Deformation of the NE Basin and Range Province: the response of the lithosphere to the Yellowstone plume?

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SUMMARY
Tectonic deformation in the NE Basin and Range Province of the western United States is concentrated in two zones, the N–S trending northern part of the Intermountain seismic belt in northern Utah, SE Idaho and SW Wyoming and the E–W trending Central Idaho seismic zone, which converge at Yellowstone in NW Wyoming. Accepting the popular view that the Yellowstone volcanic field is above an upwelling plume from the deep mantle, I explore the possibility that deformation of the brittle upper crust in these two seismic zones may be caused by return flow from this plume. This return flow is assumed sheared SW in the direction of motion of the North American plate relative to Yellowstone. The Snake River Plain (SRP), a linear outcrop of Neogene and Quaternary basalt that trends SW from Yellowstone for more than 400 km, bisecting the angle between the two seismic zones, may be a trail formed as the plate moves SW across the upwelling plume.

Beyond ~150 km from Yellowstone, slip vectors on active normal faults in both deforming zones are oriented inwards at ~45° towards the SW azimuth along the SRP. Normal slip rates on these faults are typically 1 mm yr⁻¹, implying extensional strain rate ~10⁻¹⁵ s⁻¹. Proportions of normal slip and strike slip suggest faults in the Intermountain seismic belt are rotating anticlockwise at ~2° Myr⁻¹ around vertical axes, and faults in the central Idaho seismic zone are rotating clockwise at similar rates. Within ~100 km of Yellowstone, observed sense of rotation around vertical axes in both zones appears reversed. Tilt rates of hanging wall beds suggest most faults in both zones are rotating at ~2° Myr⁻¹ (also ~10⁻¹⁵ s⁻¹) around horizontal axes oriented NW.

The relative size of rotation rate around vertical axes and horizontal extensional strain rate is shown to imply that velocity in the underlying deforming ‘fluid’ is small in the direction perpendicular to the axis of the SRP. In an initial model, this velocity is set to zero, enabling the two-dimensional pattern of flow at the horizontal upper boundary of this deforming fluid to be solved straightforwardly. This approximation enables observed rates and senses of horizontal extension and rotation around a vertical axis to be explained as consequences of the pattern of horizontal gradients of velocity parallel to the axis of the SRP in the deforming fluid beneath the brittle layer.

A refinement suggests that velocity perpendicular to the SRP near the upper boundary of the deforming ‘fluid’ is small but inward towards the SRP. This pattern of flow will act to move magma in the uppermost mantle beneath the SRP from its sides. It may thus reconcile petrological observations that SRP basalts do not show deep mantle characteristics expected had they been extruded from the Yellowstone plume, with the existence of this plume nearby.

Conservation of angular momentum around horizontal axes perpendicular to the SRP for the combined system of the plume interacting with the North American plate leads to the deduction that the plume radius is ~50 km, consistent with results of other investigations, and its upwelling velocity is ~60 mm yr⁻¹.

Key words: Basin and Range Province, continental deformation, fluid mechanics, Yellowstone

1 INTRODUCTION
Continental lithosphere may deform in response to boundary conditions imposed on it by its surroundings. These boundary conditions may result from buoyancy forces caused by large-scale topography (e.g. in Tibet; Molnar, Tapponnier & Chen 1981) or from motion of laterally adjacent stronger regions (e.g. in Italy; Westaway,
Gawthorpe & Tozzi 1989; and perhaps also in western Turkey and central Greece; McKenzie & Jackson 1983, 1986). Extensional deformation of the continental lithosphere may also conceivably be driven in some places by non-uniform motion of the mantle below, although no well-documented example has yet been identified.

The pattern of mantle convection beneath the Pacific Ocean, indicated by correlated geoid and bathymetry maxima and minima spaced 1500–2000 km apart, with geoid highs above upwelling convection plumes, has been revealed by combining marine geophysical and satellite altimetry observations (McKenzie et al. 1980). Geoid anomalies in this region are elongated parallel to the absolute direction of motion of the Pacific plate, indicating that they are sheared by the plate moving across them. No seismicity appears to occur above some of these plumes, suggesting that the oceanic lithosphere may be sufficiently strong to interact with plumes in this way without itself deforming—except for distorting to form the long-wavelength bathymetry that reveals the pattern of convection.

Continental lithosphere is weaker than oceanic lithosphere, and instead of comprising rigid plates may deform on a large scale. The possibility exists that it may deform over an upwelling plume, as well as shearing the plume in the direction of plate motion. To provide an observational test of this possibility, a region needs to be identified where an upwelling plume is adjacent to an area of distributed continental deformation, and where this deformation is not explained satisfactorily by other causes. One such region is the NE Basin and Range Province of the western United States, which lies close to the Yellowstone volcanic field, believed to be above an upwelling plume.

Ideally, it would be desirable to attempt a full solution of the dynamics of Yellowstone plume and the NE Basin and Range Province, either analytically or numerically. However, no analytic models exist for sheared plumes, and flow within the mantle is likely to be sufficiently vigorous that, given the limitations of the present generation of supercomputers (e.g. Craig & McKenzie 1987), attainable three-dimensional numerical dynamic models from first principles are unlikely to model it well. Modelling the dynamics of such a plume and its interaction with its surroundings properly is thus not yet possible. My investigation is therefore restricted to posing the problem and attempting a simplified solution for the kinematics only. Throughout this article these points should be borne in mind.

Bathymetry, geoid, and gravity anomaly signatures of plumes beneath oceanic areas are relatively easy to identify given the uniformity of oceanic lithosphere (e.g. McKenzie et al. 1980). Topography in the NE Basin and Range Province is dominated by ~2 km offsets across major active normal faults with structural wavelength ~30–100 km. Topography expected near an upwelling plume in the absence of lithosphere deformation has ~1 km amplitude and wavelength ~1000 km (e.g. McKenzie et al. 1980). Although it may be possible to filter the topography around Yellowstone to reveal a long-wavelength component, or to examine the long-wavelength pattern of topography along the direction of strike of major active faults, its signal to noise ratio is likely to be low. However, Yellowstone is associated with a ~+10 m geoid anomaly (e.g. Wagner et al. 1977) and a ~+20 mGal free air gravity anomaly that can be revealed by upward continuing the gravity field to ~160 km above the Earth's surface to filter out local features (Taylor et al. 1983) (Fig. 1). Both gravity and geoid anomalies have the magnitude and sign expected from numerical simulations of upwelling mantle plumes (e.g. McKenzie et al. 1980; Craig & McKenzie 1987), suggesting that one is present beneath Yellowstone.

Circulation in a plume and the surrounding deforming lithosphere may also cause other observable features at the Earth's surface. If extension in the overlying brittle layer is caused by circulation in a plume, the direction of this extension, revealed by the sense of slip on major active faults, should be related systematically to this circulation. In addition, vorticity within the plume may cause rotation of blocks in the overlying brittle layer. Consequently, measurement of senses of slip on faults and rotation of blocks in the brittle layer may be useful for constraining the circulation deeper in the Earth.

Having summarized this article elsewhere (Westaway 1989d), it is appropriate to present it in detail. In Section 2 I review established theoretical relationships between derivatives of the velocity field within the plastically-deforming lithosphere and patterns of deformation in the brittle layer above it, identifying features of this velocity field that are potentially observable, and develop new theory applicable to deforming zones such as the Intermountain seismic belt and central Idaho seismic zone, where a single set of parallel faults in the brittle layer takes up both extensional strain and vorticity in the underlying plastically-deforming 'fluid' (Fig. 2). In Section 3 I summarize observed deformation in the brittle layer of the NE Basin and Range Province, and derive estimates of deformation rates needed later when modelling this region. In Section 4 I briefly summarize models already suggested to explain evolution of this region over the last ~10 Myr, when the North American plate has apparently moved across the Yellowstone plume. No model that aims to explain observed deformation as an effect unrelated to Yellowstone is satisfactory. In Section 5 I discuss simple kinematic models for circulation in a plume that is sheared by a plate moving across it, and examine the extent of agreement between predictions of these models and observed deformation of the NE Basin and Range Province. Readers who are primarily interested in these suggested models rather than the theory or data on which they are based may proceed directly to Section 5.

2 RELATIONS BETWEEN OBSERVABLES AT THE EARTH'S SURFACE AND PATTERNS OF DEFORMATION AT DEPTH

One model for continental deformation (McKenzie & Jackson 1983, 1986) treats the bulk of the continental lithosphere as a deforming fluid. Fault-bounded blocks in the upper-crustal brittle layer, which in the NE Basin and Range Province is up to ~15 km thick, are regarded as floating on this fluid and responding to the pattern of flow within it. The deforming 'fluid' beneath a single set of parallel faults, should be related systematically to this circulation. In addition, vorticity within the plume may cause rotation of blocks in the overlying brittle layer. Consequently, measurement of senses of slip on faults and rotation of blocks in the brittle layer may be useful for constraining the circulation deeper in the Earth.
Deformation rate at any point in a deforming fluid can be characterized by its velocity gradient tensor, \( \mathbf{L} \) (e.g. McKenzie & Jackson 1983), where:

\[
\mathbf{L} = \begin{bmatrix}
\frac{\partial v_x}{\partial x} & \frac{\partial v_x}{\partial y} & \frac{\partial v_x}{\partial z} \\
\frac{\partial v_y}{\partial x} & \frac{\partial v_y}{\partial y} & \frac{\partial v_y}{\partial z} \\
\frac{\partial v_z}{\partial x} & \frac{\partial v_z}{\partial y} & \frac{\partial v_z}{\partial z}
\end{bmatrix} = \mathbf{E} + \mathbf{\Omega}
\]  

(2.2)

\( \mathbf{L} \) can thus be expressed as the sum of two tensors: \( \mathbf{E} \), the symmetric strain rate tensor; and \( \mathbf{\Omega} \), the rotation tensor, that is antisymmetric and quantifies local rotation rate within the fluid. Vorticity, \( \mathbf{\chi} \), where:

\[
\mathbf{\chi} = -\nabla \times \mathbf{v}
\]  

(2.3)

can be obtained from the rotation tensor as:

\[
\mathbf{\chi} \cdot \mathbf{x} = 2\mathbf{\Omega} \times \mathbf{x}.
\]  

(2.4)

With vorticity non-zero a fluid is said to be rotational. Vertical vorticity, corresponding to rotation around vertical axes, can be expressed as:

\[
\chi_z = - (\nabla \times \mathbf{v})_z = \frac{\partial v_y}{\partial y} - \frac{\partial v_x}{\partial x}.
\]  

(2.5)

The two horizontal vorticity components can be expressed as:

\[
\chi_x = - (\nabla \times \mathbf{v})_x = \frac{\partial v_z}{\partial y} - \frac{\partial v_y}{\partial z}.
\]  

(2.6)

Under equations (2.5)--(2.7), positive vorticity corresponds to clockwise rotation around orthogonal axes. If an axis is imagined in the direction of the outstretched thumb of a right hand that is clenched as a fist, then the sense of circulation associated with positive vorticity around that axis is parallel to the curved fingers of the hand. The angular momentum \( \mathbf{P} \) of a deforming fluid is:

\[
\mathbf{P} = \mathbf{I} \mathbf{\chi}.
\]  

(2.8)

where \( \mathbf{I} \) is its moment of inertia tensor.

Strain rate, being a symmetric \( 3 \times 3 \) tensor, has three orthogonal eigenvectors and three eigenvalues that indicate principal strain rates in three orthogonal directions. Given the incompressibility condition (equation 2.1), the sum of the three eigenvalues is zero. If one eigenvector is vertical, orthogonality allows the strain rate tensor to be separated into vertical and horizontal parts. The horizontal part of the strain rate tensor \( \mathbf{E}_h \), where:

\[
2\mathbf{E}_h = \begin{bmatrix}
\frac{\partial v_x}{\partial x} & \frac{\partial v_x}{\partial y} & \frac{\partial v_x}{\partial z} \\
\frac{\partial v_y}{\partial x} & \frac{\partial v_y}{\partial y} & \frac{\partial v_y}{\partial z} \\
\frac{\partial v_z}{\partial x} & \frac{\partial v_z}{\partial y} & \frac{\partial v_z}{\partial z}
\end{bmatrix}
\]  

(2.9)

indicates the pattern of local horizontal extension or shortening in the fluid. The two horizontal eigenvectors \( \varphi_1 \) and \( \varphi_2 \) of this tensor indicate the azimuths of the local principal axes of strain rate and the corresponding eigenvalues \( E_1 \) and \( E_2 \) strain rates in these directions. These
Figure 2. Major tectonic features in the NE Basin and Range Province, Neogene and Quaternary volcanic outcrops around Yellowstone and along the Snake River Plain are outlined. Northeastward movement of the Yellowstone upwelling mantle plume relative to the North American plate at \(3.5 - 4\,\text{cm yr}^{-1}\) indicated by ages of SRP volcanic rocks (e.g. Armstrong et al. 1975) implies that 10 Myr ago the active volcanic field was near Twin Falls, Idaho, 400 km SW of Yellowstone. The principal normal faults on both sides of the SRP, inferred as active from evidence of earthquakes or local geomorphology, are identified with tick marks on the hanging wall. Slip vector azimuths, where known or inferred, are indicated by short single arrows, and senses of strike-slip by small double arrows. Information is from Smith & Sbar (1974), Smith, Richins & Doser (1985), Eddington et al. (1987), Stickney & Bartholomew (1987), Pechmann et al. (1987) and Westaway & Smith (1989).

can be obtained by solving the equation:

\[
E_i \varphi_i = E_i \varphi_i \quad (i = 1, 2)
\]  

(2.10)

using standard methods. In an actively-extending region, eigenvector \(\varphi_i\), corresponding to principal extensional strain rate \(E_i\), can be assumed parallel to local slip vector azimuth on any major active normal fault in the brittle layer above the deforming 'fluid'. The horizontal part of \(\nabla \cdot \mathbf{v}\), \((\nabla \cdot \mathbf{v})_h\), where

\[
(\nabla \cdot \mathbf{v})_h = \frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y}
\]  

(2.11)

equals the trace of \(E_h\), and hence equals the sum of the eigenvalues of \(E_h\). It thus determines whether the deforming 'fluid' is, overall, extending or shortening in the horizontal plane. Incompressibility can still be maintained provided eigenvalue \(E_z\) of the vertical eigenvector, the strain rate for thinning, is equal in magnitude but opposite in sign to the sum of eigenvalues of \(E_h\).

Circulation that is divergent in the horizontal plane will thus try to extend the brittle layer above it. Blocks in the brittle layer may respond to vorticity in the underlying 'fluid' by rotating as rigid bodies. Vertical vorticity will drive rotation around vertical axes, and horizontal vorticity will drive block tilting or rotation around horizontal axes. With \(\mathbf{v}\) uniform beneath a region, then the brittle layer above it may be translated horizontally in the direction of \(\mathbf{v}\) without rotation or internal deformation; this translation may occur in addition to extension, shortening or rotation in an area
Figure 3. (a) A rectangular closed path C (1-2-3-4) in the x-y-plane, representing a plan view of a rectangular block of area A in the upper-crustal brittle layer that is rotating around a vertical axis due to the vertical vorticity in the underlying deforming fluid. Arrows show the direction in which the line integral in equation (2.14) is calculated. Its anticlockwise sense is consistent with the sign convention used by Westaway (1989~). (b) Schematic diagram indicating rotation of rectangular blocks above a deforming fluid with vertical vorticity $\zeta_z$ and extensional strain rate $\varepsilon$. See text for discussion. (c) Geometry of oblique faulting used to derive equation (2.23). AB is a unit oblique slip vector on a fault plane. AC is the direction of fault strike and CD is the horizontal direction perpendicular to AC. Angle DCB is rake, $\lambda$; and CAD is $\gamma$. Length AC is $\cos(\lambda)$ and length CB is $\sin(\lambda)$. Length CD is $\cos(\delta)$ or $\sin(\lambda)\cos(\delta)$. Length CD is also $AC\tan(\gamma)$ or $CA\cot(\lambda)\tan(\gamma)$. Equating the two expressions for length CD gives equation (2.26).

above fluid that is divergent, convergent or rotational. For example, a deforming zone of width $L$ containing uniform velocity gradients and with overall rate of right-lateral movement $S$, the vertical vorticity caused by this lateral boundary condition is:

$$\zeta_z = \frac{S}{L}$$

(McKenzie & Jackson 1983).

If upper-crustal blocks are considered as isolated floats with circular cross-section, then their rotation rate $\omega$ around vertical axes is:

$$\omega = \frac{d\varphi}{dt} = \frac{\zeta_z}{2}$$

(McKenzie & Jackson 1986). Lamb (1987) has investigated the more general problem of isolated floats with elliptical cross-section. These rotate in the same sense as $\zeta_z$, but at rate that depends on the orientation of the major axis of the float relative to $\nu$ and to the lateral boundaries of the deforming fluid. In contrast, if blocks are sufficiently large to be attached at both ends to rigid boundaries of a deforming zone, their rotation rate equals $\zeta_z$ (McKenzie & Jackson 1983) (equation 2.12).

This ‘floating block’ model has so far only been applied to investigate the large-scale patterns of deformation in a limited number of regions, most notably in central Greece (McKenzie & Jackson 1986). Central Greece comprises a zone $\sim 100 \times 100$ km in extent comprising active parallel oblique normal faults over which distributed extension is occurring. Striations on fault planes and slip vectors of recent major earthquakes indicate left-lateral slip on these faults. McKenzie & Jackson (1986) showed this left-lateral slip is consistent with clockwise rotation of blocks, implying clockwise vertical vorticity. Clockwise rotation up to $\sim 48^\circ$ over $\sim 5$ Myr has been observed in central Greece using paleomagnetism (Kissel et al. 1986), confirming this sense of vertical vorticity. These numbers are consistent with equation (2.12) with $L \sim 90$ km and $S \sim 25$ mm yr$^{-1}$, assuming that a right-lateral lateral displacement boundary condition across this deforming zone causes the clockwise vertical vorticity. However, other explanations for clockwise rotation are also possible (e.g. Westaway 1989c).

The theoretical argument that leads to the 2:1 ratio between vorticity and rotation rate for independent circular blocks as presented by McKenzie & Jackson (1983) and others is not directly applicable for zones of elongated interacting blocks, like the NE Basin and Range Province (Fig. 2), where a single set of parallel faults with roughly uniform strike forms block margins and appears to take up both extensional strain and rotation. Westaway (1989c) has recently investigated the geometrical requirements for deformation of a region to be taken up in the brittle layer in this manner. Figure 3 shows a rigid rectangular body with area $s = A$ that lies in the $x$-$y$ plane rotating around the $z$-axis at angular velocity $\omega$, on the surface of a deforming fluid moving locally at velocity $v$. Consider the line integral of the scalar product $v \cdot d\mathbf{l}$ around the margins of the rigid body $C$, where $d\mathbf{l}$ is a position vector corresponding to an
Because any block of arbitrary shape can be broken into components of vorticity and angular velocity, enabling rates the integral on the left hand side of equation (2.15) would of block tilting around horizontal axes to be related to different points within the deforming fluid. However, for the critical assumption is that the block is sufficiently small block provided they are free to rotate around vertical axes. This suggests that a useful infinitesimal element of length along C.

\[
\lim_{A \to 0} \left[ \frac{1}{A} \oint_C \mathbf{v} \cdot d\mathbf{l} \right] =
\lim_{dx, dy \to 0} \left[ -\omega y \, dx + \omega(x + dx)y - \omega(y + dy)(-dx) + \omega x(-dy) \right]
\]
\[
\frac{dx \, dy}{dx \, dy} = 2\omega.
\]  

(2.14)

Furthermore, from Stokes' theorem,

\[
\int_s \nabla \times \mathbf{v} \cdot d\mathbf{s} = \oint_C \mathbf{v} \cdot d\mathbf{l}.
\]  

(2.15)

Assuming \( s \) is sufficiently small for \( \nabla \times \mathbf{v} \) not to vary across it, the left-hand side of equation (2.15) equals \((\nabla \times \mathbf{v})_z A\). Combining equations (2.14) and (2.15) then gives:

\[
2\omega = (\nabla \times \mathbf{v})_z = \chi_z,
\]  

(2.16)

the familiar result for circular blocks (equation 2.13). Because any block of arbitrary shape can be broken into rectangular elements, this method is valid for all shapes of block provided they are free to rotate around vertical axes. The critical assumption is that the block is sufficiently small in comparison with spatial scales over which \( \mathbf{v} \) varies for \((\nabla \times \mathbf{v})_z \) to be constant across it. For blocks larger than this, the integral on the left hand side of equation (2.15) would need to be evaluated to enable \( \omega \) to be compared with \( \chi_z \) at different points within the deforming fluid. However, for sufficiently small blocks, equation (2.16) predicts the 2:1 relationship between vertical vorticity and angular velocity around the \( z \)-axis, regardless of shape. For Stokes' theorem to be valid, \( \mathbf{v} \) and all nine velocity gradient tensor elements must be continuous beneath \( s \). Because the 'fluid' beneath the brittle layer has finite viscosity, it cannot contain discontinuities in velocity, so this condition is likely always to be satisfied. The same reasoning can lead, under the same assumptions, to predicted 2:1 relationships between other components of vorticity and angular velocity, enabling rates of block tilting around horizontal axes to be related to horizontal components of vorticity. It is possible to speculate that the brittle layer beneath any region may break up during deformation on a scale such that \( \nabla \times \mathbf{v} \) not to vary substantially beneath any block, enabling the 2:1 relationship to be satisfied in all deforming regions but with different scales of blocks. This suggests that a useful measure of local characteristic horizontal scale of deformation within any region is provided by:

\[
C = \left[ \frac{\int_s (\nabla_h \mathbf{v}) \cdot (\nabla_h \mathbf{v}) \, dx \, dy}{\int_s (\nabla_h (\nabla \times \mathbf{v})) \cdot (\nabla_h (\nabla \times \mathbf{v})) \, dx \, dy} \right]^{1/2},
\]  

(2.17)

where both integrals are calculated over the same area \( s \) in the horizontal plane. \( \nabla_h \mathbf{v} \) and \( \nabla_h \mathbf{v} \) are the horizontal parts of \( \mathbf{v} \) and \( \nabla \mathbf{v} \), giving \( C \) dimensions of length.

Equation (2.16) will thus apply as a special case to actively-extending regions containing sets of parallel faults bounding blocks that are rectangular in plan view and arranged like tilted dominoes in cross-section (Fig. 3b). In this case one horizontal strain rate tensor eigenvalue \( E_1 \) is the principal horizontal extensional strain rate, and the other \( E_2 \) at azimuth perpendicular to \( E_1 \) will be zero.

Consider faults arranged as in Fig. 3(b) responding to a horizontal pattern of deformation in the underlying 'fluid' characterised by \( \chi_z \) and \( E_1 \), defining blocks with width \( D \) perpendicular to strike. Rotation rate around vertical axes will follow equation (2.16). If \( C(t) \) is distance between mid-points of adjacent blocks in the direction of \( E_1 \), the direction in which extension is occurring, and this direction is at angle \( \gamma \) to the fault strike \( \phi \) at time \( t \), then:

\[
\frac{\partial C}{\partial t} = E_1 C.
\]  

(2.18)

Given that adjacent blocks are rotating around vertical axes, and distance between block centres is increasing due to the component of normal slip, at any time the strike-slip rate \( S(t) \) on each fault is:

\[
S = \frac{X^2}{2} C \sin (\gamma).
\]  

(2.19)

Given also that \( 2\omega = \chi_z \) and separation of block centres perpendicular to fault strike is \( D \), then equation (2.19) can be simplified to:

\[
S = \omega D.
\]  

(2.20)

Finally, \( S \) must also equal the component of velocity of separation of adjacent blocks parallel to the instantaneous direction of fault strike:

\[
S = \frac{\partial C}{\partial \gamma} \cos (\gamma).
\]  

(2.21)

Deformation of any extending region where extension is taken up by a single set of equi-spaced parallel faults in the brittle layer must instantaneously satisfy equations (2.13) and (2.18)–(2.21) at any time \( t \). Note that the along-strike length \( L \) of blocks does not appear in any equation, suggesting that solutions are independent of \( L \). Extension factor \( \beta \) can be calculated as:

\[
\beta = \frac{C}{D} \sin (\gamma).
\]  

(2.22)

During oblique extension, blocks will also tilt around horizontal axes, and faults will rotate to progressively less steep dips, as described by Jackson & McKenzie (1983) and others. A measure \( H \) of the relative strength of vertical vorticity and extensional strain rate can be defined as:

\[
H = \frac{\chi_z}{2E_1}.
\]  

(2.23)

\( H \) is dimensionless when \( \chi_z \) and \( E_1 \) are expressed in the same units. The factor 2 in the denominator is inserted to approximately normalize \( H \), given that the off-diagonal element of the horizontal strain rate tensor, \( E_{xy} \), which will be comparable in size to the eigenvalue \( E_1 \) is from equation (2.9):

\[
2E_{xy} = \frac{\partial u_x}{\partial y} + \frac{\partial u_y}{\partial x},
\]  

(2.24)

whereas the definition of \( \chi_z \) (equation 2.5) contains no factor 2.

Combining equations (2.17), (2.19), (2.21) and (2.23)
as a necessary condition to be satisfied at all times for a region to take up deformation in the underlying 'fluid' on a single set of parallel oblique faults. If slip vectors on all faults lie in the same vertical plane as the eigenvector for \( \lambda \), then \( \gamma \) is the horizontal projection of rake \( \lambda \):

\[
\tan(\gamma) = \tan(\lambda) \cos(\delta)
\]

(2.26)

(see Fig. 2c) where \( \delta \) is dip of the faults. Values of \( \gamma \), \( \delta \) and \( \lambda \) consistent with instantaneous deformation being taken up on a single set of parallel faults are listed in Table 1 for \( 0 > \lambda > -90^\circ \), giving solutions consistent with clockwise rotation around vertical axes and left-lateral slip on faults.

\( H \), \( \chi_x \) and \( E_1 \) may all vary with time, but if a single set of faults takes up both finite strain and finite rotation in a region, they are interrelated. Westaway (1989c) has determined solutions for finite extension with variable \( H \) and constant \( \gamma \) but variable \( \chi_x \) and \( E_1 \) for several actively-extending regions.

\( H \) is a useful parameter and deserves a name. I suggest Holmes number is appropriate, commemorating Arthur Holmes' early investigations of mantle dynamics and thermal structure (e.g. Holmes 1928). Given that deformation is in three dimensions, \( H \) as already defined can be regarded as vertical Holmes number, \( H_e \). With horizontal strain rate eigenvalue \( E_2 \) zero, horizontal Holmes numbers \( H_x = \chi_x/(2E_3) \) and \( H_y = \chi_y/(2E_3) \) can be defined, where \( E_3 \) is crustal thinning strain rate.

Although a given \( H_e \) beneath a zone of parallel faults requires a particular oblique slip vector on all faults in the zone to take up both extensional strain rate and rotation around vertical axes, it is worth pointing out that oblique slip on faults in general does not necessarily require rotation around vertical axes. Faults may slip obliquely when \( \lambda \) is zero. If \( \chi_x \) is unknown a priori, but \( E_1 \) can be estimated from slip rate on faults and \( H_e \) from \( \gamma \), an estimate for \( \chi_x \) can be made using equation (2.23) as \( \langle \chi_x \rangle = 2E_1H_e \). The true value of \( \chi_x \) will be somewhere between zero and \( \langle \chi_x \rangle \).

### Table 1. Conditions for deformation of the brittle layer by slip on a single set of parallel faults

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</table>

The Lost River fault in the central Idaho seismic zone is represented approximately by: \( \lambda = 60^\circ; \gamma = 51^\circ; H_e = 0.82 \). The East Cache fault in the Intermountain seismic belt is represented approximately by: \( \lambda = 70^\circ; \gamma = 63^\circ; H_e = 0.51 \).

Provided the nodal plane ambiguity for earthquake focal mechanisms on a zone of parallel faults can be resolved, for example using surface faulting, an asymmetric seismic moment tensor \( M^* \) can be defined (Molnar 1983; Jackson & McKenzie 1988), where:

\[
M^* = M_{0}\mu\nu
\]  

(2.26)

or

\[
M^*_u = M_{0}u_i n_j
\]  

(2.27)

where \( M_0 \) is scalar seismic moment, and \( \mu \) and \( \nu \) are unit vectors parallel to the co-seismic slip vector and normal to the fault plane. Summation of asymmetric moment tensors for earthquakes within a zone of parallel faults gives the full velocity gradient tensor [(see equation (22) of Jackson & McKenzie (1988)) in a frame of reference that is fixed to the faults, just as summation of standard symmetric moment tensors gives the strain rate tensor (Kostrov 1974), the symmetric part of the velocity gradient tensor (equation 2.2)]. Westaway (1989c) has shown that in a zone of parallel faults the brittle layer can only take up finite extensional strain and rotation at constant strain rate and rotation rate provided the principal extensional strain rate azimuth rotates around a vertical axis at the same rate as all the faults. In these circumstances it may be a reasonable alternative to analyse finite deformation within any such zone in a frame of reference that rotates around a vertical axis at the same rate as all the faults rather than a reference frame that is fixed. Provided enough earthquakes occur on a set of parallel faults to allow \( M^* \) to be determined, it can be used to resolve the full velocity gradient tensor beneath the set of faults. This method cannot be used to obtain the overall deformation rate across any deforming zone bounded by rigid stable regions, because, as Jackson & McKenzie (1988) point out, faults at the boundaries between a deforming zone and its stable surroundings inevitably cannot rotate at the same rate as faults within the zone, and the necessary conditions for \( M^* \) to be meaningful, that all faults are parallel and rotate at the same rate, cannot be satisfied. However, this method can work in principal for deformation on a set of parallel faults within a deforming zone, such as the Central Idaho seismic zone in central
Idaho, or the Intermountain seismic belt in northern Utah and southern Idaho (Fig. 2). Unfortunately, as is shown later, too few large earthquakes have occurred in these areas in recent decades for this method to work in practice.

3 OBSERVED DEFORMATION IN THE NE BASIN AND RANGE PROVINCE

The Yellowstone volcanic region in NW Wyoming has been the object of intense geophysical investigation for many years. One popular viewpoint (e.g. Morgan 1972; Matthews & Anderson 1973; Smith & Sbar 1974; and many more recent articles) regards Yellowstone and the Snake River Plain (SRP) as the surface expression of a ‘hot spot’ or upwelling plume from the deep mantle. Three main pieces of geophysical evidence are consistent with plume theory. First, the sustained very high surface heat flow can only be explained if a concentrated local source of heat exists at depth. Second, the broad geoid and gravity highs around Yellowstone (e.g. Wagner et al. 1977; Taylor et al. 1983) are comparable to those observed above upwelling plumes elsewhere (e.g. McKenzie et al. 1980). Third, the Yellowstone plume appears to have been stationary in the same reference frame as the better-documented Hawaiian plume for more than 10 Myr (e.g. Smith & Sbar 1974), suggesting that both may be similar features. If Yellowstone is regarded as the present-day position of an upwelling plume, then the SRP, which extends SW of Yellowstone across the Basin and Range Province for more than 400 km, can be interpreted as a trail left behind by this plume as the North American plate has moved across it. Age of basalts in the SRP increases linearly from Yellowstone (Armstrong, Leeman & Malde 1975), suggesting that the North American plate has been moving SW with respect to the inferred plume at 3.5–4 cm yr\(^{-1}\) for the last 10 Myr. However, other explanations for this region have also been suggested (e.g. Thompson 1977; Dzurisin & Yamashita 1987). In particular, petrological studies (e.g. Thompson 1975, 1977) suggest a relatively shallow mantle source for the SRP volcanic rocks, not a deep mantle source within a plume. Although the geophysical evidence in support of the existence of a Yellowstone plume appears unequivocal, this large body of petrological work cannot simply be ignored.

The Borah Peak, Idaho, earthquake of 1983 October 28 (\(M_\text{L} 7.3; M_\text{w} 33 \times 10^{18} \text{Nm}\)) has drawn attention to seismic activity in the NE Basin and Range Province. As has already been pointed out by Scott, Pierce & Hait (1985), much of this activity is concentrated (Fig. 2) in a V-shaped pattern. Active normal faulting predominates along both branches of this V, indicating that extension of the brittle upper crust is occurring, with slip rates on individual major faults ~1 mm yr\(^{-1}\). One branch trends eastward across central Idaho and SW Montana, and the other, which includes the well-known Wasatch Fault, trends SSW through western Wyoming, SE Idaho and northern Utah. The two branches converge at Yellowstone, where the SW trend of the SRP bisects the two seismically active zones. Seismicity in these zones since 1900 has been sparse (see, e.g. Fig. 2 of Eddington, Smith & Renggli 1987). The central Idaho seismic zone has experienced six earthquakes with \(M_\text{L} > 6\) including two, Hebgen Lake and Borah Peak, with \(M_\text{L} > 7\). The Intermountain seismic belt has experienced only three events with \(M_\text{L} > 6\). Five of these nine events occurred before the late 1950s when global coverage of seismograph stations began to improve, facilitating detailed earthquake source studies, and their source orientations and seismic moments are not known with confidence. Even waveform inversion results (e.g. Doser 1989) for some older events are very poorly constrained and contribute nothing useful to any discussion of the active tectonics of the region. Some published maps indicate that earthquake activity persists in a northward continuation of the Intermountain seismic belt north of Yellowstone in western Montana. Although some large historical earthquakes have occurred in this region, and some late Quaternary surface faulting has been mapped (e.g. O’Neill & Lopez 1985), local deformation rate appears to be much less than in the zones west and south of Yellowstone (Scott et al. 1985). Similarly, although the Wasatch fault has been identified for more than 100 km south of Salt Lake City, it appears much less active than further north; Eddington et al. (1987) suggest average slip rate between ~0.1 and 0.3 mm yr\(^{-1}\) south of Salt Lake City.

Many studies have investigated in detail the active normal faults in the brittle upper crust of the NE Basin and Range Province. Active normal faults close to Yellowstone are oriented approximately radially to the active volcanic centre, with slip vectors approximately tangential to Yellowstone. Examples include the fault that moved in the Yellowstone Park earthquake (\(M_\text{L} 5.9\)) of 1975 June 30 (e.g. Pitt, Weaver & Spence 1979), north of the volcanic centre, during which east-west extension occurred; the much larger Hebgen Lake earthquake of 1959 August 18 (\(M_\text{L} 7.5\)), west of Yellowstone, which involved approximately North-South extension (e.g. Doser 1985); the Teton fault, south of Yellowstone, ~60 km long with ~10 km of total normal throw and more than 10 m of Holocene throw involving E–W extension (Susong, Smith & Bruhn 1987). Its Holocene slip rate is thus ~1 mm yr\(^{-1}\), a value typical for other major faults in the region also (e.g. Schwartz & Coppersmith 1984).

The two seismically active zones further west and south of Yellowstone have developed on opposite sides of the SRP. The northern zone, sometimes called the Centennial Tectonic Belt or the Idaho Seismic Zone, comprises three major normal faults bounding the west faces of the Lost River, Lemhi and Beaverhead ranges in central Idaho, together with several smaller faults in SW Montana (e.g. McKenzie & Jackson 1986; Stickney & Bartholomew 1987) (Fig. 2). The one major historical earthquake in this zone, Borah Peak, occurred on part of the westernmost of these faults that bounds the Lost River range. This ruptured a WSW-dipping normal fault segment with substantial left-lateral strike-slip (e.g. Crone & Machette 1984) (strike 150°, dip 45°, rake ~60°). Seismological and geodetic evidence indicates this fault is planar with constant 45° dip between the Earth’s surface and the 13–16 km rupture nucleation depth (e.g. Doser & Smith 1985; Stein & Barrientos 1985). Slip vector azimuth was ~200°, with maximum normal slip ~2 m and left-lateral strike-slip ~1 m. Although the 2–3 m slip in 1983 involved ~1 to 5 ratio of footwall uplift to hanging wall subsidence (e.g. Stein & Barrientos 1985), relative to the present day ground surface in the hanging wall long-term footwall uplift and
hanging wall subsidence are about equal (Fig. 2). The Lemhi and Beaverhead faults have 5 m high Holocene scarps (Stickney & Bartholomew 1987), implying average Holocene slip rate ~0.5 mm yr⁻¹. McKenzie & Jackson (1986) have suggested that the left-lateral slip on the Lost River fault may imply clockwise rotation of faults and blocks in the brittle layer within this zone around vertical axes, which may be caused by clockwise vertical vorticity in the underlying 'fluid'. Figure 4 shows a cross-section trending N 70°E across the central Idaho seismic zone. Although the planar geometry of the Lost River fault is clear (Stein & Barrientos 1985), no direct evidence exists for the dip at depth of the other two major range-bounding faults, because no major historical earthquake has occurred on either of them, so the same dip and depth limit have been assumed. Assuming that the early Tertiary volcanic unit buried beneath the hanging wall and projected above the footwall of the Lost River fault in Fig. 4 was continuous before extension began, throw on this fault is ~7 km (e.g. Stein & Barrientos 1985).

Assuming this throw accumulated at uniform 1–1.4 mm yr⁻¹ average rate (Stickney & Bartholomew 1987; Eddington et al. 1987), then the age of the Lost River fault is ~5–7 Myr. The estimated rate of movement of the North American plate relative to the upwelling plume suggests that the plume was close to the SE end of this fault ~5 Myr ago, suggesting that active faulting may have begun when the Yellowstone plume passed by it. Assuming proportions of normal and left-lateral slip in the Borah Peak earthquake are typical of the earlier history of the Lost River Fault also, they imply average Holocene strike-slip rate a substantial fraction of 1 mm yr⁻¹ and total strike-slip offset up to 4 km.

The southern seismically-active zone, the Intermountain seismic belt, extends SW from Yellowstone to the vicinity of Salt Lake City in northern Utah, comprising many normal faults that have been active in Holocene time (Fig. 5), most of which strike S and dip W. It includes the East Great Salt Lake fault of Pechmann et al. (1987), with estimated slip-rate ~0.5 mm yr⁻¹; the normal fault at the margin of Hansel Valley, Utah, that moved in the earthquake of 1934.

Figure 5. W–E cross-section in a vertical plane along 41° 54' N between the Wasatch Fault and Bear Lake, from C to D in Fig. 2, showing schematically the pattern of normal faults in the upper crust. Construction of this section, described in detail by Westaway & Smith (1989), is based on results of a study of the Cache Valley earthquake of 1962 August 30 that indicate fault plane dipping W at ~40°. As in Fig. 4, the same orientation is assumed for other faults that have not moved in recent earthquakes. E-dipping Tertiary beds outcropping at the western margin of the Cache Valley (Davis 1985) and buried beneath it are shown, identified by depth-converting a seismic section (Smith & Bruhn 1984; fig. 9). Depth of Bear Lake is schematic; 1.5 km thickness of Neogene sediments in the hanging wall of the Bear Lake fault is from Dixon (1979).
March 12 \( (M_s 6.6; M_w 8 \times 10^{18} \text{Nm}; \text{Doser} \& \text{Smith} 1982) \); the northern part of the Wasatch fault, with estimated slip rate \(-1 \text{mm yr}^{-1}\) (e.g. Swan, Schwartz \& Cluff 1980; Schwartz \& Coppersmith 1984); the East Cache fault; the Temple Ridge fault, which Westaway \& Smith (1989) have suggested moved in the Cache Valley earthquake of 1962 August 30 \( (M_s 5.7; M_w 0.3 \times 10^{18} \text{Nm}) \); the NW-dipping fault that moved in the Pocatello Valley, Idaho, earthquake of 1975 March 28 \( (M_s -6.0; M_w 0.3 \times 10^{18} \text{Nm}; \text{Bache, Lambert} \& \text{Barker} 1980) \); and faults bounding the eastern edges of Bear Lake in northern Utah and southern Idaho and Star Valley in western Wyoming (e.g. Smith \& Sbar 1974; Dixon 1978; Oriel \& Platt 1980) (Fig. 5). Much of the extension of this zone appears to have occurred on this set of sub-parallel faults.

The East Cache fault has one of the greatest topographic offsets between the hanging wall and the footwall in this zone. It is not continuous, comprising several segments each offset to the left (e.g. Westaway \& Smith 1989), consistent with a component of right-lateral slip (e.g. Ambraseys \& Tchalenko 1972) and in agreement with the deduction by Westaway \& Smith (1989) of a component of right-lateral slip on the adjacent sub-parallel Temple Ridge Fault in the Cache Valley earthquake (strike 193°, dip 43°, rake -101°). Smith \& Bruhn (1984) suggested that the East Cache fault may flatten to sub-horizontal dip at only a few km depth. However, the overall geomorphology of the region around the East Cache fault, including the northern part of the Wasatch fault to the west and the Bear Lake fault to the east, resembles the tilted 'domino' pattern observed in other regions of continental extension, such as the margins of the Red Sea (Morton \& Black 1975) and central Greece (Jackson \& McKenzie 1983). In these regions extension occurs on normal faults that are planar from mid-crustal depths, close to the depth-limit of seismicity or 'brittle-ductile transition' near which large earthquakes nucleate (e.g. Sibson 1982) to the earth's surface. As extension proceeds, these faults and the blocks between them tilt progressively, decreasing dips of the faults. On account of this tilting, sediments deposited in the hanging wall of any fault tilt progressively towards the fault as they are buried beneath younger sediments.

Figure 6 shows schematically two faults that formed with initial dips \( \delta_0 \) separated in a horizontal direction by distance \( D_0 \). After extension the faults dip at \( \delta_1 \) and are separated by distance \( D_1 \). Assuming blocks between faults do not deform internally, separation of the faults perpendicular to their planes remains constant. This condition means that:

\[
D_0 \sin (\delta_0) = D_1 \sin (\delta_1) \quad (3.1)
\]

and hence:

\[
\beta = D_1/D_0 = \sin (\delta_1)/\sin (\delta_0), \quad (3.2)
\]

where \( \beta \) is extension factor. Throw \( T \), or distance \( P_1-S_1 \), is:

\[
T = D_1 \cos (\delta_1) - D_0 \cos (\delta_0) \quad (3.3)
\]

or

\[
T = D_0[\beta \cos (\delta_1) - \cos (\delta_0)]. \quad (3.4)
\]

Extension on the East Cache fault can be investigated using equations (3.2)–(3.4). Suppose it now has dip \( \delta_1 40^\circ \), similar to the W-dipping nodal plane in the 1962 earthquake. Suppose also its initial dip \( \delta_0 \) was 60°, a typical value for active normal faults in extending regions that have not yet extended much, such as southern and central Italy (e.g. Westaway \& Jackson 1987; Westaway et al. 1989). This implies extension factor \( \beta \approx 1.35 \) on the East Cache fault. Horizontal separation of the East Cache and Wasatch faults is now \(-20 \text{km}; \text{indicating} \sim 15 \text{km} \text{original separation.} \text{Throw} \ T \text{on the East Cache fault is thus} \sim 8 \text{km.} \)

A second estimate of throw on the East Cache fault can be made using thickness of Tertiary sediments in its hanging wall. Smith \& Bruhn (1984; fig. 7) showed a seismic reflection record section crossing this fault \sim 10 \text{km} \text{S of the} \text{line in Fig. 5}, \text{indicating beds dipping towards} \text{it to} \text{echo time} 3 \text{s}. Depth-converting this section approximately using 2–3 \text{km s}^{-1} \text{seismic velocity in} \text{these beds indicates that} \text{the} \text{deepest and oldest beds} \text{dip towards} \text{the fault at} 20°–25°, \text{and suggest that} \text{the} \text{top of basement} \text{beneath the} \text{hanging wall} \text{sediments} \text{is not shallower than} \text{3–4.5 km.} \text{The} \text{dipping beds} \text{visible} \text{on} \text{this seismic section} \text{are likely to be} \text{of the} \text{'Salt Lake Group'} \text{of} \text{upper Oligocene to Pliocene elastic} \text{sediments} (\text{Davis} 1985). \text{Although part of this formation} \text{outcrops} \text{along} \text{the western margin of the Cache Valley, also} \text{dipping} \text{E at} \text{~20°}, \text{it} \text{is obscured beneath} \text{most of the} \text{valley} \text{by} \text{Pleistocene deposits of} \text{glacial Lake Bonneville.} \text{This} 20° \text{dip is consistent} \text{with that expected if the} \text{East Cache fault} \text{has tilted during extension from} \text{60° to} \text{40°} \text{.} \text{Paleozoic rocks in its footwall} \text{dip} \text{E at} \sim 20° \text{also, suggesting that they were also sub-horizontal before} \text{Neogene extension began.}
Elevation change across the East Cache fault is ~1.5 km, as well as the 3–4.5 km minimum thickness of Neogene sediments beneath its hanging wall. This indicates at least 4.5–6 km overall vertical offset. With dip of the East Cache Fault now 40°, its throw is thus at least 7–9 km. The apparent 20° change in dip of the East Cache fault over ~10 Myr suggests that it and the surrounding blocks are rotating around horizontal axes parallel to its strike at ~2° Myr⁻¹. For comparison, the same analysis leads to ~15° rotation around a sub-parallel horizontal axis over 5 Myr around the Lost River Fault, at ~3° Myr⁻¹ average rate.

A third estimate of throw on the East Cache Fault can be made using the modelling of gravity anomaly changes across it by Zoback (1983). This study suggests vertical offset up to 3.8 km, corresponding, again assuming 40° dip, to 6 km throw. Thus all three methods, given the stated assumptions, predict 6–9 km throw on the East Cache Fault. If its age is 10 Myr, then its average slip rate is close to 1 mm yr⁻¹, similar to the rate on the adjacent northern Wasatch fault (e.g. Swan et al. 1980). This slip rate may be achieved by, say, a magnitude 7 earthquake similar to Borah Peak producing ~2–3 m throw on each segment once every ~3000 yr. Given the stated assumptions, the evidence suggests that the East Cache Fault and other sub-parallel normal faults in the same actively-extending zone are similar to the major normal faults on the other side of the SRP, and are major planar features reaching to mid-crustal depths on which much of the extension of the Intermountain seismic belt has taken place.

The Pocatello Valley earthquake of 1975 March 28 occurred ~50 km NW of the Cache Valley event, outside the zone of parallel faults in Fig. 5. Aftershock locations form a zone dipping NW at ~40° (Bache et al. 1980). The mainshock focal mechanism determined by Bache et al. (1980) has one NW-dipping nodal plane with similar orientation (strike 225°, dip 39°, rake ~53°). This is most likely the fault plane, and gives slip vector azimuth 271° similar to the Cache Valley event. This and the Cache Valley event are the only normal-faulting earthquakes with $M_A > 5.5$ in the Intermountain seismic belt that have been sufficiently well recorded to contribute reliable co-seismic slip vectors.

Many of the normal faults now active in both seismic zones are reactivated reverse faults that formed in late Cretaceous and early Tertiary time during an episode of large-scale crustal shortening that created the Overthrust Belt of SE Idaho and western Wyoming (e.g. Armstrong & Oriel 1965; Ruppel & Lopez 1984). Surface traces of some of these faults lie along the boundaries of early Tertiary thrust sheets. This has led some people (e.g. Smith & Bruhn 1984) to suggest that some active normal faults in this zone may be steep at the earth's surface but flatten at a few km depth. My alternative interpretation suggests instead that these faults may reach to mid-crustal depths of ~15 km, with approximately planar dip, on both sides of the SRP. The limited evidence from northern Utah and southern Idaho suggests that individual parallel active normal faults in this region move with a component of right-lateral strike–slip, which may indicate the brittle layer is responding to anticlockwise vertical vorticity in the underlying 'fluid'. Slip vector azimuths ~295° and ~271° in the Cache Valley and Pocatello Valley earthquakes were inward towards the SRP at ~40 to 65°. Slip vector azimuth at Borah Peak was ~200° or towards the SRP at ~30°. These angles average to inward orientation at ~45°.

The Teton fault in NW Wyoming is the closest major active normal fault to Yellowstone in the Intermountain seismic belt. Heights of its Holocene scarps (Susong et al. 1987) indicate average normal slip rate 0.74 mm yr⁻¹. Substantial left-lateral slip on these scarps is consistent with en-échelon stepping to the right of adjacent segments of this fault and, given its 010° strike, indicate slip vector azimuth ~60°. This is substantially different from slip vector azimuths typical further south (Table 2), suggesting principal extensional strain rate azimuth changes across the region in between. Left-lateral slip on the Teton fault may imply local clockwise vertical vorticity, suggesting this vorticity component reverses sense from clockwise south of and near Yellowstone to anticlockwise further away. Eddington et al. (1987) have summed seismic moment tensors from microearthquakes around the Teton fault for comparison. These give a strain rate tensor that implies average local slip rate only 0.07 mm yr⁻¹, with extension direction different from that inferred from Holocene scarps. This confirms that only the largest earthquakes, with $M_A > 5.5$, have source orientations representative of the overall regional deformation pattern.

Faulting in the Hebgen Lake earthquake of 1959 August 18 ($M_A 7.5$; $M_s 100 \times 10^{18}$ Nm) (Doser 1985) also warrants comment. Several fault segments ruptured, each up to ~20 km long, forming scarps up to ~10 m high (Myers & Hamilton 1964). No clear evidence of sense of any oblique slip was observed on any segment of scarp. The largest throw occurred on the Red Canyon fault, with strongly curved surface trace; strike varied from ~100° to ~60° between its eastern and western ends. Doser (1985) matched observed teleseismic waveforms with a planar source with strike 086°, dip 50° and rake ~130°, implying a component of right-lateral strike–slip with slip vector azimuth between 220° and 230°. The strike determined is thus an average along the curved fault plane, a consequence of use of a modelling procedure that could model planar fault rupture only. Westaway (1989b) has investigated the sense of relative motion between the footwall and the hanging wall expected under the assumption that all parts of this scarp had the same slip vector. The trend of this common slip vector depends on the dip assumed at different parts of the scarp, and may range between ~170° (in which case the E end of the scarp had a small component of left-lateral slip, and the W end a small component of right-lateral slip) to ~240° (in which case the E end of the scarp had a small component of right-lateral slip, and the W end a substantially larger component of right-lateral slip). It is likely that this event involved a small amount of right-lateral strike–slip overall, with slip vector azimuth ~220°.

Figure 7 shows a vertically exaggerated section trending NW–SE, perpendicular to the axis of the SRP. It shows the topography of the Lost River Range, the footwall of the Lost River Fault, NW of the SRP, and the corresponding topography of ranges on its SE side that are an in-line northward continuation of the footwall of the Bear Lake fault. The envelope of the summits of mountain ranges on both sides of SRP dips towards it, as does the elevation of the ground surface in the hanging wall valleys between
Table 2. Values of \( r, \theta, x \) and \( y \), using the coordinate system in Fig. 8, for the extremities of some of the principal active normal faults in the actively-extending zones south and west of Yellowstone.

<table>
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<th>( r ) (km)</th>
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<th>( x ) (km)</th>
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<td>N</td>
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<td>E Wasatch S</td>
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<td>F EGSLF/Hansel S</td>
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The EGSLF is the East Great Salt Lake Fault of Pechmann et al. (1987), which reaches the earth's surface along the western shores of Antelope and Fremont Islands within the Great Salt Lake. \( \phi \) is average strike of each fault; \( \alpha \) is slip vector azimuth; \( \psi \) is the difference between \( \alpha \) and the 230° trend of the \( x \)-axis; and \( \gamma \) and \( H_z \) are defined in equations (2.26) and (2.23).

The average topographic gradient is \(-1^\circ\) towards the SRP.

Before attempting to model deformation of the NE Basin and Range Province, it is necessary to use the available field and seismological evidence to estimate deformation rates. Three independent methods exist for estimating strain rate: by calculation from observed slip rates inferred from Holocene scarp heights; via summation of seismic moment tensors; and by direct measurement using geodesy. Eddington et al. (1987) have shown that observed Holocene scarp heights indicate extensional strain rates \( E_1 \) up to \(-10^{-15} \text{ s}^{-1}\). This corresponds to \(-0.75 \text{ mm yr}^{-1}\) and fault dip 45° on a fault bounding a block 25 km wide. As mentioned in Section 2, for zones of parallel faults one may either sum standard symmetric moment tensors to obtain the strain rate tensor (Kostrov 1974), or sum asymmetric moment tensors to obtain the velocity gradient tensor (Molnar 1983). Although these methods will work well in relatively active parts of the Mediterranean region [e.g. Jackson & McKenzie (1988) who...
applied to Kostrov (1974) method], the historical record of earthquakes in the NE Basin and Range Province is too short and the observed seismicity too sparse for either method to work well there. They result in localities where a major earthquake has recently occurred being assigned strain rate comparable with those deduced from scarp heights (Eddington et al. 1987), but other localities that are apparently equivalent but have not recently had a major earthquake being assigned much lower strain rates, as illustrated by the Teton fault example. The limited record of geodetic observations confirms strain rates are $\sim 10^{-15}$ s$^{-1}$ in parts of the NE Basin and Range province (Eddington et al. 1987).

Using Holocene scarp heights to deduce average slip and strain rates gives uncertain results, because the $\sim 3000$ yr or more interval between relatively infrequent magnitude $\sim 7$ earthquakes that form the scarps is a substantial fraction of the duration of Holocene time, $t_{hh}$. If each such earthquake produces a scarp of height $h$, and $n$ events have occurred, observed scarp height is $nh$ and the simplest estimate of the Holocene slip rate is $u_n$, where:

$$u_n = nh/t_{hh}. \tag{3.5}$$

However, if no historical earthquake has occurred on a fault, the time since the last earthquake is unknown. It may be as little as $\sim 100$ yr or as much as $\sim t_{hh}/n$. If another earthquake were to happen tomorrow, the average Holocene slip rate would then be revised to $u_{n+1}$ where:

$$u_{n+1} = (n+1)h/t_{hh}. \tag{3.6}$$

Because many scarps in the NE Basin and Range Province have $n \sim 2$ (e.g. Stickney & Bartholomew 1987), the $u_{n+1}$ slip rate estimate will be typically 50 per cent larger than the $u_n$ estimate. Throughout this article slip rates follow the $u_n$ estimate, although this may underestimate true rates. This problem is complicated further both because recurrence intervals of earthquakes of a given size are intrinsically variable, and because any scarp may be produced by earthquakes of a range of sizes that recur at different intervals.

Rotation rates on faults within parallel sets can be estimated using equation (2.19) provided the strike-slip component of slip rate, $S$, is known. For example, on the Lost River fault $S$ is $\sim 1$ mm yr$^{-1}$ and $D \sim 30$ km, making $\omega \sim 10^{-15}$ s$^{-1}$ or $2 \times 10^6$ Myr$^{-1}$ clockwise. Over the 5 Myr age of this fault estimated from its throw, maximum likely total rotation around a vertical axis is $\sim 10^6$, assuming the fault has rotated at the same rate and in the same sense ever since it formed. Using equation (2.13), present-day local vertical vorticity is $\sim 4$ Myr$^{-1}$.

Thus, to summarize the available observations, the following rates and senses of deformation appear typical for major active normal faults in the NE Basin and Range Province: normal slip rate $\sim 1$ mm yr$^{-1}$; strike slip rate up to 1 mm yr$^{-1}$; extensional strain rate $10^{-15}$ s$^{-1}$; rotation rate $2 \times 10^6$ Myr$^{-1}$ around vertical axes and horizontal axes perpendicular to the SRP. Slip vectors in both seismic zones are oriented inwards at $\sim 45^\circ$ towards the SW axis of the SRP. Expected rotation around vertical axes is clockwise in the Central Idaho seismic zone and anticlockwise in the Intermountain seismic belt.

Before proceeding further it is necessary also to investigate whether the assumption that the bulk of the Neogene deformation in the NE Basin and Range Province has occurred on the limited number of major faults in Fig. 2 is likely to be valid. Many other smaller faults exist in the region, and it is worthwhile to check whether their cumulative throw could be substantial in comparison with the throw on the smaller number of larger faults that form parallel sets. If the bulk of the deformation in each seismically active zone is on a single set of major faults, then equations (2.22) to (2.25) can be applied. For the part of the central Idaho seismic zone comprising the en-échelon Lost River, Lemhi and Beaverhead faults, the Borah Peak slip vector suggests $\gamma \sim -50^\circ$. For the part of the Intermountain seismic belt between the Wasatch and Bear Lake faults, the Pocatello Valley and Cache Valley slip vectors suggest $\gamma \sim 110^\circ$ (Table 2). Given faults dip at $\sim 45^\circ$ and are approximately equi-saced, equation (2.24) predicts $H_z = -0.3$ to $-0.4$ for the East Cache fault and $\sim 0.8$ for the Lost River fault; highlighted in bold print in Table 1, or $\sim 0.6$ on average. For the Lost River fault this estimate is in reasonable agreement with the value calculated from independent estimates of $\chi_z = \gamma^2 - 4\pi$ Myr$^{-1}$ and $E_1 = 1.4 \times 10^{-15}$ s$^{-1}$, giving $H_z = 2\chi_z/(E_1) \equiv 0.8$. For the zone containing the Lost River fault at least, observed slip vector azimuth is consistent with that expected under the assumption that deformation is predominantly on a single set of parallel faults.

### 4 SUMMARY OF PREVIOUS EXPLANATIONS OF DEFORMATION IN THE NE BASIN AND RANGE PROVINCE

Many people have suggested explanations for extension and the associated magmatism in the Basin and Range Province over the last 50 Myr (e.g. Zoback, Anderson & Thompson 1981; Gough 1984; Sonder et al. 1987; Wernicke et al. 1987). These include models that link extension to changes in plate motions that affect the lateral boundary conditions to the region (e.g. Zoback et al. 1981) or with mantle upwelling, either associated with Yellowstone (e.g. Smith & Sbar 1974) or on a larger scale (e.g. Matthews & Anderson 1973; Gough 1984). More recently, Sonder et al. (1987) have shown that many features in the timing and distribution of extension can be explained as consequences of the lithosphere having been overthickened before extension began, due to earlier shortening having occurred. In this model much of the Tertiary magmatism of the Basin and Range province can be explained by adiabatic decompression of the lower lithosphere during extension. Geological observations of the timing and distribution of magmatism appear not to be related in any simple way to boundary conditions associated with plate motions. However, relaxation of boundary conditions at the edges of the region, which presumably reflected changing patterns of plate motion, was probably necessary for extension to begin $\sim 50$ Myr ago (Wernicke et al. 1987).

In the NE Basin and Range Province and on the time-scale of the last 10 Myr the observations made by McKenzie & Jackson (1986) that suggest clockwise vertical vorticity beneath the Central Idaho Seismic Zone have already been summarized in Section 3. However, they did
not discuss any physical process that could cause this vorticity. For the last 10 Myr, deformation in the NE Basin and Range Province and the associated normal-faulting seismicity appears to be unrelated to any lateral boundary condition related to plate motions (e.g. Smith & Sbar 1974). Crustal thicknesses ~40 km have been measured using seismic refraction in and around the SRP (e.g. Iyer 1984), and are approximately the same as, or slightly less than, those in many other parts of North America, including the adjacent regions that are not actively extending. Although there may be some residual thermal anomaly in the NE Basin and Range Province due to earlier crustal thinning, making this region weaker than its surroundings and hence easier to deform, as suggested by Sonder et al. (1987), the absence of any excess crustal thickness suggests that present-day extension of the NE Basin and Range Province is not a consequence of present-day excess crustal thickness. Although crustal thinning accompanies this extension, it may be as an effect rather than a cause. Furthermore, the age-relations of magnetism along the SRP and its relationship to Yellowstone are not explained by the Sonder et al. (1987) model, nor are the positions of the zones of present day extension within this region, nor are the senses of oblique slip on major normal faults.

In contrast, the relationship between the positions of Yellowstone, the SRP, and the zones of active normal faulting in the NE Basin and Range Province suggests that, if the SRP is caused by the upwelling Yellowstone plume, and if deformation in the two active zones beside it is a consequence of the pattern of flow in the deforming fluid that underlies these regions, then this pattern of flow is also related to the presence of the Yellowstone plume. There is no direct a priori evidence that the dynamics of the NE Basin and Range Province are related to the dynamics of Yellowstone and, as already mentioned, models have been proposed in which the two are unrelated. If the two features are considered dynamically unrelated, then the relationship between the positions of Yellowstone, the SRP and the active normal faults in the NE Basin and Range Province is coincidental, and separate explanations need to be found for all these features. It appears more satisfactory if a single explanation can be suggested that can account for these features and the evolution of this region on a restricted timescale of ~10 Myr. Most extension outside the NE part of this province occurred long before the Yellowstone plume was in its present position and the SRP was formed, and must have been caused by processes not directly related to these features; thinning of previously overthickened lithosphere may well be important in this context.

It could be argued that deformation in the NE Basin and Range Province may indeed be related to Yellowstone, but is caused simply by thermal contraction and subsidence while the plate cools as it moves off the plume. However, if this were the only physical process operating, deformation would be concentrated SW of Yellowstone along the SRP, where heating by the plume will have been most intense, rather than in two zones on either side of the SRP. Furthermore, it is difficult to see how thermal contraction could drive rotation around vertical axes. It appears more reasonable instead to investigate whether regional tectonic deformation is driven by circulation in the plume.

Finally, it could also be argued that the Yellowstone plume is sheared SW at present because of lateral rheology contrast between relatively cold lithosphere of usual thickness to the NE, which has remained largely undeformed for hundreds of Myr, and relatively hot, thin, weaker lithosphere in the Basin and Range Province to the SW that was affected by late Cretaceous and early Tertiary shortening and subsequent extension. Under this interpretation the strong lithosphere to the NE acts as a barrier that return flow from the plume cannot penetrate. In contrast, under the assumption that southward plate motion is shearing the plume, the region NE of the position of the upwelling at any time is undeformed by the plume because the plume has not yet reached it. Given the present-day position of the plume near the edge of relatively strong lithosphere, it is not possible to distinguish between these two hypotheses on the basis of intuitive reasoning.

5 MODELLING NE BASIN AND RANGE PROVINCE DEFORMATION

In this section I investigate possible models for the pattern of flow in the deforming fluid beneath the NE Basin and Range Province. I begin in subsection 5.1 with a radially-symmetric model, following the suggestion of Smith & Sbar (1974). Such a model cannot account for rotation around vertical axes because circulation in it is irrotational. In Subsection 5.2, the main part of this section, I investigate the relative sizes of velocity components and velocity gradient tensor elements for a plume aligned along the x-axis, the direction of motion of the North American plate. The smallest velocities and velocity gradients in this model are likely to be \( v_y \) and its derivatives. The simplest realistic two-dimensional \((x-y)\) model thus has \( v_y \) constrained to be zero. In the same subsection I explore this model quantitatively, and show that it can account for the main deformational features in the region. In Subsection 5.3 I refine the model slightly by attempting to resolve the direction of \( v_y \). Finally in Subsection 5.4 I discuss qualitatively the necessary features of three-dimensional models consistent with the surface observations.

5.1 Radially symmetric two-dimensional model

The simplest possible pattern of flow around the Yellowstone plume is divergent and radially symmetric, similar to the pattern suggested by Smith & Sbar (1974). This can readily account for crustal extension in the surrounding areas provided this is in directions that are radial or tangential to the plume. However, the observations in Section 3 indicate, first, that slip vectors in the actively deforming zones of the NE Basin and Range Province well away from Yellowstone are neither radial nor tangential to the plume. Second, flow with radial horizontal velocity is irrotational, with no vertical vorticity, and hence cannot drive rotation around vertical axes. Third, observed deformation is strongly concentrated S and W of Yellowstone, suggesting that if this deformation is an effect of Yellowstone then it is not a radially symmetric effect. Whilst this may be a consequence of the differences in lithosphere rheology S and W of Yellowstone compared with N and E of it, as discussed in Section 4, it is also consistent with the return flow of Yellowstone plume being...
sheared SW in the direction towards which the North American plate is moving.

Linear combinations of radially-symmetric plume velocity and uniform plate velocity can account for asymmetry of overall deformation between the NE and SW directions. However, any such linear combination will also be irrotational and hence cannot drive rotation around vertical axes.

5.2 Non-radially-symmetric two-dimensional model: \( v_y \) zero

To account for the suggested vorticity in the deformation zones, the Yellowstone plume cannot be radially symmetric. A Cartesian coordinate system to represent this plume can be defined with \( x \)-axis oriented roughly SW along the SRP, and \( y \)-axis oriented roughly SE towards the Intermountain seismic belt (Fig. 8). The \( z \)-axis is the downward vertical. Within the plume, \( v_y \) can be assumed to increase from zero near Yellowstone where the plume is upwelling, to a value close to the plate velocity \( A \) at great distance from the plume in the +\( x \) direction. It can either increase monotonically (profile 1 in Fig. 9) or via a maximum where it is locally greater than \( A \) (profile 2 in Fig. 9). For a plume that is not sheared sideways by an overlying plate, initial upward velocity will be deflected radially outward, making the plume resemble a mushroom. Its divergence or geometrical spreading will lead to radial horizontal velocities much smaller than the initial upwelling velocity when at distances more than a few times the radius of the upwelling 'stem' away from its axis. However, for a plume that is strongly sheared horizontally into a zone that is not many times wider than its vertical 'stem', geometrical spreading will be much less important. In these circumstances, profile 1 is likely if the upwelling velocity is much less than \( A \), and profile 2 is likely if the velocity of upwelling is comparable to or greater than \( A \).

Upwelling velocity in the Yellowstone plume can be approximately estimated simplistically using Stokes' law. A spherical blob of material with density \( \rho_p \) and radius \( a \) rising at velocity \( v_r \) through a surrounding fluid with density \( \rho_m \) experiences viscous drag force \( F_v \), where:

\[
F_v = 6\pi \eta_m a v_r. \tag{5.1}
\]

**Figure 8.** Schematic map of the proposed initial two-dimensional model for the sheared plume beneath the NE Basin and Range Province, in a co-ordinate system or reference frame fixed relative to the upwelling part of the plume. Well away from the plume, the North American plate is moving in the +\( x \) direction at velocity \( A \). Yellowstone is assumed to be at the point of maximum radial horizontal velocity and maximum upwelling velocity, at point \( Y \) with coordinates \((x_m, 0)\). The plume is arranged symmetrically about the +\( x \) axis, with outer limits defined by \( y = \pm P(x) \). On both sides of the interface at \( y = \pm P(x) \), \( v_y = A \).
If \( v_u \) is its terminal velocity, then \( F_u \) balances the buoyancy force \( 4\pi a^2 g(\rho_m - \rho_p)/3 \) due to the difference in density between the blob and its surroundings, where \( g \) is acceleration of gravity. Thus:

\[
v_u = \frac{2a^2 g(\rho_m - \rho_p)}{9\eta_m}.
\]  

Using \( g = 10 \text{ m s}^{-2} \), \( \eta_m = 10^{21} \text{ Pa s} \), both standard values, and \( (\rho_m - \rho_p) = 0.05\rho_m = 150 \text{ kg m}^{-3} \), \( v_u \) is \( \sim 25 \text{ mm yr}^{-1} \) for \( a = 50 \text{ km} \) and \( \sim 100 \text{ mm yr}^{-1} \) for \( a = 100 \text{ km} \). This realistic range of plume radius gives velocities of upwelling that bracket the value of \( A \), so neither profile 1 nor 2 in Fig. 9 can be eliminated on the basis of this test. The 5 per cent density contrast assumed is conservative, given the apparent \( \sim 10 \text{ per cent} \) P-wave velocity anomaly beneath Yellowstone relative to its surroundings (Iyer 1984).

From equation (2.5), in order for \( \chi_x \) to be negative,

\[
\chi_x = \frac{\partial v_x/\partial y - \partial v_y/\partial x}{\partial x} < 0.
\]  

Close to Yellowstone, \( v_x \) will be zero at \( (x = 0, y = 0) \) above the upwelling. It will equal \( A \) at positive values of \( x \) beyond the effect of the plume. Also, close to Yellowstone \( \partial v_x/\partial y \) will be large and positive as \( v_x \) varies between a near-zero value close to the \( y \)-axis and the plate velocity \( A \) at the edge of the plume. Thus strong variations in \( v_x \) of about tens of \( \text{mm yr}^{-1} \) will exist around and SW of Yellowstone. In contrast, \( v_y = 0 \) both outside the sheared plume and, from symmetry, along the +\( x \)-axis. It may be non-zero off-axis, but, unlike \( v_x \), there is no requirement in the model for lateral variations in \( v_y \) of tens of \( \text{mm yr}^{-1} \). The sign of \( \partial v_x/\partial x \) is thus not clear, but given that variations in \( v_x \) appear to be much greater than variations in \( v_y \), it is likely that \( \partial v_x/\partial y \) will make the dominant contribution to \( \chi_x \) and control its sign. This will make \( \chi_x \) positive, or clockwise, in the +\( y \) quadrant where \( x \) is small, consistent with the sense inferred from left-lateral slip on the Teton fault. Thus, this slip sense may indeed be due to response of the blocks around the Teton fault to the underlying vertical vorticity, and not simply caused by fault strike being oblique to the local principal extensional strain rate azimuth. Slip vectors on the Teton and Hebgen faults suggest that principal extensional strain rate is locally oriented sub-parallel to the \( x \)-axis, implying that \( \partial v_x/\partial x \) is locally much greater than any of the three other horizontal gradients of horizontal components.

Under profile 1, \( \chi_x \) will be clockwise at all points along the +\( y \) branch of the plume-lithosphere interface, and anticlockwise along the −\( y \) branch. Vertical vorticity would thus be in the opposite sense to that reported in Section 3 at large \( x \) along both seismic zones beyond −150 km from Yellowstone. However, profile 2 predicts the observed sense of vertical vorticity at large \( x \), and suggests it is in the

\[\text{Turning Point}\]

**Figure 9.** Possible profiles of \( v_u(x, y = 0) \) discussed in the text. Profile (2) was calculated using equation (5.6) with \( A = 40 \text{ mm yr}^{-1} \), \( B = 60 \text{ mm yr}^{-1} \), \( x_2 = 150 \text{ km} \) and \( x_3 = 450 \text{ km} \). Profile (1) was calculated with \( B = 0 \).
opposite sense for small $x$. Such a pattern of flow can thus explain the apparent reversal in vertical vorticity between the Teton fault and more distant faults south of Yellowstone, and between the Hebgen Lake area and more distant parts of the central Idaho seismic zone.

Slip vector azimuth also constrains the velocity field between $-200$ and $400$ km from Yellowstone. Its inward orientation towards the $x$-axis at $\psi \sim 45^\circ$ implies that strain rate tensor element $E_{xy}$ is negative in the $(+x, +y)$ quadrant. From the form of this element in equation (2.2), this means that:

$$2E_{xy} = \partial v_x / \partial y + \partial v_y / \partial x < 0.$$  \hfill (5.4)

Equations (5.3) and (5.4) can only be satisfied simultaneously if $\partial v_y / \partial y < 0$ and $|\partial v_x / \partial y| > |\partial v_y / \partial x|$. This latter condition requires that $v_x$ varies with $x$ along $y = 0$ as in profile 2 in Fig. 9.

A further constraint comes from the observations in Section 3 that suggest rotation rate ($-0.5 \chi_x$) and principal extensional strain rate in the two seismically active zones beyond $-200$ km from Yellowstone are numerically roughly equal, with values $\sim 10^{-15}$ s$^{-1}$ (or $\sim 2''$ Myr$^{-1}$), equivalent in terms of equation (2.23) to $H_2 \sim 0.6$–0.8. This implies from equations (5.3) and (5.4) that $|\partial v_x / \partial y| \gg |\partial v_y / \partial x|$. The simplest model satisfying this condition has $v_x = 0$ everywhere. Making this assumption avoids the need to tackle for the time being the sign of $v_x$ in each quadrant, which has not yet been determined. Regardless of its sign, the similar numerical values of $\chi_x$ and $E_1$ suggest that $v_x$ is likely to be small compared with $v_y$, and derivatives of $v_x$ are also likely to be small compared with the derivatives of $v_y$. For the time being I investigate in detail the consequences of constraining $v_x$ to be zero, before returning in Subsection 5.3 to investigate the magnitude and sign of $v_x$.

Consider the pattern in Fig. 8 of the upper boundary of a plume sheared southwestward by the North American plate moving across it. The upwelling is assumed to begin to affect the brittle layer at point $(x = 0, y = 0)$, and the plume is sheared in the $+x$ direction due to the velocity $v_x = A$ of the lithosphere moving across it. The plume is assumed to penetrate the lithosphere such that circulation near its upper boundary influences deformation in the brittle layer across a profile defined by:

$$y = P(x) = y_0 [1 - \exp(-x/x_0)],$$  \hfill (5.5)

where $y_0$ and $x_0$ are constants. The principal velocity gradients are likely to be near the interface between the plume and the surrounding lithosphere. Since the plume is likely to have lower viscosity than the surrounding lithosphere, velocity gradients are likely to be predominantly on the plume side of this interface. For a further approximation I assume the material on the lithosphere side of the interface is moving uniformly at velocity $A$ in the $+x$ direction, thus neglecting all velocity gradients outside the plume. A profile of $v_x$ along $y = 0$ like profile 2 in Fig. 9 can be generated using:

$$v_x(x, y = 0) = (A + B)[1 - \exp(-x/x_2)] - B[1 - \exp(-x/x_3)].$$  \hfill (5.6)

where $B > 0$ and $x_3 > x_2$. Profile 1 can be generated if required by setting $B = 0$.

It is important to stress that equations (5.5) and (5.6) are not based on fluid mechanics. They are simply convenient algebraic forms that give profiles like those in Fig. 9 using the minimum number of free parameters. Equation (5.6) has been invented to suggest a form for $v_x(x, y = 0)$ that can vary smoothly as either profile 1 or profile 2 of Fig. 9. Similarly, equation (5.5) has been invented to suggest a form of the profile of the outer margin of the sheared plume that varies smoothly so as to form an envelope that bounds the outer limits of deformation of the brittle layer (Fig. 10). These two equations thus constitute a simple empirical kinematic description of a sheared plume involving only five free parameters: $B$, $x_0$, $x_2$, $x_3$ and $y_0$, which can be constrained from observations of the overall shapes and deformation rate of the two seismically-active zones. It would of course be preferable to use velocity fields derived from proper numerical simulations of plume dynamics, or from analytic solutions for such plumes, if these existed. In the absence of such solutions, equations (5.5) and (5.6) provide a reasonable guess as to the form of the horizontal velocity field within a sheared plume.

Symmetry considerations indicate that along $y = 0$,

$$v_x(x, y = 0) = 0.$$  \hfill (5.7)

For the time being I continue to assume that $v_x = 0$ elsewhere. Along the interface at $y = P(x)$, I assume a boundary condition:

$$v_x(x, y = P) = A.$$  \hfill (5.8)

The plume–lithosphere interface and the $x$-axis are not parallel; they diverge at angle $\zeta$, where:

$$\tan (\zeta) = \frac{\partial y}{\partial x} = \frac{y_0}{x_0} \exp(-x/x_0).$$  \hfill (5.9)

Thus, $v_x$ within the plume is not parallel to its boundaries. I assume that $v_x$ at all points is such that the incompressibility condition (equation 2.1) is satisfied given $v_x$. Between its limiting values, I assume that $v_x$ within the plume can be interpolated as:

$$v_x(x, y) = A + [v_x(x, y = 0) - A] \cos^2 \left(\frac{\pi y}{2P}\right).$$  \hfill (5.10)

Figure 10 shows observed values of $y$ against $x - x_m$ for some of the principal active normal faults in the NE Basin and Range province, using data in Table 2. The free parameter $x_m$ positions the sheared plume velocity field, determined by parameters $x_0$, $y_0$, $x_2$, $x_3$ and $B$, relative to the geographical coordinate system. The point $(x = x_m, y = 0)$ denotes the centre of Yellowstone. The overall width across both deforming zones at their ends in the $+x$ direction is $\sim 500$ km, constraining $y_0$ to be at least $250$ km, but no greater than $350$ km. A value $300$ km is preferred. The gradient of the envelope of the deforming zone, $\partial P/\partial x$ or $\tan (\zeta)$, equals $y_0/x_0$ close to $x = 0$ (equation 5.9). The two zones converge at Yellowstone from directions roughly $90^\circ$ apart, suggesting that near the origin $\partial P/\partial x \approx 1$ and hence $y_0 \approx x_0$. The Yellowstone caldera and the regions of very high surface heat flow, anomalous density, and P- and S-wave velocity, within it (e.g. Iyer 1984), have radius $\sim 50$ km, providing an independent estimate for the size of the initial upwelling. Given this radius, and given that the
width of the model plume velocity field decreases to zero at 
\((x = 0, y = 0)\), \(x_m \sim 50 \text{ km} \) suitably positions the centre 
of Yellowstone relative to the origin. Fig. 10 suggests also that 
observed positions of major active faults relative to 
Yellowstone are consistent with \(x_m \sim 50 \text{ km} \). Parameter \(x_2 \)
controls the scale over which \(v_x \) builds up to its maximum 
value in the model, increasing from zero to values \(\sim A \) 
and greater. The suggested reversal in sense of vertical vorticity 
\(\sim 150 \text{ km} \) from Yellowstone suggests \(x_2 \sim 150 \text{ km} \). The only 
constraint on \(x_2 \) is that it must be sufficiently large, given \(B \),
to give the turning point of \(v_x(x, y = 0) \) sufficiently far along 
the \(+x\) axis to reproduce the observed extent of extension in 
the \(+x\) direction; \(x_2 = 450 \text{ km} \) is reasonable. Observed 
vertical vorticity is up to \(4^\circ \text{Myr}^{-1} \), providing an 
order-of-magnitude estimate for the velocity gradient 
\(\partial v_x / \partial y \). Across a 250 \text{ km} \) wide zone this gradient leads to a 
change in \(v_z \) of \(-15 \text{ mm yr}^{-1} \), suggesting that \(v_x \) is up to 
\(\sim 55 \text{ mm yr}^{-1} \) on the \(+x\) axis.

Figure 11 indicates the velocity field, vorticity, and strain 
rate eigenvalue \(E_1 \) and eigenvector orientation \(\psi \) determined 
using this model for \(x_0 \sim 300 \text{ km}, y_0 \sim 300 \text{ km}, x_2 \sim 
150 \text{ km}, x_3 \sim 450 \text{ km}, A \sim 40 \text{ mm yr}^{-1} \) and \(B \sim 60 \text{ mm yr}^{-1} \).
Because \(v_x \) is zero everywhere, \(E_1 \) is the only non-zero 
horizontal strain rate tensor eigenvalue. To generate this 
figure, a grid with \(\Delta x = 5 \text{ km} \) and \(\Delta y = 5 \text{ km} \) was first 
generated covering the area shown. \(v_x \) was calculated as 
a function of \(x \) along \(y = 0 \) using equation (5.6). \(v_x \) was also 
constrained equal to \(A \) everywhere along and outside 
\(y = \pm P(x) \) using equation (5.5). \(v_x(x, y) \) was then 
interpolated between its \(y \) limits using equation (5.10).

Because \(v_x \) is constrained zero everywhere, observable 
quantities \(x_1, E_1 \) and \(\psi \) depend on the derivatives \(\partial v_x / \partial x \) 
and \(\partial v_x / \partial y \). These were estimated by taking differences in 
values of \(v_x \) between adjacent grid elements in a symmetric 
manner:

\[
\frac{\partial v_x}{\partial x}(x, y) = \frac{v_x(X + \delta x, Y) - v_x(X - \delta x, Y)}{2\delta x} \quad (5.11)
\]

\[
\frac{\partial v_x}{\partial y}(x, y) = \frac{v_x(X, Y + \delta y) - v_x(X, Y - \delta y)}{2\delta y} \quad (5.12)
\]

Given these derivatives, the horizontal strain rate tensor is 
defined. Its eigenvalue \(E_1 \) and eigenvector orientation \(\psi \) can 
be found using standard methods. Given equation (2.4), \(\chi_x \)
can also be determined. Finally, derived parameters \(H_1 \) and 
\(C \) can be calculated using equations (2.23) and (2.17). These 
quantities are contoured as functions of \(x \) and \(y \) in Fig. 11.

Given that values of the free parameters in this model 
have been constrained to within \(-30 \% \) per cent by features 
observed directly in the deforming zones in the NE Basin 
and Range Province, the good agreement between this 
model and the observed features is perhaps not surprising 
(Fig. 11). The model predicts two zones of distributed 
extension in the expected positions, with principal strain 
rate up to \(~2 \times 10^{-15} \text{ s}^{-1} \) in direction oriented at \(-45^\circ \) 
to the \(x\)-axis. Vertical vorticity in the model is anticyclonically 
in the \((+x, +y) \) quadrant, the Intermountain seismic belt, and 
clockwise in the \((+x, -y) \) quadrant, the central Idaho.
Figure 11. Contour plots of the horizontal velocity field and some of its derivatives within the sheared plume as functions of $x$ and $y$, given the $v_x$ distribution in equation (5.6) with $A = 40$ mm yr$^{-1}$, $B = 60$ mm yr$^{-1}$, $x_2 = 150$ km and $x_3 = 450$ km, with a plume with profile described using equation (5.5) with $x_0 = y_0 = 300$ km. (a) $x$-component of velocity, $v_x$ in mm yr$^{-1}$. (b) Vertical vorticity, $\omega_y$ in $^\circ$Myr$^{-1}$. (c) Eigenvalue $E_x$ of the horizontal strain rate tensor corresponding to observed principal horizontal strain rate in s$^{-1}$. (d) Angle $\psi$ between the $x$-axis and the principal horizontal strain rate azimuth, in $^\circ$. (e) Vertical Holmes number $H_x = (2E_x)$. (f) Horizontal length-scale parameter $C$ (equation 2.17) in km.

seismic zone. Near the origin, the model predicts the opposite sense of vertical vorticity in each quadrant compared with further away, and near the transition between senses of vorticity the model predicts extensional strain rate sub-parallel to the $x$-axis; both in agreement with observations from the Teton and Hebgen Lake areas. Extensional strain rate and vertical vorticity predicted by the model are up double observed values. This may partly be because true observed fault slip rates are underestimated because Holocene slip rate determinations do not take account of time since the last earthquake (equations (3.5) and (3.6)). Because the model overestimates both $\omega_y$ and $E_x$, by approximately the same proportion, its estimate of $H_x$, their ratio, agrees with observation. The length scale parameter $C$ suggests that $(\nabla \times v)_x$ varies by only a small proportion of its typical value on $\sim 100$ km scale along $\sim 200$ km long zones beneath the Intermountain seismic belt and central Idaho seismic zone. Given the reasoning embodied in equations (2.14)–(2.23), this implies that the model predicts these are the only parts of the NE Basin and Range province where oblique faults bounding large $\sim 100$ km size blocks can take up extensional strain and vertical vorticity in the underlying ‘fluid’. This agrees with the observation that fault-bounded blocks $\sim 100$ km long exist in both zones (Fig. 2). Elsewhere in the NE Basin and Range province, $(\nabla \times v)_x$ is smaller and varies as a proportion of its typical value on smaller spatial scale. Thus, elsewhere only relatively small fault-bounded blocks, with dimensions $<\sim 10$ km, can take up local strain rate and vertical vorticity in the underlying ‘fluid’. This length scale parameter $C$ may thus explain why large ($M_t \sim 7$) earthquakes that require relatively large (length along strike $\sim 30$ km) faults only occur in the two principal seismically active zones. Elsewhere, the region may be broken up with smaller faults and blocks to take up local strain rate and vertical vorticity, and only smaller earthquakes can occur.
This model can thus account for the observed concentration of extension in the Intermountain Seismic Belt south of Yellowstone and in the central Idaho seismic zone west of Yellowstone, and explains the observed present-day sense and rate of extension and rotation around vertical axes in these zones. However, the limited observational data have all been used to derive this model, and thus no data exist that can test it. Objective testing can only be done when additional independent data have been gathered. This may include paleomagnetic rotation observations that give independent estimates of $x$ or field investigations that give senses and rates of slip on faults. It is important to stress the need for fieldwork to investigate sense and rate of strike-slip on major predominantly normal faults, as well as the normal slip rate that is larger and easier to measure.

This same model can test the alternative suggestion that the upwelling part of the plume is trapped and sheared SW by lateral rheology contrast at the NE edge of the Basin and Range province between weaker lithosphere to the SW and stronger lithosphere to the NE. This is equivalent to setting $A = 0$, and looking for solutions of the form of profile 2 in Fig. 9. Such solutions would have vertical vorticity clockwise throughout the central Idaho seismic zone and anticlockwise throughout the Intermountain seismic belt. The observed reversal in vertical vorticity in both zones ~100 km from Yellowstone thus cannot be reconciled with this suggestion. This implies that plate motion is the main cause of SW shearing of the plume. The upwelling plume may well be able to melt and burn its way in future through the relatively strong lithosphere of Montana. In 10 Myr, when its projected course takes it to NE Montana close to the Canadian border, circulation associated with it may deform flat land in southern Montana such that it resembles central Idaho today.

5.3 Non-radially-symmetric two-dimensional model; $v_y$ non-zero

There is no requirement for $v_y$ to be zero other than along the $y$-axis; $v_y$ was set to zero to simplify the analysis in Subsection 5.2 given evidence that it is small. The only available evidence to constrain $v_y$ is the observation that in the parts of both deforming zones where sets of parallel en echelon faults exist, $H_z \approx 0.6$ on average, implying from equation (2.23) that $|2E_i| \approx |x|$ for these localities, Fig.
9 shows that in the (+x, +y) quadrant \( \psi = -45^\circ \) and in the (+x, -y) quadrant, \( \psi = +45^\circ \). With \( v_y \) and its derivatives zero to a first approximation, equation (2.10) simplifies to:

\[
E_{xx} \cos(\psi) + E_{xy} \sin(\psi) = E_1 \cos(\psi) \\
E_{xy} \cos(\psi) = E_1 \sin(\psi).
\]

With \( \psi = \pm 45^\circ \), equation (5.14) indicates that \( |E_{xy}| \approx |E_1| \), and hence \( H_1 = |x_i|/|2E_{xy}| \). Eigenvalues and eigenvector orientations of \( \mathbf{E} \) are stationary with respect to small changes in \( \mathbf{E} \), and hence \( H_1 \) and \( |x_i|/|2E_{xy}| \) will remain roughly constant even if \( v_y \) is no longer constrained zero. \( |H_1| < 0.6 \) both observed in patterns of faulting and deduced in Fig. 11(e) from the numerical model, beneath both the Intermountain seismic belt and the Central Idaho seismic zone, implies \( 2E_{xy} \approx |x_i|/H \approx 1.7 |x_i| \). In terms of equations (5.3) and (5.4), if \( 2E_{xy} > |x_i| \), then \( \partial v_x/\partial x \) and \( \partial v_y/\partial y \) have the same sign in both quadrants. This means that \( \partial v_x/\partial x < 0 \) in the (+x, +y) quadrant and >0 in the (+x, -y) quadrant. At a given positive value of \( y \) close to the leading edge of the plume, \( v_y \) will be zero. With \( \partial v_y/\partial x < 0 \) in the plume, at greater \( x \) \( v_y \) will build up in the \(-y\) direction. The same reasoning suggests that \( v_y > 0 \) inside the plume in the (+x, -y) quadrant. Thus, beneath both deforming zones \( v_y \) will be inwards towards the x-axis. The same reasoning suggests, given that \( v_y = 0 \) along \( y = 0 \), that \( \partial v_y/\partial y \) is negative in the (+x, +y) quadrant also.

This result appears counterintuitive; it seems plausible a priori that any upwelling at Yellowstone will persist along the x-axis, spilling out in the +y and -y directions, to make \( v_y \) directed outwards. However, the deduction that beneath the deforming zones at distances between 100 and 300 km from Yellowstone \( v_y \) is inwards follows directly from the observation that the normal component of slip rate on faults in these regions is larger than the strike-slip component, consistent with the evidence in Section 3.

More generally, with \( E_1 = |E_{xy}| \tan(\psi) \) from equation (5.14), still assuming gradients of \( v_y \) are small compared with those of \( v_x \), equations (5.3) and (5.4) can be solved simultaneously to give:

\[
\frac{\partial v_x}{\partial x} = H_1 \cot(\psi) - 1 \frac{\partial v_y}{\partial y} \\
\frac{\partial v_y}{\partial x} = H_1 \cot(\psi) + 1 \frac{\partial v_y}{\partial y}.
\]

Solutions to this equation are listed in Table 3. With \( H 0.6 \) and \( \psi 45^\circ \), \( \partial v_y/\partial x \sim 0.25 \partial v_y/\partial y \). Thus the observations that both \( H \sim 0.6 \) and \( \psi \sim 45^\circ \) indicate this is a configuration...
where $\frac{\partial v_y}{\partial x}$ is indeed small compared with $\frac{\partial v_x}{\partial y}$. It is likely that $\frac{\partial v_y}{\partial y}$ will be of similar size to $\frac{\partial v_x}{\partial x}$ beneath many localities. Thus, for the NE Basin and Range province horizontal gradients of $v_y$ are likely to be small compared with horizontal gradients of $v_x$.

With $v_y$ non-zero, the strain rate tensor may have two non-zero eigenvalues, associated with orthogonal eigenvectors. Given that for $H_y \sim 0.6$ and $\psi \sim 45^\circ$ horizontal gradients of $v_y$ are small compared with horizontal gradients of $v_x$, the first eigenvector will have a similar azimuth to that in Fig. 11, and the associated eigenvalue will also be similar to that in Fig. 11. The second eigenvector will be oriented at azimuth $90^\circ$ different, and its eigenvalue $E_2$ will be small compared with $E_1$. Likewise, $\frac{\partial v_y}{\partial x}$ contributes to the vertical vorticity, but provided it remains small the distribution of vertical vorticity will change little from Fig. 11(a).

5.4 Three-dimensional model

The flow in the deforming 'fluid' close to its upper boundary is likely to be parallel to this boundary, and hence horizontal, and the assumption made that the pattern of flow that influences the behaviour of blocks in the brittle layer is horizontal is reasonable. However, vertical velocity will become non-zero at greater depth, and vertical derivatives of all three velocity components are likely to be non-zero below this upper boundary also. The two horizontal vorticity components, $\chi_x$ and $\chi_y$, depend on these vertical derivatives (equations 2.6 and 2.7), and hence are also likely to be non-zero near this upper boundary. $\chi_y$ will cause rotation in the brittle layer around the $y$-axis. Most faults in both seismic zones dip W or SW, particularly at distances greater than $\sim 150$ km from Yellowstone, and under the 'domino' model (equations 3.2-3.4) will rotate during extension with axis of rotation along the $-y$-axis, such that $\chi_y < 0$. The decrease in topographic elevation from both seismic zones towards the SRP has already been highlighted in Fig. 7. If this is a feature related to the circulation in the underlying fluid, and in particular to the $x$-component of vorticity twisting the Earth's surface around the $x$-axis, it suggests that this component of vorticity is along the $+x$-axis for $y > 0$, and along the $-x$-axis for $y < 0$.

Patterns of observables dependent on $\chi_x$ and $\chi_y$ are antisymmetric with respect to inversion across the $y$-axis, whereas those dependent on $\chi_y$ are symmetric; in agreement
with the symmetry properties predicted given the definitions of these quantities in terms of Cartesian coordinates and derivatives of velocity (equations 2.5-2.7), given that symmetry considerations require both \( u_x \) and \( u_y \) to have the same sign on both sides of the \( y \)-axis, whereas \( u_z \) changes sign. Senses of the three components of vorticity on both sides of the \( y \)-axis, whereas \( v_y \) changes sign. Senses of the three components of vorticity on both sides of the SRP suggested by observation are shown schematically in Fig. 12.

The sense of \( u_x \) in Fig. 12 is consistent with the inward \( v_y \) near the upper boundary of the deforming 'fluid' deduced in Subsection 5.3. Part of the three-dimensional circulation beneath the SRP beyond \(-150 \text{ km}\) from Yellowstone appears to comprise inward movement from both sides towards the axis of the SRP, revealed by this vorticity component and the sense of \( v_y \). Given that negligible deformation, neither extension nor shortening, appears to be occurring along the axis of the SRP, incompressibility (equation 2.1) requires that inward movement towards the SRP is accompanied by downwelling beneath it (Fig. 13a). A proportion of the magma that has erupted to form the widespread Neogene and Quaternary SRP basalts may have been carried inward by this pattern of flow being erupted, and may have come from a relatively shallow mantle source elsewhere and not from the Yellowstone plume. Thus this pattern of flow may account for the lack of deep mantle characteristics (e.g. Thompson 1977) in SRP basalts.

The contributions to \( \chi_x \) and \( \chi_y \) from both sides of the SRP are equal in magnitude but opposite in sign, and hence zero total angular momentum is associated with both \( \chi_x \) and \( \chi_y \) when summed overall. In contrast, \( \chi_y \) has the same sign on both sides of the SRP, and the angular momentum associated with it beneath the region more than \(-150 \text{ km}\) from Yellowstone does not sum to zero. The motion of the plume, initially upwelling in the \(-z\) direction, then sheared in the \(+x\) direction by the plate motion, can impart no angular momentum around the \( x \)- or \( z \)-axis, but appears to impart a component of angular momentum, \( P_{x,y} \), along the \(-y\) direction into the deforming 'fluid' beneath the region between \(-150\) and \(-350 \text{ km}\) from Yellowstone, resulting in this overall non-zero summation of the \( y \) component of angular momentum there.

Neither the initial upwelling plume nor the North American plate in the absence of the plume possesses any angular momentum around the \( y \)-axis. Conservation of angular momentum requires that the combined system of
the plume and the plate interacting and deforming each other must also have no overall angular momentum around this axis. The component of angular momentum described above must be balanced by an equal and opposite component $P_{y,c}$ within the part of the plume that is beneath and close to Yellowstone, which is being deflected from vertical to predominantly horizontal motion by a rotation around the $+y$-axis (Fig. 13b).

Conservation of angular momentum around the $y$-axis provides an independent estimate of the upwelling velocity in the plume. The zone with the $-y$ component of angular momentum extends between about $y = +250$ to $-250$ km and $x = 150$ to $350$ km. Section 3 suggested that rotation rates of fault-bounded blocks around the $-y$-axis imply typical $y$-component of vorticity $(x_y) \sim 4 \, \text{Myr}^{-1}$. If this zone rotating around the $-y$-axis is approximated as a cylinder with density $\rho_c, 3000 \, \text{kg m}^{-3}$, radius $r_c, 100$ km and length along axis $L_c, 500$ km, then its moment of inertia is $\pi r_c^4 L_c/2$ and its angular momentum is:

$$P_{y,c} = \frac{\pi \rho_c r_c^4 (x_y)}{2}. \quad (5.16)$$

If the part of the plume being deflected horizontally is assumed relatively narrow, with radius $a$ and density $\rho_p$, moving with velocity $v_z$ in a quarter of a circular arc with radius $R_p$ between the upward direction and the positive $x$-direction, then its angular velocity is $v_z/R_p$ and its approximate moment of inertia, assuming $r_p \ll R_p$, is $(\pi a^2)(\pi R_p/2)(\rho R_p^2)$. Given this moment of inertia, its
Figure 13. (a) Schematic cross-section, not to vertical scale, in the y-z plane across the SRP Plain close to the line of section E–F in Fig. 2. Large arrows indicate the direction of velocity at depth given the signs of $\partial u_y/\partial z$ and $\partial u_z/\partial y$ deduced in the text and the sense of vorticity component $\chi_x$ shown in Fig. 11. Small arrows show possible sense of movement of magma before it erupts into the SRP. (b) Schematic cross-section, not to scale, indicating the suggested circulation in the x-z plane beneath and close to Yellowstone. Arrows indicate local direction of v.

Angular momentum is:

$$P_{x.p} = \frac{\pi^2 \rho_p a^2 R_p^2 v_z}{2}. \quad (5.17)$$

Equating $P_{x,c}$ and $P_{x.p}$ gives:

$$v_z = \frac{\rho_c \pi^4 L_c \chi_x}{\pi \rho_p a^2 R_p^2}. \quad (5.18)$$

Approximating $\rho_c = \rho_p$, for $a = 35-50$ km and $R_p = 100$ km, equation (5.18) gives $v_z \sim 60-130$ mm yr$^{-1}$, similar to the range of values calculated using the Stokes' law method (equation 5.3).

Upwelling velocity scales differently with plume radius under the two methods. For given angular momentum necessary to deform the NE Basin and Range Province with the observed $\chi_x$, under equation (5.18), $v_z \propto a^{-2}$, whereas under equation (5.3) for given density contrast between the plume and its surroundings and mantle viscosity, $v_z \propto a^2$. With the values assumed for the other quantities, $a \sim 50$ to $\sim 100$ km is consistent with both methods, suggesting this is a reasonable value for upwelling plume radius. This is similar to radii of the heat flow and seismic velocity anomalies around Yellowstone, and consistent with numerical plume simulations (e.g. Craig & McKenzie 1987).

For $\chi_x$ to be positive in the +y quadrant when greater than $\sim 150$ km from Yellowstone where $\partial u_y/\partial y$ is likely to be positive ($v_z$ is upward and hence negative beneath Yellowstone, and its absolute value will be likely to decrease.
away from the centre of the upwelling; hence both \( \partial u_x / \partial y \) and \( \partial v_x / \partial x \) are likely to be positive moving in the +x and +y directions away from Yellowstone (equation 2.6), \( \partial u_x / \partial y \) must be sufficiently large and positive to overwhelm \( \partial v_x / \partial x \), suggesting that \( v_x \) increases strongly with depth. The same reasoning suggests that \( v_x \) increases strongly with depth in the -y direction in the (+x, -y) quadrant. Similarly, to make \( \chi_x \) negative in both the (+x, +y) and (+x, -y) quadrants requires \( \partial u_x / \partial x \) is sufficiently large and positive to overwhelm \( \partial v_x / \partial x \), indicating that \( v_x \) also increases strongly with depth. This discussion thus suggests that vertical derivatives of horizontal velocity components are likely to be more important than horizontal derivatives of \( u_x \) in determining the sense of horizontal vorticity components beneath the NE Basin and Range Province.

Topographic rotation of \(-1^\circ\) around the x-axis at the 5 Myr SRP age point in Fig. 7 suggests that local rotation rate of the brittle layer around the x-axis is \(-0.2^\circ\)Myr\(^{-1}\), implying \( \chi_x \approx 0.4^\circ\)Myr\(^{-1}\). The circulation associated with this vorticity component will be inward towards the y-axis near the upper boundary of the fluid and downward beneath the y-axis. If this circulation has radius, say, 100 km, then the maximum \( v_x \) close to the upper boundary is no greater than \( \sim 1\) mm yr\(^{-1}\). This confirms suggestions made in Subsection 5.2 that \( v_x \) is small compared with \( v_y \).

Further quantitative investigation of the properties of the three-dimensional velocity gradient tensor in three dimensions will require a proper three-dimensional model for velocity beneath Yellowstone and the NE Basin and Range province, which is beyond the scope of this study.

6 DISCUSSION

The \( \sim 35-40\) mm yr\(^{-1}\) southwestern velocity of the North American plate relative to the Yellowstone plume (e.g. Armstrong et al. 1975; Iyer 1984) means that the position of the plume changes by about 35-40 km every 1 Myr relative to the plate. The change in position of the horizontal velocity field caused by the plume will thus be small over a million years in comparison with its overall dimensions. Consequently, the instantaneous pattern of deformation investigated in Section 5 is likely to be a reasonable approximation for the deformation on a time-scale of the order of 1 Myr. I make no attempt to predict quantitatively the sense of the strike-slip earlier than this on any fault, or the sense of vertical vorticity at any point. However, many of the faults in the region that are active now are older than this, and the normal components of throw appear to be consistent with normal slip rates that have been uniform over time for up to \( \sim 10\) Myr. Paleomagnetic observations, as already illustrated for central Greece, can provide a measure of the overall rotation on a long time-scale around a vertical axis in a region. Given the age of a rock unit and the rotation observed in it, an average rotation rate can be calculated. However, the suggested model for the NE Basin and Range Province implies that a region may rotate in one sense when the Yellowstone plume is close to it, but in the opposite sense after the plume has moved further away. For example, the proposed model predicts that the blocks separated by the Lost River fault and the other sub-parallel faults in the same zone are at present rotating clockwise, but would have been rotating anticlockwise a few million years ago when close to the upwelling plume. Blocks may thus rotate through substantial angles in one sense and then in the other sense in quick succession, possibly resulting in very small or zero net rotation. In these circumstances, a time-averaged rotation rate alone is unlikely to be a helpful measure of integrated deformation.

For the model suggested in Section 5 to be valid, the crust beneath the SRP has to be moving SW away from Yellowstone, at the same rate as its surroundings, and hence moving relative to undeformed parts of the North American plate east of Yellowstone. The velocity profile in Fig. 9 suggests that flow in the underlying deforming fluid will act to try to cause relative movement along the length of the SRP. However, any casual inspection of the SRP will show that it is flat and uniform and appears undeformed, with no evidence at the Earth’s surface of any large-scale relative movement. Many people have investigated the crustal structure of the SRP, using seismic refraction, teleseismic delay times, heat flow, and magnetotelluric and other techniques (see Iyer 1984, for a review). It comprises \( \sim 5\) km average thickness of late Tertiary and Quaternary lavas, underlain by thinned continental crust, with some evidence that this crust has been intruded with igneous bodies also in late Tertiary time. Average surface heat flow is \( \sim 200\) mW m\(^{-2}\) (Brott, Blackwell & Ziaogos 1981), much higher than in typical continental crust, suggesting temperature \( \sim 350^\circ\)C at depth \( \sim 5\) km. Brott et al. (1981) also suggested temperature \( \sim 1000^\circ\)C at 25 km depth at the 5 Myr age point close to the southern end of the Lost River Fault. The strength of typical continental crust appears to be controlled by the rheology of quartz, which deforms plastically at geological strain rates when \( > \sim 300^\circ\)C (e.g. Sibson 1983). The observed depth limit of seismicity in many continental regions coincides approximately with this isotherm, suggesting that the base of the brittle layer beneath the SRP is only \( \sim 5\) km deep. In contrast, brittle deformation beneath the fault-bounded blocks in the regions surrounding the SRP continues to \( \sim 15\) km depth, the depth limit of seismicity in these regions, and extension there in this depth range occurs by slip on major faults as already discussed.

Several suggestions may reconcile the observed character of the SRP with the model. First, the entire remaining thickness of continental crust beneath the SRP below the \( \sim 5\) km brittle layer may be too hot to deform seismically, but may be deforming plastically at depth at the rate suggested in the model, without any relative movement occurring at the earth’s surface. Average model \( v_x \) throughout the region SW of a point \( \sim 100\) km SW of Yellowstone is \( \sim 40\) mm yr\(^{-1}\), the same as A. The whole SRP above this zone may be moving SW at velocity A, and local differences between this velocity and the predicted model velocity may be taken by distributed aseismic plastic deformation at depth.

Alternatively, intermittent volcanic activity from cinder cones and fissures along the SRP may allow extension of parts of it. For example the Holocene (\( \sim 1000\) yr ago) eruption in the Craters of the Moon area (e.g. Thompson 1977) occurred along a zone of NNW-trending fissures that is roughly in line with the southern end of the Lost River fault. Opening of fissures in this manner may enable the SRP to extend so as to accommodate the suggested slight
extensional strain rate beneath it. However this process cannot account for the suggested slight shortening strain rate at distances greater than ~500 km from Yellowstone, where model \( v_r \) begins to decrease back to the plate velocity \( A \). As a third alternative, the SRP may distort in a distributed manner that is not readily identifiable.

At present, no observational data exist that can distinguish between these possibilities. Suitable data may be gathered by carrying out repeated geodetic surveys to measure relative motion between points on the SRP and points in undeforming parts of the North American plate NE of Yellowstone and west of the two seismically active zones in Fig. 2. Suggestion 1 predicts no relative motion between points on the Snake River Plain, or between the SRP and undeforming parts of the North American plate. Suggestions 2 and 3 predict relative motion between points on the SRP and between the SRP and its undeforming surroundings. Given that suitable geodetic observations do not yet exist, there is no point speculating further about the kinematics of the SRP.

It is interesting to compare senses of rotation around vertical axes now occurring with those during the crustal shortening phase that formed the ‘Overthrust belt’ of northern Utah, Wyoming and Idaho. When viewed from the SW, the Overthrust Belt is concave; its southern part in northern Utah trends at ~010°, whereas in central Idaho it trends at ~300°. Paleomagnetic investigations (Grubbs & Van der Voo 1976) suggest that during shortening the southern part of this zone rotated clockwise by ~30°, and the northern part rotated anticlockwise, forming this concave shape. The inferred present-day sense of rotation, clockwise in central Idaho and anticlockwise in SE Idaho and SW Wyoming, is straightening out the curvature that now exists in this belt, for reasons unconnected with the original pattern of crustal shortening. This contrasts with what is occurring in the Aegean Sea region, where present-day lateral variation in extension rate is progressively sharpening the overall structural curvature (e.g. Kissel et al. 1986). It is possible that in a few million years the Overthrust Belt will be substantially straightened, and the curvature there now will only be detectable using palaeomagnetic measurements on rocks of suitable age.

The assumption that volume within the deforming fluid is conserved (equation 2.1) appears at odds with the eruption of very large volumes of volcanic rocks around Yellowstone. These total ~4000 km³ over the past 2 Myr, at average rate of production ~2 x 10⁶ m³ yr⁻¹ (R. B. Smith, personal communication, 1987). However, this rate of volume extrusion corresponds to convergence rate across a front 100 km wide (in the y-direction) and 100 km deep at the leading edge of the plate that is in contact with the plume that is only ~0.2 mm yr⁻¹, about 2 orders of magnitude smaller than the relative velocity \( A \). Thus, the volume extruded at Yellowstone is only a very small proportion of the volumes of the plume and the North American plate that are interacting, and conservation of volume may be a good overall approximation.

The model described in Subsection 5.2 shows that a simple circulation can explain a reversal in the sense of vertical vorticity in both deforming zones. Local reversals in the sense of rotation angles about a vertical axis have been observed in some regions. In NW Turkey for example, palaeomagnetic observations (Kissel et al. 1987) suggest that the deforming zone can be subdivided into adjacent regions with opposite senses of rotation. Local reversals in sense of vertical vorticity in the underlying deforming ‘fluid’ there may explain this reversal in senses of rotation. However, an alternative explanation for the observations in NW Turkey, that some blocks are forced to rotate around a vertical axis in the opposite sense to others by frictional interactions at their adjacent edges (Westaway 1989a) cannot be discounted on the limited available evidence.

The two seismically active zones in the NE Basin and Range province are, to first order, symmetrical about the SRP (Fig. 2), and, assuming the average deformation rate and orientation parameters \( H_z \approx 0.6 \) and \( \psi \approx 45° \) apply to both zones, the deduction in Section 5 that \( v_r \) is inwards towards the SRP from both sides is reasonable. However, detailed inspection of Fig. 2 and Table 2 indicates that the two zones are not exactly symmetrical. For the Intermountain seismic belt, \( H_z \approx 0.4 \) and \( \psi \approx 55° \), giving \( v_r \) inward towards the SRP (Table 3), as already suggested. However, \( H_z \approx 0.8 \) and \( \psi \approx 30° \) for the Central Idaho seismic zone gives \( \psi \approx 0 \) or perhaps slightly negative, implying \( v_r \) is either zero or slightly outward from the x-axis beneath this zone (Table 3), in contrast with the sense deduced from the local topographic tilting. Given the limited available data, further discussion concerning whether this apparent asymmetry between the two seismically-active zones is real or merely an artefact of limited sampling, is not worthwhile. Hopefully, future field investigations will attempt to measure the small component of strike-slip as well as the larger and more obvious component of normal slip on major faults in both zones, enabling models for the overall pattern of deformation to be improved.

If the Yellowstone–SRP system is a continental analogy to sheared upwelling mantle plumes observed in oceanic regions (e.g. McKenzie et al. 1980), then the apparent twisting of the regional topography towards the SRP by the x-component of vorticity in the underlying deforming ‘fluid’

### Table 3. Relationships between \( \partial v_r/\partial x \) and \( \partial v_r/\partial y \) for the deforming zones on both sides of the SRP

<table>
<thead>
<tr>
<th>( H_z )</th>
<th>( \psi )</th>
<th>( \partial v_r/\partial x = 0 )</th>
<th>( \partial v_r/\partial y = 0 )</th>
<th>( \partial v_r/\partial y = 0 )</th>
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<tbody>
<tr>
<td>0.1</td>
<td>6°</td>
<td>0.70</td>
<td>0.82</td>
<td>0.89</td>
</tr>
<tr>
<td>0.2</td>
<td>11°</td>
<td>0.49</td>
<td>0.67</td>
<td>0.79</td>
</tr>
<tr>
<td>0.3</td>
<td>17°</td>
<td>0.32</td>
<td>0.53</td>
<td>0.70</td>
</tr>
<tr>
<td>0.4</td>
<td>22°</td>
<td>0.18</td>
<td>0.43</td>
<td>0.62</td>
</tr>
<tr>
<td>0.5</td>
<td>27°</td>
<td>0.07</td>
<td>0.33</td>
<td>0.55</td>
</tr>
<tr>
<td>0.6</td>
<td>31°</td>
<td>-0.02</td>
<td>0.25</td>
<td>0.49</td>
</tr>
<tr>
<td>0.7</td>
<td>35°</td>
<td>-0.10</td>
<td>0.18</td>
<td>0.42</td>
</tr>
<tr>
<td>0.8</td>
<td>39°</td>
<td>-0.16</td>
<td>0.11</td>
<td>0.37</td>
</tr>
<tr>
<td>0.9</td>
<td>42°</td>
<td>-0.22</td>
<td>0.06</td>
<td>0.32</td>
</tr>
<tr>
<td>1.0</td>
<td>45°</td>
<td>-0.26</td>
<td>0.00</td>
<td>0.26</td>
</tr>
</tbody>
</table>

For each value of \( H_z \), a single slip vector azimuth \( \psi \) relative to the 230° azimuth of the SRP exists for which \( \partial v_r/\partial x \) is zero. For other representative values of \( \psi \), 30, 45 and 60°, the ratio \( \partial v_r/\partial x \)/\( \partial v_r/\partial y \) is calculated for each value of \( H_z \). Because the condition \( \psi \ll v_r \) was used to obtain equation (5.12) as part of the procedure for obtaining these relationships, solutions where \( \partial v_r/\partial x \) and \( \partial v_r/\partial y \) are comparable are likely to be only crude approximations.
may have a counterpart in oceanic regions. An axial
topographic low with width ~100 km would be expected
along the trailing direction of each sheared plume. Because
McKenzie et al. (1980) applied a Gaussian filter with 150 km
half-width to the ocean floor topography, any axial low as
narrow in the y-direction as the SRP would have been
masked. Investigation of topography over sheared plumes in
oceanic regions using a sharper filter may be worthwhile.
Although oceanic lithosphere may be too strong to twist in
this manner, if axial lows could be identified they would
provide further confirmation that the suggested analogy is
valid, as well as providing a useful method for distinguishing
the leading and trailing directions of sheared plumes
beneath oceanic regions where the absolute plate motion
direction has not been well determined.

Zones of distributed extension of similar scale to those in
the NE Basin and Range Province exist in various other
parts of the world, and may also indicate deformation linked
to upwelling mantle plumes. Examples include the various
branches of the East African rift system, the Baikal rift
system in Siberia, and the Shansi graben system in central
China (see, e.g. Dunkelman, Karson & Rosendahl 1988 and
Molnar et al. 1981, for location maps). Because each
observable element of the strain rate and rotation tensors
depends on the difference between two velocity gradient
tensor elements, small changes in these elements may lead to
major changes in observables, including reversals of sign.
Thus differences in shape between the deforming zones in
the NE Basin and Range Province and in these other
regions do not necessarily make this suggested explanation
inapplicable. However, because these other regions have
been studied in much less detail than the NE Basin and
Range Province, and their senses and rates of deformation
are not known with confidence, this suggestion cannot yet
be tested.

7 CONCLUSIONS

The northern part of the N-S trending Intermountain
seismic belt and the E-W trending Central Idaho seismic
zone are bisected by the NE-SW trending SRP, regarded as
the trail left behind by the Yellowstone plume as the North
American plate moves SW across it. Available seismological
and geomorphological evidence suggests that major normal
faults in the upper-crustal brittle layer more than ~150 km
from Yellowstone move with small components of oblique
slip, implying ~4° Myr⁻¹ vertical vorticity that is clockwise
beneath the Central Idaho seismic zone and anticlockwise
beneath the Intermountain seismic belt. Slip vector azimuths
ψ in large earthquakes are inward in both zones at ~45°
to the axis of the SRP, the +x-axis. Nearer Yellowstone,
major faults in both zones show the opposite sense of
oblique slip, suggesting that vertical (z) vorticity may
change sign close to Yellowstone in both zones. Most major
faults in both zones more than 150 km from Yellowstone dip
W or SW, and assuming their rotation around horizontal
axes follows the ‘domino’ model (equations 3.2–3.4), this
rotation implies horizontal (y) vorticity in the underlying
deforming ‘fluid’ oriented between N and NW. Topographic
elevation in both zones decreases towards the SRP.
Assuming this topography is caused by torsion of the Earth’s
surface by a second horizontal (x) vorticity component in
the underlying deforming fluid, it implies this component is
oriented SW when SE of the SRP and oriented NE when
NW of it.

The North American plate is moving SW relative to
Yellowstone. Like similar upwelling plumes in oceanic
regions, the plume beneath Yellowstone is likely to be
sheared in the direction of this plate motion. This shearing
of the plume can be regarded as a rotation around the
+y-axis only, and can account for the opposite sign of both
the x- and z-components of vorticity on both sides of the
plume, as conservation of angular momentum constrains the
total angular momentum should the x- and z-axis to be zero.
The y-component of vorticity is finite when summed
throughout the NE Basin and Range Province beyond
~100 km from Yellowstone, giving finite (equation 5.11)
angular momentum around the –y-axis. Conservation of
angular momentum requires this to be equal and opposite to
the angular momentum in the part of the plume beneath
Yellowstone that is deflected from vertical to predominantly
horizontal motion (equation (5.12)). Conservation of
y-component of angular momentum predicts upwelling
plume radius ~50 km and upwelling velocity ~60 mm yr⁻¹
(equation 5.13), consistent with an independent method
using Stokes’ law (equation 5.3).

Extensional strain rate E₁ ~10⁻¹¹ s⁻¹ in both zones,
revealed by Holocene scarp heights and by the throw on
major normal faults given their age. Tilt rates of individual
fault-bounded blocks suggest y-component of vorticity is
~4° Myr⁻¹ in both zones, oriented along the –y-axis
beneath both zones. Topographic tilting towards the SRP is
~1° at its 5 Myr age point (Fig. 9) suggesting that the
x-component of vorticity is ~0.4° Myr⁻¹. Relative sizes of
vertical vorticity χ₃ and principal extensional strain rate E₁
can be expressed as a dimensionless vertical Holmes number
Hᵢ or χ₃/(2E₁). Observed slip vectors on sets of parallel
faults in both deforming zones constrain |Hᵢ| ~0.6
independently of both quantities when beyond ~150 km
from Yellowstone. Given the estimated value for E₁,
χ₃ ~4° Myr⁻¹ is predicted. Given that E₁ is oriented
towards the +x-axis at ψ ~45° at these localities, E₁ = Eₓ₉,
the off-diagonal strain rate tensor element. Provided
0.5 < Hᵢ < 1 (Table 3), gradients of uₓ are small compared
with gradients of uᵧ in both deforming zones.

Symmetry considerations require uᵧ zero on the y-axis,
and it can be constrained zero everywhere on the upper
boundary of the deforming fluid to establish an approximate
two-dimensional kinematic model for the horizontal velocity
field there. Near Yellowstone ∂ₓ/∂y dominates ∂ᵧ/∂x in
determining vertical vorticity. In the +y quadrant,
∂ₓ/∂y > 0 and in the –y quadrant ∂ₓ/∂y < 0. E₁ sub-parallel to the SRP near Yellowstone is primarily caused
by the strong positive ∂ₓ/∂x velocity gradient in both
quadrants. These velocity gradients are caused by the local
acceleration of the plume in the +x direction by its
interaction with the overlying and surrounding lithosphere.

Observed sense of vertical vorticity beyond ~150 km
south and west of Yellowstone can be explained with
∂ₓ/∂y < 0 in the +y quadrant and ∂ₓ/∂y > 0 in the –y quadrant. Observed principal extensional strain rate
azimuth agrees with the model prediction, although predicted strain rate and vertical vorticity overestimate
observed values by factor up to ~2. This discrepancy may be
partly because strain rates from average Holocene fault slip rates do not take account of time since the last major earthquake (equations 3.5 and 3.6) and may substantially underestimate true strain rate. Predicted vertical Holmes number $H_v$ agrees well with observations.

Inferred senses of the two horizontal vorticity components agree with the likely signs of vertical gradients of horizontal velocity components, and suggest that these dominate the effect of horizontal gradients of vertical velocity. The observation that faults in the two deforming zones have predominantly normal slip, with $H_v \sim 0.6$ and $\psi$ inwards at $\sim 45^\circ$, implies $v_z$ inwards at the upper boundary of the deforming fluid in both zones. The sense of the $x$-component of vorticity suggests this inward flow downwells beneath the SRP (Fig. 13). This inward movement may contribute to the magma supply for SRP basalts, and may explain why these rocks do not show the characteristics expected for a source within a plume from the deep mantle (e.g. Thompson 1977).

These preliminary kinematic results suggest that it is realistic to consider the actively deforming zones on both sides of the SRP in the NE Basin and Range Province as an effect of the interaction between the Yellowstone plume and the moving North American plate. They suggest that this problem warrants further investigation, perhaps by combining detailed field, geodetic and paleomagnetic studies aimed at determining the senses and rates of slip on faults and rotation of fault-bounded blocks with the use of more sophisticated methods to examine the dynamics of the region and to obtain proper three-dimensional solutions for the velocity, vorticity and strain rate fields. Many of the interrelationships discussed between strain rate tensor elements and vorticity components will be necessary features of such solutions, and will constrain more sophisticated models for this region.

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