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Mass transfer along the Yellowstone hotspot track II: Kinematic constraints on the volume of mantle-derived magma

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ABSTRACT

This paper evaluates the Neogene kinematic evolution of the Eastern Snake River Plain (ESRP) to estimate the volume of mafic magma transferred from mantle to crust during passage over the Yellowstone hotspot. Four major kinematic processes affected the ESRP including densification due to emplacement of gabbro into the middle crust, subsidence and consequent filling of a 6-km-deep basin with volcanic and plutonic rocks, 20% extension via dike injection or faulting/ductile attenuation, and lower crustal flow from ESRP to the adjacent Basin and Range. Two kinematic models involving these processes are proposed. Overall, the two models are viable only if the initial crustal thickness were 43–47-km-thick; otherwise, the lost mass of the ESRP is not balanced by the excess mass in the Basin and Range. In the first model (densification, eruption, and extension by dike injection), the ESRP crust receives mantle-derived gabbro equivalent to a 16-km-thick layer, but would need to lose 10 km of crust via lower crustal flow. In the second model (densification, eruption, and extension by faulting and ductile attenuation), the ESRP crust receives a 9.4 km equivalent thickness of mantle-derived gabbro, and would need to lose 6 km of crust via lower crustal flow. The cumulative amount of mantle-derived gabbro in the ESRP crust derived by this kinematic analysis (equivalent to a 9.4–16-km-thick layer) is nearly identical to the amount estimated independently using a mass balance petrologic model, suggesting it is a robust result.

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1. Introduction

Basalt magma was transferred from mantle to crust during passage of the ESRP over the Yellowstone hotspot. Fractionation, assimilation, freezing, and remelting accompanied magma ascent, resulting in an intimate mixture of mantle and crustal sources that formed two compositional extremes, rhyolite and tholeiitic basalt (Leeman, 1982a; Hildreth et al., 1991). Some of this mixture was erupted as volcanic rock, while the rest remained underground as completely to partially solidified plutonic rock. The volume of volcanic rock was huge: individual rhyolitic eruptions exceeded 10^3 km^3 and the cumulative volume of erupted rhyolite and basalt was possibly as large as 10^5 km^3 (Leeman, 1989; McCurry et al., 2008). An even greater volume of plutonic rock is interpreted to exist according to petrologic mass balance calculations (McCurry and Rodgers, 2009–this volume). While some of the volcanic and plutonic rock merely represents crustal mass that moved from one place to another, McCurry and Rodgers (2009–this volume) calculated that $\sim 3.4 \times 10^5 \text{ km}^3$ was mantle-derived gabbro, equivalent to a layer 14-km-thick beneath the ESRP.

How did the ESRP crust accommodate the ascent of magma from the mantle? The ESRP crust could dilate, densify, or lose mass to

adjacent regions (Fig. 1). Dilation would be accomplished by crustal thickening, widening, or lengthening. Densification would involve the packing of more mass into a constant volume. Mass escape refers to the transfer of ESRP crust (mantle-derived or not) to the surface or to the Basin and Range crust. This paper will evaluate the likelihood of whether each process occurred, individually or in concert with other processes, by interpreting the kinematic evolution of the crust during hotspot tectonism. Where some form of crustal dilation, densification, or mass escape can be documented, the magnitude of volume or mass change will be calculated. The intent is to provide an estimate of the amount of mass transferred from mantle to crust during hotspot tectonism, independent of petrologic mass balance calculations, in more detail than before (Rodgers et al., 2002a).

Two types of data are available for documenting the kinematic evolution of the crust in the ESRP region. First, modern crustal characteristics including the length, width, thickness, and density of different layers are known through field study of surface rocks, drilling, gravity analysis, heat flow studies, and especially through seismic reflection, refraction, and tomography (Doherty et al., 1979; Braille et al., 1982; Sparlin et al., 1982; Kuntz et al., 1992; Pierce and Morgan, 1992; Peng and Humphreys, 1998; Gilbert and Sheehan, 2004; Schutt et al., 2008). Second, the nature of the crust prior to hotspot tectonism can be interpreted by removing the depositional and deformational changes associated with hotspot and Basin-Range tectonism. Some changes can be documented by studying the

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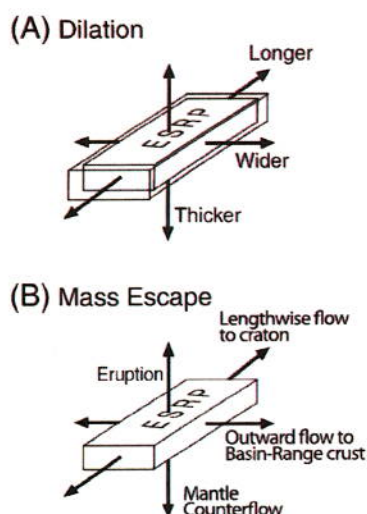


Fig. 1. Schematic illustration of how the ESRP crust could accommodate emplacement of mantle-derived mafic magma. (A) Dilation involves an increase in ESRP crustal volume through extension or thickening. Only axis-parallel extension has been recognized for the ESRP. (B) Mass escape involves the transfer of mass (ESRP crust or mantle-derived mafic magma) beyond the original boundaries of the ESRP via eruption, mantle-counterflow, or lower crustal flow. Only eruption and outward flow to the Basin-Range have been recognized for the ESRP. Not shown is the process of densification, in which ESRP crust experiences a net density increase due to emplacement of mantle-derived magma.

structural and volcanic evolution of the upper crust and extrapolating the results to deeper levels (Armstrong et al., 1975; Brott et al., 1981; Leeman, 1982a,b; Morgan et al., 1984; Anders et al., 1989; Leeman, 1989; Rodgers et al., 1990; Pierce and Morgan, 1992; McQuarrie and Rodgers, 1998; Parsons et al., 1998; Bindeman et al., 2007). Other changes can be interpreted by measuring the kinematic history of surface rocks over the past 10–50 years (e.g. Eddington et al., 1987; Puskas et al., 2007; Payne et al., 2008) and extrapolating the results back in time. Still other changes are identified by contrasting the modern crust to analogous crust beyond the hotspot domain.

2. Current crustal architecture

The ESRP is an elongate topographic and structural depression that cuts obliquely across the northern Basin and Range Province (Fig. 2). Its upper layer is a basin filled with Neogene–Quaternary volcanic, plutonic, and sedimentary rocks (Table 1). Drilling indicates that about 1–2 km of interbedded Pliocene–Pleistocene tholeiitic basalt, rare evolved lavas, and minor terrestrial sediments are present (Doherty et al., 1979; Kuntz et al., 1992; Anderson and Liszewski, 1997). These are underlain by Miocene–Pliocene rhyolite lavas, tuffs, and epizonal plutons, whose thickness is 2–5 km based on modeling of gravity and seismic data (Braile et al., 1982; Sparlin et al., 1982) or 5 km based on flexural modeling of the ESRP basin (McQuarrie and Rodgers, 1998). Modeling the entire basalt + rhyolite + sedimentary basin as a rectangular body, with vertical sides coincident with the ESRP margins and a flat floor 6-km-deep, provides a first-order estimate of its dimensions.

Paleozoic clastic and carbonate rocks, which are widely exposed beyond the ESRP, are inferred from 6 to 10 km depth based on seismic velocities (Sparlin et al., 1982). Miocene granitic rocks associated with hotspot magmatism are also likely at these depths.

An abrupt increase in seismic velocity at about 10 km depth marks the top of the ESRP mid-crustal “sill”, a crustal layer interpreted to contain a mixture of typical upper crustal rock and denser gabbro. Sparlin et al. (1982) interpreted the sill to have a density of 2880 kg/m³ and thickness of 10 km. McQuarrie and Rodgers (1998) used flexural modeling to calculate a thickness of 17 km, and tomographic data led Peng and Humphreys (1998) to interpret a sill density of

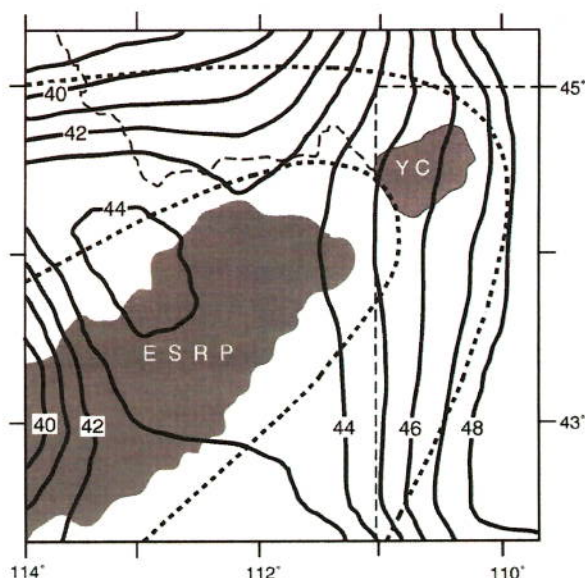


Fig. 2. Map of the Eastern Snake River Plain (ESRP) and Yellowstone Caldera (YC) region showing state boundaries (narrow dashes), the inner and outer boundaries of the seismic parabola (bold dashes) after Anders et al. (1989), and isopachs of crustal thickness (bold lines) after Schutt et al. (2008). The Basin and Range province surrounds the ESRP while the unextended craton is located east and north of outer boundary of the seismic parabola.

2830 kg/m³ and thickness of 10 km. Dueker et al. (2007) reported a thickness of 10 km and a density of 2880 kg/m³, which is used in this paper. The width of the layer was estimated to be 5–10 km beyond the ESRP margins based on gravity modeling (Sparlin et al., 1982), but flexural modeling and revised gravity modeling is more compatible with a width similar to the ESRP (Peng and Humphreys, 1998; McQuarrie and Rodgers, 1998).

Below this layer, the lower crust extends to a depth of 42 km (Eaton et al., 1978; Sparlin et al., 1982; Peng and Humphreys, 1998) or 43 km (Schutt et al., 2008). S-wave velocity is slow in the lowermost crust, evidence of a 3–5-km-thick zone of partial melting (Greensfelder, 1981; Peng and Humphreys, 1998). Seismic velocity below the Moho is indicative of mantle rock which appears to be uniformly hot (up to 2% partial melt) and anisotropic with a strong SW alignment of olivine crystals (Schutt et al., 1995; Schutt and Dueker, 2008).

The northeastern Basin and Range province surrounds the ESRP. Its heterogeneous upper crust includes uplifted horsts of late Proterozoic to Eocene sedimentary, volcanic, and plutonic rock and intervening

Table 1

Stratigraphic architecture of the ESRP and B&R crust, including the thickness (*T*) and density (ρ) of major layers. References include E (Eaton et al., 1978), G (Greensfelder, 1981), S (Sparlin et al., 1982), P&H (Peng and Humphreys, 1998), M&R (McQuarrie and Rodgers, 1998), D (Dueker et al., 2007), Sch (Schutt et al., 2008).

	<i>T</i> (m)	ρ (kg/m ³)	Reference
ESRP layers			
Basalt, sediments	1000	2000	Drilling
Rhyolite, granite	5000	2500	S, M&R
Granite/PZ crust	4000	2670	S, P&H
Mid-crustal sill	10,000	2880	S, P&H, D
Lower crust	18,000	3000	E, S, P&H
Partial melt	5000	2950	G, P&H
Mantle	≥43,000 depth	3300	Sch
B&R layers			
Upper crust seds	5000	2450	S
Upper crust	15,000	2670	S, P&H, Sch
Lower crust	24,000	3000	S, P&H, Sch
Mantle	≥44,000 depth	3300	Sch

half-grabens filled with upper Cenozoic detrital and volcanic rocks. Active extension is oriented east–west (to the south) or northeast–southwest (to the north) and concentrated in a parabolic zone (Fig. 2) whose vertex coincides with the Yellowstone hotspot (Anders et al., 1989; Smith et al., 1989; Pierce and Morgan, 1992). Seismic velocities typical of upper crust are observed to about 20 km depth and typical lower crustal velocities are observed below this, though different depths to the Moho have been reported. Just south of the ESRP, crustal thickness was measured at 38 ± 2 km (depending upon the choice of Moho; Peng and Humphreys, 1998), 41–43 km (Gilbert and Sheehan, 2004), or 42–44 km (Schutt et al., 2008). In contrast, crustal thickness just north of the ESRP was reported to be 40 ± 2 km (Peng and Humphreys, 1998), 39.7 km (Zeiler et al., 2005), or 42–45 km (Schutt et al., 2008). The most recent measurements of crustal thickness, based on joint inversion of phase velocity maps and Moho P_{ms} times (Schutt et al., 2008), are used in this paper (Fig. 2). The uppermost mantle beneath this part of the Basin and Range province is seismically fast but anomalously buoyant (Saltzer and Humphreys, 1997), which Humphreys et al. (2000) interpreted as a partial melt residuum that was displaced via convection from beneath the ESRP.

3. Tectonic history

Continental rifting and passive margin sedimentation affected the ESRP crust from Proterozoic to Devonian time, and additional shallow marine to terrestrial sediments accumulated from Mississippian to Triassic time. From Late Jurassic to early Paleogene time the Sevier fold and thrust belt developed throughout the region, and during the Paleogene parts of south central Idaho were strongly extended. These events created a heterogeneous crust whose thickness just prior to Neogene tectonism is poorly constrained. One estimate of thickness can be made by measuring the crustal thickness in an adjacent area that was unaffected by Neogene hotspot magmatism or associated extension, and extrapolating that value to the pre-hotspot ESRP. The nearest region of this type is the craton about 130 km southeast of the ESRP, where Peng and Humphreys (1998) located the Moho at a depth of 47 km and Schutt et al. (2008) placed it at 47–49 km. Simply assuming that the cratonal thickness characterized the pre-hotspot ESRP is not entirely accurate, however, because the southern Idaho crust experienced several tectonic events that the craton did not. Because of this uncertainty, a range of values from 40–60 km is used for the crustal thickness of the ESRP and adjacent Basin and Range prior to Neogene tectonism, and all results are presented in terms of a range of initial thicknesses.

Neogene silicic volcanism initiated at ~17 Ma in a diffuse zone (Christiansen et al., 2002; Brueseke et al., 2008) that overlaps in time and space with southern parts of the Columbia River basalt large igneous province (Camp et al., 2003; Brueseke et al., 2007). While basalt volcanism waned rapidly (Brueseke et al., 2007), spatially diffuse eruptions of rhyolites transitioned towards the west (e.g. Jordan et al., 2004) and especially northeast, along the YSRP volcanic track, into a ~80 km wide, physiographically well-defined structural downwarp (Kirkham, 1931) referred to as the eastern Snake River Plain (ESRP). Each large volcanic field remained active for ~2–4 m.y., and each produced ~5 to 10×10^3 km³ of distinctive Snake River-type rhyolite deposits (Armstrong et al., 1975; Bonnicksen, 1982; Embree et al., 1982; Williams et al., 1982; Leeman 1982a,b; Morgan et al., 1984; Rytuba and McKee, 1984; Pierce and Morgan, 1992; Kellogg et al., 1994; Perkins et al., 1998; Perkins and Nash, 2002; Pierce et al., 2002; Morgan and McIntosh, 2005; Nash et al., 2006; Branney et al., 2008; Bonnicksen et al., 2008). Minor olivine tholeiite lavas are interbedded with the rhyolites in some areas (e.g. Bonnicksen et al., 2008). Rhyolite magmatism shifted to the Yellowstone Plateau between 4 and 2 Ma and remains there today (Christiansen, 1984; Smith and Braile, 1994; Hildreth et al., 1991; Christiansen, 2001; Lowenstern and Hurwitz, 2008).

Olivine tholeiite volcanism postdates and overlaps spatially with rhyolite volcanism. In the eastern Snake River Plain these basalts accumulated at a nearly constant rate over the last 4 m.y. to a maximum thickness of 1–2 km (Kuntz et al., 1992; Champion et al., 2002). Small volumes of young (<2 Ma) olivine tholeiite lavas also erupted across an arc that overlaps large parts of the western and eastern tracks of silicic volcanism (Christiansen et al., 2002; Jordan et al., 2004).

The crust adjacent to the ESRP extended in an east–west to NE–SW direction during the Late Cenozoic to create the northeastern Basin and Range province. Extension appears related in space and time to hotspot volcanism: just south of the ESRP, a mid-Miocene pulse of minor extension was succeeded by an 11 to 0 Ma pulse of major extension. The locus of major extension was slightly ahead of, and beside, the locus of eastward-migrating rhyolite volcanism (Allmendinger, 1982; Rodgers et al., 1990, 2002b). Measurements of the modern kinematic activity are compatible with the long-term geologic record. Analyses of earthquakes (e.g. Stickney and Bartholomew, 1987; Eddington et al., 1987) indicate east–west to NE–SW extension, localized in a parabolic zone that flanks the ESRP (Anders et al., 1989). Decade-long monitoring of crustal motions using GPS technology (Puskas et al., 2007; Payne et al., 2008) indicates that, relative to a fixed North America reference frame, the ESRP and surrounding Basin and Range are rotating clockwise about a pole located in northern Idaho. Separation from the craton is accomplished throughout the parabolic zone of seismicity (Anders et al., 1989) and Quaternary faulting (Pierce and Morgan, 1992).

The topographic and structural depression that characterizes the ESRP reflects prolonged crustal subsidence that was first recognized by Kirkham (1931). The elevation decrease from northeast to southwest was interpreted to reflect thermal subsidence in the wake of the hotspot (Brott et al., 1981; Anders and Sleep, 1992). Subsidence with respect to the adjacent Basin and Range, where heat flow is significantly less, is evidence of subsidence due to loading. The load that initiated subsidence is interpreted to be an unusually dense mid-crustal “sill” (Anders and Sleep, 1992; McQuarrie and Rodgers, 1998).

Extension, subsidence, and emplacement of the mid-crustal “sill” exemplify the crustal kinematics and mass transfer that accompanied hotspot tectonism. As we attempt to quantify the effects of these and related processes, we seek evidence that the ESRP (and adjacent) crust has experienced a net increase in mass during the past 17 Ma. Our approach is to quantify the kinematic history of the ESRP and surrounding crust to document its net change in size, shape, and density.

4. Crustal dilation, densification, and mass escape

To accommodate the injection of magma from the mantle, the ESRP crust could dilate, densify, or lose mass to adjacent regions (Fig. 1). Densification of the ESRP middle crust was recognized by Sparlin et al. (1982) and Braile et al. (1982) based on seismic and gravity data, and resolved in more detail by Peng and Humphreys (1998) and Schutt and others (2008). Dilation of the entire ESRP crust via dike injection was mentioned by Rodgers et al. (1990) and expanded upon by Parsons and Thompson (1991), Kuntz et al. (1992) and Parsons et al. (1998). Loss of mass via lower crustal flow away from the ESRP was discussed by McQuarrie and Rodgers (1998), Peng and Humphreys (1998), and most recently by Schutt and others (2008). Each process is considered in the following section — some are shown to be insignificant while others important in accommodating the influx of mantle-derived mass. Where evidence permits the transfer of mass from mantle to crust, we consider that mass to be mafic magma whose density, once solidified to gabbro within the crust, is 3000 kg/m³. To better conceptualize the huge volumes of gabbro that were transferred, calculated volumes are divided by the modern areal dimensions of the ESRP and described as thicknesses of a layer beneath the ESRP. With average ESRP dimensions of 80 km

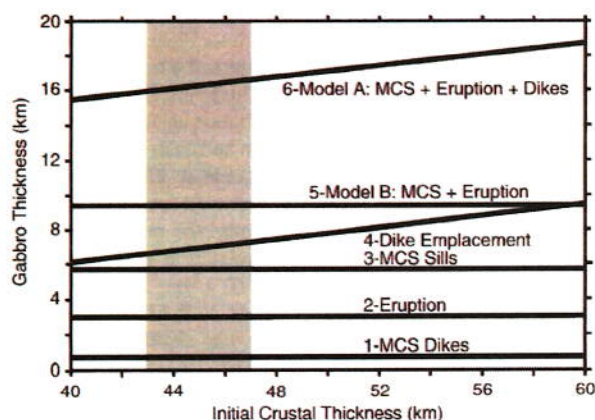


Fig. 3. Graph showing amount of gabbro added to the ESRP crust by several processes as a function of initial crustal thickness. Gabbro volume is divided by ESRP surface area to yield an equivalent “gabbro thickness”. Each line represents a magmatic process, or combination of processes, that would add gabbro to the crust. Horizontal lines indicate the process is independent of initial crustal thickness. As described in detail in the text, these processes include (1) dike injection into the mid-crustal sill (MCS); (2) gabbro that left the initial ESRP crust via eruption but remains in the overlying ESRP basin; (3) sill injection into the MCS; (4) dike emplacement throughout the entire initial crustal thickness (except the MCS); (5) a combination of 1, 2, and 3; and (6) a combination of 1, 2, 3, and 4. The vertical gray band from 43–47 km represents the preferred range of ESRP initial crustal thickness, based upon a mass balance (discussed in text and Fig. 7) between gabbro added to the ESRP and crust that left the ESRP via lower crustal flow to the surrounding Basin and Range.

wide and 300 km long, each kilometer thickness of gabbro is equivalent to $2.4 \times 10^4 \text{ km}^3$ of gabbro or $7 \times 10^{16} \text{ kg}$ of mass beneath the ESRP. In the following analysis, we ignore the top layer of Pliocene–Pleistocene basalt, since it constitutes a relatively small fraction of the total mantle-derived mafic magma in the ESRP.

4.1. Dilation via crustal thickening

If thickening were the only change in ESRP crustal dimensions during hotspot tectonism, the increase in thickness would record the amount of mantle-derived magma now residing as vertically stacked sills or horizontally sheared dikes. By measuring the modern crustal thickness and comparing it to the pre-hotspot thickness, the volume of new magma could be calculated.

The modern base of the ESRP crust is 43-km-deep and generally horizontal with minor undulations (Schutt et al., 2008; Sparlin et al., 1982; Peng and Humphreys, 1998; Dueker et al., 2007). Because the modern ESRP includes a 6-km-thick Neogene basin on top of the former ESRP crust, a 37-km-thick crust is what remains of the pre-hotspot crust. With a modern thickness of 37 km and an estimated pre-hotspot thickness of 40–60 km, the cumulative effect of hotspot tectonism on the former ESRP crust would have been thinning, not thickening. Magma injection via sills could not have happened alone. Only if the crust was thinned by other processes, to a thickness less than 37 km, would there be space for the emplacement of sills and sheared dikes. One such process is crustal extension.

4.2. Dilation via crustal extension

The emplacement of mantle-derived magma could be accommodated by extension of the ESRP crust. Two end-member scenarios are possible. First, extension at all depths might have occurred via emplacement of vertical mafic dikes, in which case the amount of extension would directly record the volume of emplaced magma. Alternatively, extension may have occurred by normal faulting in the upper crust and ductile attenuation in the lower crust, resulting in a specific amount of crustal thinning due to plane strain deformation. If the actual crustal thickness exceeds the predicted thickness due to

extension, syn-extensional sills and horizontally sheared dikes might be present.

Evidence for axis-parallel extension of the ESRP crust comes from analysis of modern landforms and study of Neogene structures in the surrounding Basin and Range Province. The modern ESRP is characterized by NNW-trending volcanic fissure zones (Kuntz et al., 1992) that cut across slightly older Quaternary basalt flows. Parsons et al. (1998) inferred the size and frequency of dike events to suggest that the rate of Pliocene–Quaternary extension was similar to that in the northern Basin and Range. In another study, Rodgers et al. (2002b) summarized evidence that Neogene normal faults of the Basin and Range Province continue beneath the ESRP where they are buried beneath younger volcanic rocks. They interpreted that the kinematics and timing of ESRP extension is similar to the kinematics and timing of immediately adjacent Basin and Range faults. The cumulative amount of Basin and Range extension south of the ESRP is 25% (Rodgers et al., 2002b) and north of the ESRP is 10–15% (Janecke, 1993; Anders et al., 1993; Rodgers et al., 1995). Fault timing is not well-constrained north of the ESRP, but slip on major faults south of the ESRP was predominantly pre- to syn-eruption of nearby silicic volcanic rocks, such that a pulse of normal faulting migrated ahead of, and beside, the locus of hotspot volcanism (Allmendinger, 1982; Anders et al., 1989; Rodgers et al., 1990, 2002b). Overall, the ESRP crust is interpreted to have experienced 20% extension ($\beta = 1.2$) parallel to its ENE axis at times immediately before, during, and after hotspot volcanism.

If ESRP extension was accommodated solely by dike emplacement, as seems to be the dominant style since the Pliocene, then the extension amount can be used to calculate the volume of emplaced gabbroic magma. The volume of gabbro as a function of initial crustal thickness is shown in Fig. 3 (Dike Emplacement plus MCS Dikes) to be equivalent to adding a 6.7–10-km-thick layer of gabbro beneath the ESRP, depending upon the initial crustal thickness. If β is not 1.2 but instead as much as 20% more or less ($1.15 < \beta < 1.25$), the equivalent thickness of gabbro would range from 5.9 to 9.0 km for a 45-km-thick initial crust (Fig. 4).

The simple dike model contradicts the observed dimensions of the ESRP in one important way. Extension via dike emplacement results in no change in crustal thickness, whereas the former ESRP crust is interpreted to have thinned from 40–60 km to 37 km. If dike emplacement was significant, some other process must have acted in concert to thin the crust by an additional 3–23 km.

Alternatively, extension may have been accommodated by normal faults in the upper crust and ductile attenuation in the lower crust. In this plane strain model, crustal extension would be matched by proportional crustal thinning: 20% extension of a 40–60-km-thick ESRP crust would create a 33–50-km-thick crust (Fig. 5). Compared to the 37 km modern thickness of the ESRP crust (not including the overlying basin), this indicates that simple extension of the ESRP crust by normal faults and ductile flow would only create space for mantle-derived sills if the initial crust were thinner than 44 km. If the initial

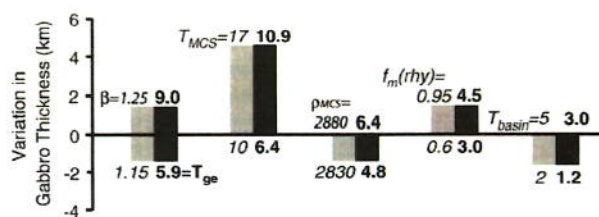


Fig. 4. Histogram showing the sensitivity of calculated gabbro thicknesses associated with different magmatic processes (Fig. 3) to varying input parameters. Variation of each parameter from the preferred value is shown by the gray bar above or below the zero line, and the consequent change in gabbro thickness (T_{ge}) is shown by the adjacent black bar. Abbreviations: β = extension factor for the ESRP, T_{MCS} = thickness of the mid-crustal sill, ρ_{MCS} = density of the mid-crustal sill, $f_{m(rhy)}$ = fraction of rhyolite derived from mantle, T_{basin} = thickness of ESRP basin.

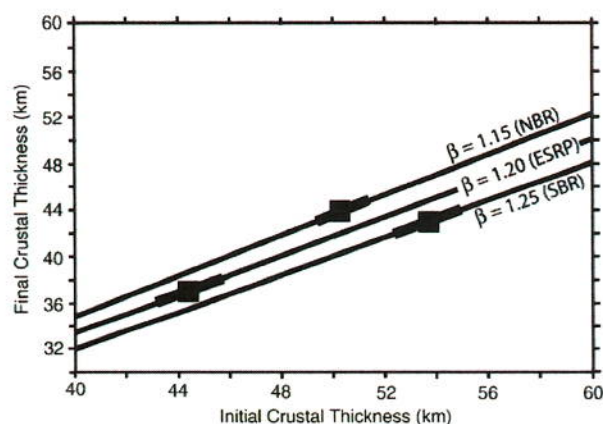


Fig. 5. Graph showing final crustal thickness after plane strain extension of various initial crustal thicknesses. Measured extension amounts (Rodgers et al., 1990, 2002b) shown for the Basin and Range just north (NBR) and south (SBR) of the Eastern Snake River Plain (ESRP). Modern crustal thicknesses described by Schutt et al. (2008) are indicated by black square (average thickness) and bold black lines (range of thicknesses). ESRP thickness does not include the 6-km-thick overlying basin that postdates extension.

crustal thickness exceeded 44 km, other processes would have to operate in concert with crustal thinning to accommodate the emplacement of mantle-derived gabbro.

4.3. Dilation via crustal widening

Like crustal extension parallel to its axis, extension normal to its axis could make the ESRP increase in crustal volume and accommodate the addition of mantle-derived magma. Dikes or normal faults would be oriented parallel to its axis, and ductile attenuation could occur at depth. The amount of crustal widening could be used to estimate the volume of emplaced magma.

However, there is insufficient evidence of ESRP crustal widening. The surface of the modern ESRP shows few normal faults oriented parallel to its axis. An abrupt gravity gradient along the northern margin of the ESRP led Pankratz and Ackermann (1982) to interpret the presence of a concealed normal fault there, but other workers found no significant normal faults (McQuarrie and Rodgers, 1998; Rodgers et al., 2002b) or velocity gradients (Peng and Humphreys, 1998) in the marginal rocks and instead interpreted the gravity gradient to reflect the sharp edge of a dense mid-crustal layer. Plain-parallel faults with minor (100s of meters) amounts of normal offset are present in the Basin and Range Province within 30 km of the ESRP (Zentner, 1989; Rodgers et al., 2002b) and within the interior of the seismic parabola (Janecke et al., 2000). The faults were interpreted to reflect extension along the upper arc of flexure associated with ESRP subsidence, not regional NW–SE crustal extension. No shortening structures oriented parallel to the ESRP axis have been recognized in the adjacent Basin and Range, which might indicate contractional strain that compensated for an ESRP extensional strain field. Overall, there is no evidence to indicate the ESRP crust extended normal to its axis.

4.4. Densification

If the cumulative density of the ESRP crust increased during hotspot tectonism, the net increase could reflect the addition of new mass to the crust. When comparing the density stratification of the ESRP to that in the Basin and Range, the only significant contrast occurs in the middle crust where the ESRP is characterized by the mid-crustal “sill” (MCS) whose density is higher (2880 kg/m³ vs 2670 kg/m³) than upper crust in the adjacent Basin and Range (Sparlin et al., 1982; Peng and Humphreys, 1998; Dueker et al., 2007). The MCS is widely interpreted to be a mixture of mantle-derived gabbro sills and dikes

emplaced into the former upper crust (Sparlin et al., 1982; Leeman, 1982a,b). A simple mixing model with end-members of gabbro (3000 kg/m³) and upper crust (2670 kg/m³) indicates that a gabbro: upper crust ratio of 64:36 is needed to create a final crust of 2880 kg/m³. Current estimates of the MCS indicate that its cumulative thickness is 10 km (Dueker et al., 2007), suggesting it consists of about 6.4 km equivalent thickness of mantle-derived gabbro. A small fraction of that mass (0.7 km equivalent thickness) would accommodate the 20% horizontal extension (Fig. 3, MCS dikes), while the remaining 5.7 km would reflect vertical thickening of the initial crust (Fig. 3, MCS sills). Variations in the input parameters yield small variations in the calculated gabbro thickness (Fig. 4): if the mid-crustal sill were as thick as 17 km (McQuarrie and Rodgers, 1998), the equivalent thickness of gabbro would increase to 10.9 km, and if the mid-crustal sill were less dense (2830 kg/m³, Peng and Humphreys, 1998), the equivalent gabbro thickness would diminish to 4.8 km. Neither of these estimates are as well supported by the most recent geophysical data, so 6.4 km equivalent thickness of gabbro in the mid-crustal sill is the preferred value.

MCS densification could not happen alone, since the nearly 6 km of thickening is not supported by the modern ESRP thickness. Excess crustal material must have been expelled beyond the ESRP during the densification process via eruption or outward-directed lower crustal flow, as described below.

4.5. Mass escape — eruption

Mass escaped from the ESRP crust during ascent and eruption of silicic magma associated with hotspot tectonism. Most of the silicic volcanic rock ultimately accumulated in the overlying basin, though a lesser amount of ash fall and ignimbrite traveled to more distant locations (Izett, 1981; Perkins et al., 1998, 2002). Volumes of individual caldera-forming eruptions from the Heise volcanic field in the ESRP range from 750–1800 km³ (Morgan and McIntosh, 2005). In addition, some silicic magma ascended into the lower portion of the basin and solidified as plutonic rock (Bindeman et al., 2007). Whether the silicic magma was derived from a mixture of former crust and mantle-derived gabbro, or was only former crust that, by escaping, made room for the emplacement of mantle-derived dikes and sills, the volume of silicic rock records mass transfer beyond the top of the original ESRP crust.

The 5-km-thick layer of rhyolite and epizonal granite (not including the 1 km of Plio-Pleistocene basalt) in the ESRP basin is derived from partial melting of the MCS and is conservatively estimated to consist of ~60% mantle-derived mass, based on Nd isotope analyses and a mixing model that invokes complete homogenization of the MCS (McCurry and Rodgers, 2009–this volume). Thus, rhyolite in the ESRP basin records the transfer from mantle to basin of a layer of gabbro about 3-km-thick (Fig. 3, Eruption). The remaining rhyolite mass, about 2-km-thick, was derived from ESRP lower or middle crust and merely relocated, via magmatic processes, to the basin. Variations in these input parameters also yield small variations in the calculated gabbro thickness (Fig. 4): if the layer of rhyolite and granite was thinner than 5 km (as thin as 2 km based on the 3.16 km total depth penetrated by drilling (Doherty et al., 1979; Morgan et al., 1984)) the equivalent thickness of gabbro would decrease to 1.2 km, while if the mass fraction of rhyolite derived from the mantle were higher than 0.6 (up to 0.95, McCurry and Rodgers, 2009–this volume) the equivalent gabbro thickness would increase to 4.5 km. Neither of these estimates are as well supported, so 3.0 km equivalent thickness of gabbro in the ESRP basin is the preferred value.

4.6. Mass escape — outward flow

Outward flow of ESRP lower crust to the adjacent Basin and Range Province could accomplish mass transfer and thin the ESRP crust. The

Table 2

Summary of thickness (T) and thickness changes (ΔT) of the Eastern Snake River Plain (ESRP) crust due to different kinematic processes discussed in text, assuming an initial crustal thickness of 40–60 km. Also shown is excess thickness of Basin and Range (B&R) crust, assuming 15% (north) and 25% (south) extension of the initial crust, using the modern thickness measured by Schutt et al. (2008). All numbers refer to the thickness (in kilometers) of a layer beneath the ESRP. Results are plotted in Fig. 7.

$T_{\text{crust initial}}$	40	45	50	55	60
ΔT_{ESRP} by different processes					
Densification	5.7	5.7	5.7	5.7	5.7
Eruption/basin	–2.0	–2.0	–2.0	–2.0	–2.0
Dike injection	0	0	0	0	0
Faults/ductile flow	–6.7	–7.5	–8.3	–9.2	–10.0
Model A: densification + eruption + dike injection					
ΔT_{ESRP}	3.7	3.7	3.7	3.7	3.7
$T_{\text{ESRP final}}$	43.7	48.7	53.7	58.7	63.7
$T_{\text{ESRP measured}}$	37.0	37.0	37.0	37.0	37.0
$T_{\text{ESRP loss}}$	6.7	11.7	16.7	21.7	26.7
Model B: densification + eruption + faults/ductile flow					
ΔT_{ESRP}	–3.0	–3.8	–4.6	–5.5	–6.3
$T_{\text{ESRP final}}$	37.0	41.2	45.4	49.5	53.7
$T_{\text{ESRP measured}}$	37.0	37.0	37.0	37.0	37.0
$T_{\text{ESRP loss}}$	0.0	4.2	8.4	12.5	16.7
Excess thickness of the B&R crust					
$T_{\text{B\&R excess}}$	12.7	7.4	2.2	–3.0	–8.2

driving force for lower crustal flow would be densification of the ESRP upper crust: as gabbro was injected into the ESRP upper crust, the underlying lower crust would flow outward to maintain isostatic equilibrium. Peng and Humphreys (1998) and McQuarrie and Rodgers (1998) provide evidence for outward-directed lower crustal flow based on isostatic models of elevation differences and basin formation. More recently, Dueker et al. (2007) identified unusually low velocity lower crust not only beneath the ESRP but also within 40–60 km of the eastern ESRP in the Basin and Range, and interpreted this as strong evidence of lower crustal flow. However, they did not report a thickness for the anomalous lower crust. For the present, a more indirect estimate of the lower crustal flow volume must be made by comparing the modern Basin and Range crustal thickness to the thickness predicted according to plane strain extension of the initial crust.

This approach requires that, during the past 15 m.y., the Basin and Range experienced no mass transfer besides lower crustal flow from the ESRP and no densification. In support of these assumptions, geophysical studies provide no evidence that Basin and Range crustal layers have unusual thicknesses or densities (Gilbert and Sheehan, 2004; Schutt et al., 2008). Only rare Miocene to recent volcanism has been documented in this part of the Basin and Range: there are very few volcanic vents north of the ESRP, and the ones located south of the ESRP are interpreted as off-axis ESRP vents (Ford et al., 2004; Ford, 2005). The amount of sediment lost or gained during the Neogene and Quaternary has not been estimated, but little is suggested by the roughly equal proportion of uplifts and half-graben basins.

The expected thickness of Basin and Range crust after extension can be estimated by invoking plane strain extension of a 40–60-km-thick initial crust. As shown in Fig. 5, 15% extension ($\beta = 1.15$) north of the ESRP yields a crust that is 35–52-km-thick and 25% extension south of the ESRP yields a crust that is 32–48-km-thick. Compared to the estimated thicknesses for the modern Basin and Range crust – about 43 km (range of 42–44) south of the ESRP and 44 km (range of 42–45) north of the ESRP (Schutt et al., 2008) – this indicates that simple extension by normal faults and ductile flow would create space for ESRP-derived lower crustal flow if the Basin and Range crust were initially thinner than 50 to 54 km (Fig. 5).

If the Basin and Range crust were initially thinner than 50–54 km, extension of that crust would yield a crust thinner than the modern

Basin and Range crust, and the difference between the calculated and modern thicknesses could be excess mass that flowed outward from the ESRP. To estimate the volume of lower crust that potentially flowed, the excess Basin and Range thickness (Fig. 5, Table 2) can be multiplied by the lateral extent of flow and by the length of the ESRP. Dueker et al. (2007) provided evidence of flow 50 ± 10 km beyond (north and south) the eastern ESRP, manifested by a low velocity lower crustal layer. Dividing the excess Basin and Range volume by the surface area of the ESRP allows us to compare the amount of excess mass to lost mass beneath the ESRP. Fig. 7 shows the excess lower crustal mass in the Basin and Range (both north and south) as a function of initial and modern crustal thickness. The excess Basin and Range mass is equivalent to an ESRP layer from 0–13-km-thick, depending upon the initial thickness of the Basin and Range crust (Fig. 7).

5. Kinematic models

The original question posed in this paper was: How did the ESRP accommodate the addition of mantle-derived magma? Based on the previous discussion only four processes appear to have been significant: extension, densification, eruption, and outward-directed lower crustal flow. There is no geologic or geophysical evidence to support other means of dilation (crustal thickening and widening) or mass escape (axis-parallel lower crustal flow, downward- or mantle-counterflow).

How much new mass was added by a combination of kinematic processes? To calculate this requires good knowledge of the previous dimensions and composition of the crust prior to hotspot and Basin and Range tectonism. Major unknowns include the original crustal thickness and the means by which the ESRP crust extended. We investigate a range of original thicknesses from 40–60 km and, to account for the unknown ESRP extension style, propose two end-member models and calculate the amount of mantle-derived mass in each.

In the first model, a 40–60-km-thick ESRP crust experiences 20% axis-parallel extension via vertical dike emplacement throughout the crust (Fig. 6, Model A). A 10-km-thick mid-crustal sill with density 2880 kg/m^3 forms within the new ESRP crust, replacing an Archean crust whose density was 2670 kg/m^3 . Of the new gabbro within the MCS, some is emplaced as dikes to accommodate 20% extension while the remainder (equivalent to a ~6-km-thick layer) is emplaced as sills. A 6-km-thick basin, that contains 2 km of crustal-derived mass and 3 km of mantle-derived mass, forms above the former ESRP surface. The new ESRP crust has a predicted thickness of 44–64 km

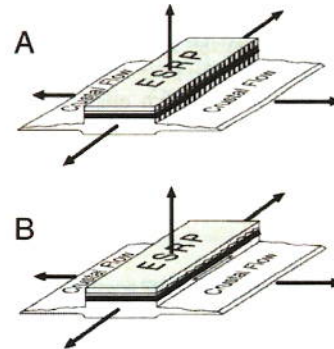


Fig. 6. Block diagrams illustrating two kinematic models of ESRP evolution. Both models involve densification via MCS formation (black layer), upward mass escape to fill the overlying basin with volcanic and plutonic rocks (gray layer), and outward mass escape via lower crustal flow into the adjacent Basin and Range. The models differ in the way they accommodate 20% axis-parallel extension: via emplacement of mantle-derived dikes in Model A (top) or via upper crustal normal faulting, MCS dike injection, and lower crustal ductile attenuation in Model B (bottom).

plus 6 km of new basin thickness for a total thickness of 50–70 km (Table 2). For Model A, the equivalent thickness of mantle-derived mass in the new ESRP crust would be 15.7–18.7 km including 7–10 km due to dike injection, 5.7 km due to MCS densification, and 3 km associated with mass escape into the basin (Table 2; Fig. 3).

In the second model, a 40–60-km-thick ESRP crust experiences 20% axis-parallel extension via plane strain normal faulting in the upper crust and ductile extension in the lower crust (Fig. 6, Model B). All other initial conditions, magmatic processes, and kinematic processes are identical to Model A. The new ESRP crust has a predicted thickness of 37–54 km plus 6 km of new basin for a total thickness of 43–60 km (Table 2). For Model B, the equivalent thickness of mantle-derived mass in the new ESRP crust would be 9.4 km including 0.7 km due to dike injection (only in the MCS), 5.7 km due to MCS densification (sills only), and 3 km due to mass escape into the basin (Table 2; Fig. 3).

Both Models A and B yield an excessively thickened ESRP crust. Compared to the measured thickness of the modern crust, Model A would be 7–27 km too thick and Model B would be 0–17 km too thick, depending upon the initial ESRP crustal thickness (Fig. 7, Models A and B). Only if the excess thickness were reduced to the modern thickness, through lower crustal flow to the adjacent Basin and Range, would these models be viable.

As discussed previously, the Basin and Range contains from 0–13 km excess mass that is interpreted to be derived via lower crustal flow from the ESRP (Fig. 7). If this interpretation is valid, a balance of ESRP lost mass and Basin and Range excess mass is indicated where the Models A and B lines cross the Basin and Range line in Fig. 7 (Points 1 and 2). Overall, the line between Points 1 and 2 represents the range of possible initial and final crustal thicknesses that would yield mass balance between the two provinces for end-member Models A and B. For Model A, mass balance is only possible if the initial crust were about 43–km-thick. For thicknesses >43 km, the Basin and Range was not thinned sufficiently to accommodate the additional mass expelled from the ESRP by lower crustal flow. For Model B, mass balance is only possible if the initial crust were about 47–km-thick. For thicknesses >47 km, the Basin and Range was not thinned sufficiently to accommodate the additional mass expelled from the ESRP by lower crustal flow.

Overall, the two models indicate a net increase in the total ESRP crustal mass during hotspot tectonism. In Model A (with an initial crustal thickness of 43 km), mantle-derived gabbro equivalent to a

16-km-thick layer was added (Fig. 3) while lower crust equivalent to a 10-km-thick layer was expelled via crustal flow (Fig. 7), yielding the net addition of an ESRP layer 6-km-thick. In Model B (with an initial crustal thickness of 47 km), mantle-derived gabbro equivalent to a 9.4-km-thick layer was added (Fig. 3) while lower crust equivalent to a 6-km-thick layer was lost (Fig. 7), yielding the net addition of an ESRP layer about 3.4-km-thick.

6. Comparison to petrology results

The results of this kinematic analysis can be compared to an independent assessment of mantle-derived gabbro in the ESRP published in a companion paper by McCurry and Rodgers (2009-this volume). These authors proposed a conceptual model of magmatism that included ascent of basaltic magma, crustal assimilation and homogenization, freezing in the middle crust, remelting, ascent to the upper crust, and ultimately eruption or upper crustal emplacement of silicic magma. A mass balance using Nd isotopes of rhyolite, basalt (as a proxy for mantle), and Archean crust has been calculated and indicates that the amount of mantle-derived gabbro emplaced into the ESRP crust was equivalent to a layer 14-km-thick (McCurry and Rodgers, 2009-this volume). This result is equivalent to the 9.4–16-km-thick layer of gabbro estimated in this paper using a kinematic model. The equivalent solution from two completely independent analyses indicates this is a robust result.

7. Conclusions and future work

This paper identifies several kinematic processes that altered the size, shape, and density of the ESRP crust during passage over the Yellowstone hotspot. The four major processes are densification due to emplacement of gabbro into the middle crust, subsidence and basin formation, axis-parallel extension via dike injection or faulting/ductile attenuation, and lower crustal flow from the ESRP to the adjacent Basin and Range. A quantitative analysis of these processes is limited by several unknowns including the initial crustal thickness, exact extensional process, and (to a lesser extent) detailed measurement of the modern crustal thickness. Despite these limitations, we estimate of the total amount of mantle-derived gabbro added to the ESRP crust is equivalent to a layer 9.4–16-km-thick, with the range depending upon whether ESRP extension was accommodated more by plane strain or dike injection. This thickness is equivalent to 2.3×10^5 to 3.8×10^5 km³ of gabbro, or 6.7×10^{17} to 11.2×10^{17} kg of mass. Independent estimates of the amount of mantle-derived gabbro, based upon petrologic mass balance calculations (McCurry and Rodgers, 2009-this volume), strongly support this conclusion. To make room for this gabbro, some ESRP crust, equivalent to a layer 6–10-km-thick, is interpreted to have flowed laterally into the adjacent Basin and Range crust. The final result was a net gain in ESRP crustal mass equivalent to a 3.4–6-km-thick crustal layer.

This study provides the kinematic framework to interpret crustal mass redistribution during passage over the Yellowstone hotspot. To refine the model results, at least two new measurements would provide better input parameters. First, the amount of extension could be resolved with higher spatial resolution. This paper uses two values of extension, one for each side of the ESRP, but in reality the amount of extension varies at length scales of ~50 km both parallel and perpendicular to the ESRP axis. Additional mapping and extension measurement within this portion of the Basin and Range would yield more highly resolved estimates of the amount of Neogene crustal extension and thinning. Second, if it were possible to directly measure the thickness of ESRP-derived lower crust in the Basin and Range, perhaps due to its seismic anisotropy or anomalous composition/temperature, the volume of lower crustal flow would not need to be indirectly calculated through estimates of initial crustal thickness and extension amount.

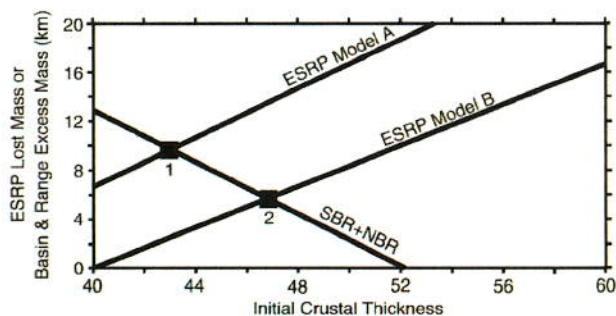


Fig. 7. To estimate the amount of lower crustal flow from Eastern Snake River Plain (ESRP) to Basin and Range (B&R), this graph shows intersecting lines. Mass gain or loss (Y axis) is measured in terms of the thickness of a lower crustal layer beneath the modern ESRP. The amount of mass lost by the ESRP is calculated by the difference between its modern crustal thickness (Schutt et al., 2008) and its predicted thickness according to Model A (MCS + basin + dike injection) or Model B (MCS + basin + plane strain extension). The amount of mass gained by the Basin and Range is calculated by the difference between its modern crustal thickness (Schutt et al., 2008) and its predicted thickness assuming plane strain extension, across a width of 50 km (Dueker et al., 2007) north and south of the ESRP. The intersection of lines shows where ESRP loss is balanced by Basin and Range gain and represents possible values of the amount of lower crustal flow for Model A (10 km, point 1) and Model B (6 km, point 2), as well as the range of possible values for initial crustal thickness (43–47 km).

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