Sci. Lett., v. 150, p. 41-53.

Wannamaker, P. E., Hulen, J. B., and Heizler, M. T., 2000, Early Miocene lamproite from the Colorado Plateau tectonic province, Utah: J. Volc. Geotherm. Res., v. 96, p. 176-191.

Wernicke, B. P., 1985, Uniform sense normal simple shear of the continental lithosphere: Can. J. Earth Sci., v. 22, p. 108-125.

Wernicke, B., 1992, Cenozoic extensional tectonics of the U. S. Cordillera, in Burchfiel, B. C., Lipman, P. W., and Zoback, M. L., editors, The Cordilleran orogen: coterminous U. S.: The Geology of North America, v. G-3, Geol. Soc. Amer., p. 553-581.

Wernicke, B., and Snow, J. K., 1998, Cenozoic tectonism in the central Basin and Range: motion of the Sierran-Great Valley block: Internat. Geol. Rev., v. 40, p. 403-410.

Wernicke, B., Clayton, R., Ducea, M., Jones, C. H., Park, S., Ruppert, S., Saleeby, J., Snow, J. K., Squires, L., Fliedner, M., Jiracek, G., Keler, R., Klemperer, S., Luetgert, J., Malin, P., Miller, K., Mooney, W., Oliver, H., and Phinney, R., 1996, Origin of high mountains in the continents: the Southern Sierra Nevada: Science, v. 271, p. 190-193.

Wernicke, B., Friedrich, A. M., Niemi, N. A., Bennett, R. A., and Davis, J. L., 2000, Dynamics of plate boundary fault systems from Basin and Range Geodetic Network (BARGEN) and geologic data: GSA Today, v. 10, no. 11, p. 1-7.

Westaway, R., 1999, The mechanical feasibility of low-angle normal faulting: Tectonophys., v. 308, p. 407-443.

White, R., and McKenzie, D., 1989, Magmatism at rift zones: The generation of volcanic continental margins and flood basalts: J. Geophys. Res., v. 94, p. 7685-7729.

Willis, G. C., 1986, Geologic map of the Salina Quadrangle, Sevier County, Utah: Utah Geol. Min. Surv. Map Ser., no. 83, scale 1:24,000.

Wills, S., and Anders, M. H., 1999, Tertiary normal faulting in the Canyon Range, eastern Sevier Desert: J. Geol., v. 107, p. 659-681.

Wills, S., Anders, M. H., and Otton, J. K., 1996, Western frontal fault of the Canyon Range: Is it the breakaway of the Sevier Desert detachment?, Reply: Geology, v. 24, p. 667-669.

Witkind, I. J., 1981, Reconnaissance geologic map of the Redmond Quadrangle, Sanpete and Sevier Counties, Utah: U. S. Geol. Surv., Map I-1304-A, scale 1:24,000.

Wolfe, J. A., Schorn, H. E., Forest, C. E., and Molnar, P., 1997, Paleobotanical evidence for high altitudes in Nevada during the Miocene: Science, v. 276, p. 1672-1675.

Wolfe, J. A., Forest, C. E., and Molnar, P., 1998, Paleobotanical evidence of Eocene and Oligocene paleoaltitudes in midlatitude western North America: Geol. Soc. Amer. Bull., v. 110, p. 664-678.

Wong, I. G., and Chapman, D. S., 1990, Deep intraplate earthquakes in the western United States and their relationship to lithospheric temperatures: Bull. Seis. Soc. Amer., v. 80, p. 589-599.

Wong, I. G., and Humphrey, J. R., 1989, Contemporary seismicity, faulting, and the state of stress in the Colorado Plateau: Geol. Soc. Amer. Bull., v. 101, p. 1127-1146.

Zandt, G., and Owens, T. J, 1980, Crustal flexure associated with normal faulting and implications for seismicity along the Wasatch front, Utah: Bull. Seismol. Soc. Am., v. 70, p. 1501-1520.

Zandt, G., Meyers, S. C., and Wallace, T. C, 1995, Crust and mantle structure across the Basin and Range-Colorado Plateau boundary at 37^{o}N latitude and implications for Cenozoic extensional mechanism: J. Geophys. Res., v. 100, p. 10,529-10,548.

Zoback, M. L., and Anderson, R. E., 1983, Style of Basin-Range faulting as inferred from seismic reflection data in the Great Basin, Nevada and Utah, in The role of heat in the development of energy and mineral resources in the Northern Basin and Range province, Geotherm. Resour. Counc. Spec. Rept. 13, p. 363-381.

Zoback, M. L., McKee, E. H., Blakely, R. J., and Thompson, G. A., 1994, The northern Nevada rift: regional tectonomagmatic relations and middle Miocene stress direction: Geol. Soc. Amer. Bull., v. 106, p. 371-382.

Zucca, J. J., 1984, The crustal structure of the southern Rhinegraben from re-interpretation of seismic refraction data: J. Geophys., v. 55, p. 13-22.