A simple method for estimating the horizontal velocity field in wide zones of active deformation—I. Description, with an example from California

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SUMMARY
The velocity field is the most complete description of the active kinematics of a wide zone of active deformation. A simple method is described here for estimating the horizontal velocity field in a network of traverses that span the zone. The sum of the relative velocities in such a zone must equal the relative motion of the boundaries, which can usually be determined independently from sea-floor magnetic anomalies. Also, if traverses intersect each other, the velocity of the point of intersection relative to one of the boundaries of the zone, must be the same regardless of the traverse segment along which relative velocities have been summed. In addition, rates of deformation within parts of the deforming zone can be estimated from field observations. All these constraints can be expressed as linear equations. The aim is to maximize the difference between the number of equations (N) and unknowns (M), with the formed exceeding the latter. The problem can be expressed in matrix form: \( \mathbf{A} \mathbf{m} = \mathbf{d} \), where \( \mathbf{A} \) is an \( N \times M \) model matrix of constraints, \( \mathbf{d} \) is a column vector of data, and \( \mathbf{m} \) is the solution vector of unknowns. If \( N > M \), a solution for \( \mathbf{m} \) can be determined by minimizing a function of \( \| \mathbf{A} \mathbf{m} - \mathbf{d} \| \), using standard techniques and taking into account any uncertainty in the data vector \( \mathbf{d} \). This method is used to estimate the velocity field in the Californian plate-boundary zone. A weighted least-squares solution is found that predicts tectonic rotations. Simple tests illustrate the effect on the derived velocity field of changing features of the velocity model.

Key words: deformation, USA.

INTRODUCTION
Whereas oceanic lithosphere is strong and forms rigid plates surrounded by narrow deforming belts, continental lithosphere is relatively weak, with very large areas of distributed deformation between the rigid plates (Molnar & Tapponnier 1975; Tapponnier & Molnar 1976; England & McKenzie 1982; McKenzie & Jackson 1983). The mountain ranges of the Andes on the western margin of South America, and the Alpine–Himalayan belt, including the whole of central Asia between the Indian and Siberian shields, are the most extensive examples of this, where deformation occurs in a zone 100's to 1000's of kilometres across, and 1000's of kilometres long. The description of the deformation in such large regions is an important problem in geology, and provides insight into how continental lithosphere responds both to the relative motion of the boundaries and also body forces generated by the topography within the deforming zone.

Major active faults, with displacement rates that are a significant fraction of the magnitude of relative plate motion, have been recognized in many wide zones of continental deformation. Some of these faults, such as the Altyn Tagh fault in Tibet (Molnar & Tapponnier 1975; Tapponnier & Molnar 1976, 1977), San Andreas fault in California (Wallace 1990), and Alpine and other faults in New Zealand (Wellman 1955), can be traced for 100's km and form major features of the deforming zone. For these reasons, a considerable amount of effort has gone into characterizing these faults and determining their slip rates for the Holocene and Pleistocene. However, it has long been recognized that slip on a few major faults, for instance in New Zealand (Walcott 1978) and California (Luyendyk 1989), does not account for all the motion in the deforming zone. The remaining motion is accommodated by both rigid-body rotation about a vertical axis of the intervening crustal blocks, and diffuse small-scale faulting, including rotation about horizontal axes. Diffuse faulting is difficult to quantify in detail, but can be measured in a bulk sense, if all the deformation in the brittle crust is expressed seismically,
by summing seismic-moment tensors (Jackson & McKenzie 1988; Ekstrom & England 1989). In addition, the deformation over wide regions can be measured geodetically over periods of years to 10's of years.

The complex displacements and rotations of the finite deformation in wide zones of deformation are difficult to analyise. However, the short-term deformation in a large region, over periods less than 100 000 yr, can be described in terms of a velocity field, which shows the velocities of points in the deforming zone relative to an external frame of reference, such as one of the bounding plates. This is the most complete description of the active deformation in a deforming zone, and in this paper I discuss ways of estimating the horizontal velocity field from the available information.

**Velocity field in a continuum**

In a uniformly and continuously deforming zone, the variation with position of velocities, with components \( u, v, w \), relative to orthogonal axes \( x, y \) (horizontal) and \( z \) (vertical) respectively, is described by the velocity gradient tensor (Malvern 1969).

\[
\begin{pmatrix}
    \frac{\partial u}{\partial x} & \frac{\partial u}{\partial y} & \frac{\partial u}{\partial z} \\
    \frac{\partial v}{\partial x} & \frac{\partial v}{\partial y} & \frac{\partial v}{\partial z} \\
    \frac{\partial w}{\partial x} & \frac{\partial w}{\partial y} & \frac{\partial w}{\partial z}
\end{pmatrix}
\]

Assuming incompressible deformation (\( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \)), eight independent variables are needed to describe a constant general velocity field. Four independent variables are needed to describe a constant general horizontal velocity field (\( \frac{\partial u}{\partial x}, \frac{\partial u}{\partial y}, \frac{\partial v}{\partial y}, \frac{\partial w}{\partial x} \)). It is often convenient to talk about the components of strain rate:

\[
\begin{align*}
\dot{\gamma}_1 &= -\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}; \\
\dot{\gamma}_2 &= \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}; \\
T &= -0.5\left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right), \\
W &= \frac{\partial u}{\partial y} - \frac{\partial v}{\partial x},
\end{align*}
\]

where \( 2T \) and \( W \) are the horizontal dilatation and vertical vorticity respectively, and \( \dot{\gamma}_1 \) and \( \dot{\gamma}_2 \) are the two components of shear-strain rate. The horizontal dilatation rate describes the rate of change of surface area per unit surface area (\( +ve \) increasing, \( -ve \) decreasing). If the velocity field is equivalent to that of a rotating rigid body, the vertical vorticity will be twice the angular velocity of the rigid body.

If the components of strain rate are constant, then the horizontal components of velocity \( u, v \) of any point, with coordinates \( x, y \) can be specified in terms of them (Fig. 1):

\[
\begin{align*}
 u &= \frac{1}{2}[(\dot{\gamma}_1 - 2T) \cdot x + (\dot{\gamma}_2 + W) \cdot y] \\
u &= \frac{1}{2}[(\dot{\gamma}_2 - W) \cdot x + (-2T - \dot{\gamma}_1) \cdot y].
\end{align*}
\]

**Velocity field in the brittle crust**

The velocity gradient tensor in eq. (1) applies to a continuum. However, the brittle crust in an active zone of deformation is broken up by faults. One can consider the velocity field of such a region at a number of scales, both in time and space. At a very small length scale, such as the spacing of individual small faults, the velocity is likely to vary, depending on the time period chosen. At time-scales that are much less than the seismic cycle, this velocity field is likely to be continuous, reflecting the local accumulation of elastic strain, and can be described by a spatially varying velocity gradient tensor. However, at longer time-scales, the velocity field will be complex with abrupt changes as a consequence of slip on individual faults.

If the velocity field is viewed on a length-scale that is much greater than the spacing between individual small faults, then the local spatial and temporal effects of the build-up of elastic strain may not be apparent (Walcott 1984). In this case, this large-scale velocity field can also be described by a velocity gradient tensor, such as that in eq. (1), as if the crust deforms continuously. Of course, this description will not be applicable to the detailed behaviour of the deforming zone at smaller length-scales. In addition, very large-scale faults within the deforming zone, which accommodate a significant proportion of the motion, will still form discontinuities. In this case, we can divide the deforming zone into regions that at the chosen length-scale deform continuously, characterized by a velocity gradient tensor, and bounded by discontinuities which represent the large-scale faults.

**Determination of the velocity field**

If a deforming region is treated in isolation, all four components of horizontal strain rate \( \dot{\gamma}_1, \dot{\gamma}_2, 2T, W \) in eq. (2) are needed to fully define the horizontal velocity field. Direct geodetic measurement of the deformation, using satellite techniques such as the global positioning system (GPS), will yield the full velocity field relative to some specified frame of reference. These surveys are made over a short period (2 to 5 yr), and have as yet only been carried out in a few zones of active deformation. Unfortunately, it is
generally not possible with other measuring techniques, which have been made over much longer periods (10's to 100's of years), and are more readily available, to directly determine the full velocity field. For instance, measurement of angle changes in repeated triangulation surveys will only yield the two components of shear-strain rate \( \gamma_1 \) and \( \gamma_2 \). We can determine the bulk dilatation rate \( \tau \) from earthquakes, as well as the two components of shear-strain rate, if the deformation is accommodated by seismic slip on planar faults (Jackson & McKenzie 1988). Thus, in the absence of GPS measurements, the determination of the vertical vorticity remains difficult without additional information (Jackson & McKenzie 1988), and is the most important stumbling block in estimating the velocity field in a zone of deformation. This is because slip on a single planar fault can only give information about relative translation, while the determination of the vertical vorticity requires information about the rotation of the fault about a vertical axis, relative to some fixed frame of reference.

In principle, if deformation is accommodated solely by slip on numerous faults, and all the faults and their slip vectors are known right up to the boundary of the zone, one must have a complete description of the deformation, relative to this boundary, including any internal rotation accommodated by slip on these faults. Haines (1982) developed this idea by showing that if a deforming zone can be approximated as a continuum, and one boundary of the deforming zone is specified, then the variation in the \( \gamma_1 \) and \( \gamma_2 \) components of shear-strain rate, right up to the boundary of the zone, gives information about both the dilatation and vertical vorticity relative to this boundary. If the distribution of the dilatation rate is known in advance, then the vertical vorticity can be determined everywhere relative to a point. The local components of average shear strain and dilatation generated by seismic slip on a fault can be deduced from the earthquake-moment tensor. Thus, Holt & Haines (1993) and Jackson, Holt & Haines (1992) fitted smooth distributions of the components of shear-strain rate and dilatation rate to the pattern of seismic-moment release this century in Western Asia and the Aegean, and determined the velocity field relative to one of the bounding plates, accommodated by seismic faulting.

In practice, Haines’s (1982) method is not straightforward to use as it requires a knowledge of the spatial distribution of the components of the strain-rate tensor. In inverting for the velocity field, smooth functions are fitted to the rates of strain. Available observations of the strain-rate tensor in wide zones of deformation, either from geodetic measurements or the summation of seismic moment tensors, are generally incomplete, and thus there is an element of approximation and interpolation in fitting this data to continuous functions of position. In addition, it may not be straightforward to take into account the effects of slip on individual large faults, which may be known from displacements of Holocene markers but may not be represented in the short-term seismic record.

**SIMPLE VELOCITY MODEL**

I describe here a simple method of estimating the horizontal velocity field, which is intended to complement Haines’s (1982) rigorous method, providing a first approximation of the general features of the velocity field. In particular, it can be used as a tool for investigating the effect of specific patterns of faulting on the velocity field. The key element of the method is the use of the constraints provided by both the spatial variation in deformation, and the total motion across the deforming zone. Regardless of the path used to cross a deforming zone, the sum of all the deformation along the path must be equal to the total motion across the deforming zone (Fig. 2a). The latter may be equal to the relative plate motion, which is known independently from ocean-floor magnetic anomalies to within a certain precision.

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**Figure 2.** (a) Diagram illustrating features of a general continental plate-boundary zone. Relative motion between plates A and B is accommodated in a zone, 100's to 1000's kilometres wide with irregular margins. The effect of the deformation crossed by any path across the plate-boundary zone, between points A and B, will be to accommodate the relative plate motion between these points. However, if the two bounding plates are rotating relative to each other, the relative plate motion will vary between different points on the bounding plates. For instance, the motion of point B, relative to plate A, will not be the same as that of point C. (b) Diagram illustrating how the relative horizontal motion between two points, A and B, can be divided into two components. One component is parallel to the line joining the two points, and describes any shortening or extension between the two points. This component is an intrinsic feature of the deformation between the two points, and is independent of the frame of reference in which the relative motion is observed. The second component is orthogonal to the line joining the two points. This component can be considered as a consequence of the rotation of this line, relative to a fixed frame of reference. Therefore, this component depends on the frame of reference chosen, such as one of the bounding plates.
These constraints can be used to determine unknown aspects of the deformation, such as rigid-body rotations about vertical axes, which contribute to the horizontal velocity field.

A velocity model is developed in a network of traverses that span the deforming zone, which is approximated as a flat plane. The traverses may cross, either regions of diffuse deformation, described by a velocity gradient tensor, or major faults.

Relative velocities

The description of the deformation is reformulated in terms of relative velocity horizontal vectors across segments of the deforming zone. The relative horizontal velocity vector between any two points in the deforming zone can be considered in terms of two components (Fig. 2b). There will be a component of motion that is parallel to the line joining the two points. This component is a consequence of either local extensional or compressional strains in the zone, and is an intrinsic feature of the deformation and is not dependent on the choice of a frame of reference. It can usually be estimated from studies of the internal deformation, such as faulting or folding. The other component of relative motion is orthogonal to the line-joining points. If the points are widely separated, then any rotation about a vertical axis of the material in the deforming zone, relative to a fixed frame of reference, will contribute to this component. This component is dependent on the choice of frame of reference. Thus, if one of the bounding plates is chosen as a frame of reference, the orthogonal component cannot be determined from studies of the internal deformation, without reference to this frame of reference. For this reason, this component of motion is generally not known.

A simple illustration may help to clarify the significance of the choice of the frame of reference when determining the relative velocity between two points. Imagine two children, Sarah and Simon, who are riding on the Big Wheel at the fair (Fig. 3). From Sarah’s point of view, Simon is stationary as he is riding on the same wheel that she is also on.
(ignoring the local movement of the swing seats). However, the rest of the world will appear to rotate anticlockwise around her (Fig. 3b). In other words, the velocity of Simon relative to Sarah, in a frame of reference fixed to the rotating wheel, is zero. However, they are being watched by their mother who is standing on the ground nearby. She sees Simon moving past the crest and is beginning to descend. So, to their mother, the velocity of the Big Wheel rotating clockwise (Fig. 3a). Sarah is still rising up, while Simon has moved past the crest and is about to descend. So, to their mother, the velocity of Simon relative to Sarah is the vector difference of the individual velocities of Simon and Sarah (Fig. 3a). In this case, in the frame of reference of his mother on the ground, Simon is moving relative to Sarah with a velocity that has a large downward component.

**Velocity constraints**

The velocity of any point in the network of traverses, relative to one of the margins, must be the sum of the relative velocities between that point and the margin. Thus, the velocity of a point, relative to one of the boundaries, at the intersection of traverses, must be the same, regardless of which traverse is used to sum relative velocities. The total velocity across the zone of deformation must be the sum of all the internal relative velocities. These constraints can be expressed as linear vector equations. If enough linear equations exist, then these can be solved for the unknown components of relative velocity, and hence the velocity field in the zone of deformation.

**Slip on a single fault**

The simplest boundary between two moving plates, A and B, is a single fault (with orientation \( \alpha \)). It is clearly a trivial matter to solve for the slip on the fault. If the slip vector (\( \mathbf{v}_f \) in Fig. 4) is resolved parallel to orthogonal axes \( x, y \) in Fig. 4, then the following equations must hold:

\[
\begin{align*}
\mathbf{v}_x &= V_{ab} \cos(\alpha) \\
\mathbf{v}_y &= V_{ab} \sin(\alpha)
\end{align*}
\]

where \( V_{ab} \) is the local magnitude of the relative plate velocity vector, for plate A relative to plate B, and \( \mathbf{v}_x \) and \( \mathbf{v}_y \) are the horizontal components of the slip rate on the fault.

**Diffuse deformation bounded by slip on faults**

If the boundary zone between two moving plates, A and B, is wide, then clearly deformation must be more than slip on a single fault. For instance, the relative motion between the plates may be accommodated by a combination of slip on major faults and more diffuse or distributed deformation in an intervening region (Fig. 5a). This example can be analysed by considering the deformation along two traverses that span the deforming zone, choosing any orientation for the traverses. We can then consider relative velocities along the traverse, which may either be the slip vectors on the faults or the effect of the diffuse deformation. Let \( \mathbf{v}_i \) refer to the \( j \)th relative horizontal velocity vector (fault slip vector or result of diffuse deformation) along the \( i \)th traverse. Relative velocity vectors are in a frame of reference fixed to one of the bounding plates. Thus, this relative velocity vector will vary, depending on whether it is defined in a frame of reference fixed to plate A or plate B. This is because any rigid body rotation about a vertical axis of the traverse segment, relative to an external frame of reference such as one of the bounding plates, may contribute to \( \mathbf{v}_i \). If \( \mathbf{v}_{ij} \) is defined in a frame of reference fixed to plate B, then the overall constraint of the relative plate motion must be satisfied (Fig. 5a, b):

\[
\sum_i \mathbf{v}_i = \mathbf{v}_{abi}
\]

where \( \mathbf{v}_{abi} \) is the velocity of plate A at the end of the \( i \)th traverse, relative to plate B. Furthermore, if plate A is rotating about a vertical axis relative to plate B with a rate of rotation \( \omega_{abi} \), then the relative plate velocities across traverses 1 and 2 are related by:

\[
\mathbf{v}_{ab12} = \mathbf{v}_{ab1} + d\omega_{abi} \mathbf{n}_a
\]

where \( \mathbf{n}_a \) is a unit vector perpendicular to the margin of plate A, and \( d \) is the distance along this margin between traverses 1 and 2 (Fig. 5a). The relative velocities must also satisfy a compatibility equation which relates the velocities in traverse 1 to those in traverse 2 along the linking tie-line 3 which intersects these traverses at points a fraction \( \alpha, \beta \) (for any \( \alpha, \beta \)) along traverses 1 and 2, respectively. If \( \mathbf{v}_{31} \) is the relative velocity across the tie-line linking traverse 1 and 2 (Fig. 5b), in a frame of reference fixed to plate B, then:

\[
\mathbf{v}_{11} + \alpha \mathbf{v}_{12} + \mathbf{v}_{31} = \beta \mathbf{v}_{22} + \mathbf{v}_{21}.
\]

The relative velocity vectors across any segment of the network can be specified by two orthogonal components. I assume that the diffuse deformation can be described by a uniform horizontal velocity gradient tensor, with components of strain rate \( \gamma_1, \gamma_2, T \), and vertical vorticity \( W \) (positive \( W \) is clockwise). Eq. (3) can be used to specify the components of velocity across the zone of diffuse deformation in terms of the components of the velocity gradient tensor, relative to plate B. For instance, for the \( i \)th traverse with length \( d_i \) and orientation \( \phi_i \), the \( x \) and \( y \) components (east, north) of relative velocity across the zone of diffuse deformation have the form (Fig. 1):

\[
\begin{align*}
\mathbf{v}_i(x \text{ component}) &= \frac{1}{2} \cdot d_i \cdot [(\gamma_1 - 2T) \cdot \cos(\phi_i) + (\gamma_2 + W) \cdot \sin(\phi_i)] \\
\mathbf{v}_i(y \text{ component}) &= \frac{1}{2} \cdot d_i \cdot [(\gamma_2 - W) \cdot \cos(\phi_i) + (-2T + \gamma_1) \cdot \sin(\phi_i)].
\end{align*}
\]
Figure 5. Diagram illustrating how the relative plate motion between Plate A and B is accommodated in a wide zone consisting of a region of diffuse or distributed deformation, bounded by two major faults. (a) Two traverses (1 and 2) cross the plate-boundary zone, with a linking tie-line. $V_{ab1}$ and $V_{ab2}$ are the velocities of Plate A, relative to Plate B, across traverses 1 and 2. The dimensions of the network are defined by lengths $d$, $d_1$, $d_2$, $d_3$. All velocities can be resolved parallel to orthogonal axes $x$, $y$. A unit vector which is orthogonal to the northern boundary of the zone is defined by $n_a$. (b) The relative velocities across traverse segments are also shown: $v_{ij}$ denotes the relative horizontal velocity across the $j$th segment of the $i$th traverse; $v_{11}$, $v_{13}$, $v_{21}$, $v_{23}$ define slip vectors on the bounding faults, and $v_{12}$, $v_{22}$, $v_{33}$ define relative velocities across the region of diffuse deformation.

Thus, for the three relative velocities $(u_{12}, v_{22}, v_{33})$ eq. (8) yields six equations. In addition, the vector eqs (5)–(7) can be expressed in terms of two orthogonal components generating a further eight equations (four from eq. 5 for traverses 1 and 2, two from eq. 6, and two from eq. 7 for tie-line 3). If the dimensions and shape of the network are known, then there are in total 23 unknowns (18 velocity components, four components of the velocity field for the diffuse deformation, and the relative plate angular velocity) and 14 equations. The significance of this disparity between the number of equations and unknowns is discussed below.

Distributed faulting and block rotation

A zone of distributed deformation may consist of a number of well-defined faults. In this case, the horizontal deformation can be described in terms of slip on the faults and rotation about a vertical axis of the intervening fault blocks. Thus, if the $i$th traverse in our network crosses such a zone between two plates A and B (Fig. 6), then we can replace eq. (5) with:

$$\sum_{i} v_{ij} + d_i \omega_i n_i = V_{abi}$$

(9)

where $v_{ij}$ is the horizontal slip vector on the $j$th fault ($u_{ij}$ is the horizontal slip rate), $\omega_i$ is the rotation rate about a vertical axis relative to plate B of the intervening fault blocks, which is presumed to be the same for all blocks. $V_{abi}$ is the velocity of plate A relative to plate B in the $i$th traverse, and $n_i$ is the unit vector that is orthogonal to the $i$th traverse. The length of the $i$th traverse is $d_i$. Note that the reference frame for $\omega_i$ is the same as that for the relative plate velocity $V_{abi}$, which in this case is plate B. In principle, $\omega_i$ can be determined from palaeomagnetic studies and is therefore of considerable interest.
Thus, if there are sufficient estimates, so that \( N \) is greater than \( M \), \( A \) will have more rows than columns and eq. (10) becomes an overdetermined problem. A solution for \( m \) might be expected to minimize some function of \( \| A \cdot m - d \| \), taking into account the uncertainty associated with the elements of \( d \). Menke (1989) shows that weighted least-squares minimization of this has the form:

\[
\mathbf{m} = [A^\top W A]^{-1} A^\top W d
\]

where \( A^\top \) is the transpose of \( A \), and \( W \) is a square matrix with only non-zero diagonal components, which are the reciprocals of the variances of the components of \( d \). The covariance matrix for the weighted least-squares solution is:

\[
\text{cov} \, \mathbf{m} = \sigma^2 [A^\top W A]^{-1}
\]

where \( \sigma^2 \) has the form:

\[
\sigma^2 = 1/(N - M) \cdot (Am - d)^\top W(Am - d).
\]

The predicted variances of the components of the solution vector \( \mathbf{m} \) should be less than those for the \textit{a priori} estimates used to determine the solution. If this is not the case, then there may be a problem with the model. We can use \( \sigma^2 \) to assess how well the solution vector (\( \mathbf{m} \)) fits the model (\( A \)) to the data (\( d \)). We can informally define the mean weighted error or degree of fit ‘\( F \)’ of the solution as:

\[
F = (N - M)/N \cdot \sigma^2.
\]

Thus, a large value of \( F \) implies a poor fit between the model and data, and a small value of \( F \) implies a good fit. If \( F \) is zero, then the fit is perfect. Also, we can express eq. (12) as:

\[
\mathbf{m} = A^{-x} \cdot \mathbf{d}
\]

where \( A^{-x} \) is the effective inverse of \( A \) used to determine the solution. Thus, it is possible to assess the sensitivity of the solution for \( \mathbf{m} \) to the components of \( \mathbf{d} \) by examining the components of the effective inverse matrix \( A^{-x} \).

**Geological example**

In the following sections, I illustrate the method described above, by using it to assess the active deformation in the wide and actively deforming plate-boundary zone of western North America in California. In particular, it may help to answer two closely related questions: (i) given the observed pattern of faulting, should one expect significant tectonic rotations about vertical axes; and (ii) can one account for observed tectonic rotations with our present understanding of the pattern of faulting? It is clear, that in any real zone of deformation, a realistic description will generate a large number of unknown parameters that must be solved for. The method in this case becomes somewhat unwieldy. In addition, the computer time necessary to solve the problem is proportional to a power of the number of unknowns. Thus, I attempt to simplify the zone of deformation and only examine the effects of the dominant structures. If a simplified conception of the zone of deformation is insufficient to answer the questions above, then it probably means that an important aspect of the deformation has been overlooked.
The 'bend region' of the San Andreas fault in California is part of a complex zone of active deformation at least 300 km wide (Fig. 7a), referred to as the San Andreas fault system. Here, structures in the plate-boundary zone between the Pacific and North American plates are at a high angle to both the general trend of the plate-boundary zone, further north and south, but also the relative plate velocity vector. In addition, clockwise rotations about vertical axes have been observed from palaeomagnetic studies (Luyendyk 1991), which are not apparent in the plate-boundary zone to the immediate north and south. There is an enormous body of work on the tectonics of this part of California, which has recently been summarized (Wallace 1990). Also, new geodetic studies, including very long baseline interferometry, have determined velocities over the last few years in this region relative to the bounding plates (Donnellan et al. 1993; Feigl et al. 1993). However, the interpretation of these results is complicated by the problem of distinguishing between short-term effects related to the build-up of elastic strain and the long-term motions. The consequences and causes of the tectonic rotations in this area remain unclear and a number of models have been proposed.

![Figure 7](https://academic.oup.com/gji/article-abstract/119/1/297/568506/)

**Figure 7.** (a) Map of the western part of North America, in the region of bend in the San Andreas Fault, California. North and south of the bend, the San Andreas fault trends nearly 135°, and accommodates the bulk of the relative plate motion. In the bend region, complex faulting occurs both to the west, making up the west-trending Western Transverse ranges, and to the east in the Mojave Desert. Various estimates of the Holocene slip rate (mm yr⁻¹) on major structures are shown in brackets. (b) Location of traverses across deforming region which are used in a velocity model. Velocities in three traverses across the deforming zone (traverse 1, 2 and 3) and a linking tie-line (traverse 4) are analysed.
proposed (Jackson & Molnar 1990; Luyendyk 1991). Therefore, a question of some interest is whether the observed tectonic rotations are explicable in terms of our present understanding of the first-order active kinematics in this part of the Californian plate-boundary zone.

I address this question by examining the deformation in a number of traverses crossing the plate-boundary zone.

**Velocity model**

I simplify the zone to four regions, characterized by both distributed deformation and slip on straight discontinuities or faults (Fig. 7a), including the San Andreas and Garlock faults. Thus, we can consider the 'bend region', where the San Andreas fault trends ESE, as bounded to the west by a west-trending zone comprising the Western Transverse ranges, and to the east by the Mojave region. North and south of the 'bend region', structures trend more nearly NW. In addition, fault-plane solutions and field observations suggest that, as a first-order approximation, we can make the following simplifying assumptions.

(1) The San Joaquin Valley region, east of the northern part of the San Andreas fault, is treated as effectively the North American plate. Thus, I use NUVEL-1 (DeMets et al. 1990) to constrain the total motion across the deforming zone, making a small correction for the deformation (11 ± 1 mm yr⁻¹ extension in a N36°E ± 2 W direction, Argus & Gordon 1991) in the Basin and Range province, even further east.

(2) The regions to the west of the San Andreas fault, north and south of the 'bend region', are deforming by dextral simple shear parallel to the San Andreas fault. However, in the north, the San Andreas fault itself is treated as having a component of shortening orthogonal to its trend. Thus, there are no velocity gradients parallel to the San Andreas fault in these regions.

(3) Slip on the San Andreas fault in the 'bend region' is pure right lateral.

(4) Deformation in the Western Transverse ranges, to the west of the 'bend region' is accommodated on a series of west-trending sinistral reverse faults, with slip vectors orthogonal to the trend of the San Andreas fault in the 'bend region' (Jackson & Molnar 1990), and there is no length change parallel to the trend of the Western Transverse ranges. Also, fault blocks in the Western Transverse ranges may be rotating about a vertical axis. It is assumed that the entire length of the Western Transverse ranges (circa 250 km) rotates about a vertical axis at a uniform rate.

(5) The San Andreas fault in the 'bend region' is treated as a passive marker, which rotates about a vertical axis at a rate determined by the internal deformation in the Western Transverse ranges to the west.

(6) The Mojave block, in the vicinity of the traverses, forms a rigid triangular zone. Note that the prominent active NW-trending faults in the Mojave region are found further east of the region studied here. In addition, slip on the Garlock fault, which bounds the Mojave region in the north, is pure left lateral where it is crossed by the network. The Mojave block rotates about a vertical axis at the same rate as the San Andreas fault in the 'bend region'.

**Equations**

With the simple assumptions outlined above, I have developed a velocity model in three traverses which span the plate-boundary zone, with a linking tie-line forming a fourth traverse (Figs 8 and 9). Traverse 2 is close to the northern margin of the Western Transverse ranges. The assumptions

![Figure 8. Detailed view of network of traverses, illustrated in Fig. 7(b), showing the definition of various dimensions in the network (d₁ - d₉). The length of the Western Transverse ranges is defined by d₁ and d₉ and their width along traverse 4 by d₇. The length of the eastern margin of the Western Transverse ranges, along the San Andreas fault in the bend region, is d₈. Also shown is a unit vector n₀, which is orthogonal to the line linking the northern ends of traverses 1 and 2.](https://academic.oup.com/gji/article-abstract/119/1/297/568506)
can be expressed as linear vector equations.

Sum of relative velocities in traverse 1:

\[ \sum_{j=1}^{3} v_{1j} = v_{ab1}. \]  \hspace{1cm} (17)

Sum of relative velocities in traverse 2:

\[ \sum_{j=1}^{2} v_{2j} + (d_1 + d_2) \omega_1 n_1 + d_2 \omega_2 n_2 = v_{ab2}. \]  \hspace{1cm} (18)

Sum of relative velocities in traverse 3:

\[ \sum_{j=1}^{4} v_{3j} + d_3 \omega_3 n_3 = v_{ab3}. \]  \hspace{1cm} (19)

Tying together traverses 1 and 2 along tie-line 4:

\[ v_{11} - v_{43} - v_{44} = d_4 \omega_1 n_4. \]  \hspace{1cm} (20)

Tying together traverses 2 and 3 along tie-line 4:

\[ \frac{d_4 v_{43}}{d_5} + v_{44} + d_2 \omega_2 n_2 = d_2 \omega_1 n_1. \]  \hspace{1cm} (21)

Tying together the ends of traverses 1 and 2:

\[ v_{ab2} = v_{ab1} + d_4 \omega_{ab} n_4. \]  \hspace{1cm} (22)

Tying together the ends of traverses 2 and 3:

\[ v_{ab3} = v_{ab2}. \]  \hspace{1cm} (23)

Linking the Mojave rotation rate with the Western

Figure 9. Detailed views of network of traverses crossing the San Andreas fault system and shown in Fig. 7(b). The relative horizontal velocities \( (v_i) \) along each traverse are defined, where the subscripts \( ij \) refer to the \( j \)th relative velocity in the \( i \)th traverse. Each relative velocity either represents slip on a fault or distributed shear across a region. Unit vectors which are orthogonal to traverse segments are also defined \((n_i - n_j)\). The Western Transverse ranges and Mojave regions are rotating about a vertical axis at rates \( \omega_1 \) and \( \omega_2 \), respectively (positive clockwise). (a) Definition of vectors in traverses 1, 2 and 3. (b) Definition of vectors in traverse 4, which links traverses 1, 2, and 3.
Transverse Range rotation rate:

\[ \omega_1 = \frac{v_{i2}}{d_{i1}} = \omega_2 \]

where \( v_{ij} \) is the horizontal velocity across the \( j \)th major fault or segment in the \( i \)th traverse (\( v \) is the horizontal slip rate) defined in Fig. 9; \( n_{i-4} \) are unit vectors orthogonal to traverse segments and \( n_{i0} \) is a unit vector orthogonal to the line linking the North American ends of traverses 1 and 2, defined in Figs 8 and 9; \( \omega_1 \) and \( \omega_2 \) are rates of rotation about a vertical axis of the intervening fault blocks in the Western Transverse ranges and Mojave region respectively, relative to the Pacific plate (Fig. 9); \( d_{i1} - d_{i2} \) are dimensions of the network, shown in Fig. 8. \( \mathbf{V}_{ab} \) is the corrected velocity of the North American plate relative to the Pacific plate at the North American end of the \( i \)th traverse, and \( \omega_{ab} \) is the local rotation about a vertical axis of the North American plate relative to the Pacific plate. Note that traverses 2 and 3 have the same North American endpoint (\( \mathbf{V}_{a2} = \mathbf{V}_{a3} \)). Traverse 4 refers to the tie-line linking traverses 1, 2 and 3 (Fig. 9).

Equations (17)-(23) can be re-expressed as linear scalar equations (two components for each vector equation). This yields 13 linear scalar equations in all, involving 17 unknowns. This set of simultaneous equations has no unique solution, because there are more unknowns \((N = 17)\) than equations \((M = 13)\). However, many of these unknowns can be estimated from field observations or sea-floor magnetic anomalies. These estimates and their associated uncertainties (standard deviation) can be expressed as linear equations themselves:

\[ \mathbf{V}_{ab} = \mathbf{V}_{abi \text{ (estimated)}} \pm \epsilon \]

where \( \epsilon \) is the uncertainty (standard deviation) associated with any estimate. Thus, we can add to eqs (15)-(24) as many additional equations as we have estimates. If the number of estimates is greater than 4, then we will have an overdetermined set of linear simultaneous equations. We can therefore use eqs (9)-(14) to find a weighted least-squares solution. The solution will depend both on our estimated parameters and the uncertainties of these estimates. We can therefore investigate the effect of changes in these estimates on the solution.

### Estimates

All the parameters, except the rotation rates about a vertical axis in the Western Transverse ranges \((\omega_1)\) and Mojave region \((\omega_2)\), can be estimated with a fair degree of confidence (Tables 1 and 2; Weldon & Sieh 1985; Eberhart-Phillips et al. 1990; Wallace 1990; McGill & Sieh 1993; Donnellan et al. 1993; Feigl et al. 1993). The least certain of the estimates is probably the dextral strike-slip rate on the San Andreas fault in the 'bend' region, defined by \( v_{21} \). An estimate of this depends upon whether it is assumed that all the dextral strike-slip on the San Jacinto fault is transferred to the San Andreas fault where these two faults meet (Eberhart-Phillips, Lisowski & Zoback 1990). Weldon & Sieh (1985) have determined the strike-slip rate for the San Andreas fault SSE of this junction. I therefore investigate two sets of solutions. In the first set of solutions (solutions A-E in Table 1), the \textit{a priori} estimate for the dextral strike-slip rate on the San Andreas fault \((v_{21})\), where it is crossed by traverse 2, is the same as that determined by Weldon & Sieh (1985) further SSE (circa 24 mm yr\(^{-1}\)). In the second set of solutions (solutions F-J in Table 2), this estimate is Weldon & Sieh’s value plus the dextral strike-slip rate on the San Jacinto fault, or circa 34 mm yr\(^{-1}\).

### Table 1. Solutions for horizontal velocity model for part of the San Andreas fault system, California, relative to the Pacific plate. See Fig. 9 for the definition of the velocities.

| Rel Plate motion: | Velocities (mm/yr) | \( M^o \) | Estimate(1) | A | B | C | D | E |
|------------------|-------------------|---------|-------------|---|---|---|---|---|
| Traverse 1       | \( \mathbf{V}_{ab1} \) | 144     | 372±2       | 372±1 | 34±3 | 39±3 | 37±1 | 37±1 |
| Traverse 2       | \( \mathbf{V}_{ab2} \) | 144     | 372±2       | 36±1 | 34±3 | 39±3 | 37±1 | 37±1 |
| Traverse 3       | \( \mathbf{V}_{ab3} \) | 144     | 372±2       | 36±1 | 34±3 | 39±3 | 37±1 | 37±1 |
| San Andreas F.:  |                   |         |             |     |     |     |     |     |
| N. of Bend (D)   | \( v_{11} \) | 140     | 35±5        | 32±2 | 33±6 | 30±5 | 31±3 | 31±3 |
| Bend region (D)  | \( v_{13} \) | 230     | 5±5         | 3±1  | 3±2  | 3±2  | 3±1  | 3±1  |
| S of Bend (D)    | \( v_{33} \) | 130     | 5±5         | 5±2  | 1±5  | 9±4  | 6±2  | 5±2  |
| Simple shear:    |                   |         |             |     |     |     |     |     |
| N. of Bend (D)   | \( v_{11} \) | 140     | 5±5         | 5±2  | 1±5  | 9±4  | 6±2  | 5±2  |
| S. of Bend       | \( v_{33} \) | 130     | 5±5         | 5±2  | 1±5  | 9±4  | 6±2  | 5±2  |
| San Jacinto F. (D)| \( v_{32} \) | 130     | 10±3        | 10±3 | 10±7 | 9±5  | 10±3 | 10±3 |
| Garlock F. (S)   | \( v_{22} \) | 240     | 7±3         | 6±2  | 13±2 | 4±3  | 3±2  | 6±2  |

### Rotation of Mojave

| Rotation of Mojave | \( \omega_{2} \) (Ma) | - | -0.7±0.5° | -2.9±0.7° | -1.5±0.4° | -0.8±0.6° |
|-------------------|---------------------|---|-----------|-----------|-----------|-----------|
| Traverse Range:   | \( \omega_{1} \) (Ma) | \( v_{41} \) | 10±5 | 10±5 | 10±5 | 10±5 | 10±5 |
| \( \omega_{2} \) (Ma) | \( v_{42} \) | 205 | 3±3 | 2±2 | 0±4 | 0±3 | -1±2 | 3±3 |
| \( \omega_{3} \) (Ma) | \( v_{43} \) | 205 | 0±5 | 10±2 | 0±3 | 2±3 | 2±3 | 2±3 |
| \( \omega_{4} \) (Ma) | \( v_{44} \) | 100 | 0±5 | 3±2 | 0±4 | 6±3 | 4±2 | 3±2 |

\[ \text{Fit} = \frac{N-M}{2} \]

Notes: (1) Estimates based on Weldon & Sieh (1985), Eberhart-Phillips et al. (1990), Wallace (1990), McGill & Sieh (1993), Donnellan et al. (1993), Feigl et al. (1993). (2) In solutions B-D the parameters in brackets have been fixed; solution J is based on the velocity model without equation (24) (see text).
Table 2. Further solutions for horizontal velocity model for part of the San Andreas fault system, California, relative to the Pacific plate. See Fig. 9 for the definition of the velocities.

NUVEL-1 (De Mets et al. 1990) relative plate motion (North American plate relative to Pacific), corrected for deformation in the Basin and Range (Argus & Gordon 1991): 37±2 mm/yr @ 144°.

| Velocities (mm/yr) | Az(°) | Estimate(1) | F | G | H | I | J |
|--------------------|-------|-------------|----|---|---|---|---|
| Traverse 1:        |       |             |    |   |   |   |   |
| V_ab1              | 144   | 37±2        | 37±1|35±3|39±3|38±1|37±1|
| Traverse 2:        |       |             |    |   |   |   |   |
| V_ab2              | 144   | 37±2        | 37±1|34±3|39±2|38±1|37±1|
| Traverse 3:        |       |             |    |   |   |   |   |
| V_ab3              | 144   | 37±2        | 37±1|35±3|39±3|38±1|37±1|
| San Andreas F.:    |       |             |    |   |   |   |   |
| N. of Bend (D)     |       |             |    |   |   |   |   |
| v12                | 140   | 35±5        | 33±2|34±5|30±5|32±3|33±2|
| v13                | 230   | 52±5        | 31±1|32±2|32±2|31±1|31±1|
| bend region (D)    |       |             |    |   |   |   |   |
| v21                | 115   | 34±5        | 33±2|37±4|24±4|29±2|33±2|
| v23                | 130   | 25±5        | 26±2|27±7|23±6|25±3|26±2|
| Simple shear:      |       |             |    |   |   |   |   |
| N. of Bend (D)     |       |             |    |   |   |   |   |
| v11                | 140   | 52±5        | 41±1|1±4|9±4|6±2|4±2|
| S. of Bend (D)     |       |             |    |   |   |   |   |
| v31                | 130   | 52±3        | 5±1|2±4|3±3|3±2|3±1|
| San Jacinto F. (D) |       |             |    |   |   |   |   |
| v32                | 130   | 10±3        | 10±2|11±5|9±6|10±3|10±2|
| Garlock F. (S)     |       |             |    |   |   |   |   |
| v22                | 240   | 7±3         | 9±2|14±2|4±3|3±2|9±2|
| Rotation of Mojave  |       |             |    |   |   |   |   |
| Transverse Ranges: |       |             |    |   |   |   |   |
| ω2 (Ma)            |       |             |    |   |   |   |   |
| v41                | 205   | 3±3         | 2±1|0±3|0±3|1±2|2±2|
| v42                | 205   | 10±5        | 8±2|0±3|22±3|(15)|8±2|
| v43                | 090   | 0±2         | 2±1|0±3|2±3|3±2|2±1|
| v44                | 000   | 0±2         | 2±1|0±3|6±3|3±2|2±1|
| ω2 (Ma)            |       |             |    |   |   |   |   |
| Fit = N.M/σ²       | 0.19  | 2.11        | 1.81|0.47|0.2|

Notes: (1) Estimates based on Weldon & Sieh (1985), Eberhart-Phillips et al. (1990), Wallace (1990), McGill & Sieh (1993), Donnellan et al. (1993); Feigl et al. (1993). (2) In solutions G and H the parameters in brackets have been fixed; solution J is based on the velocity model without equation (24) (see text).

I further investigate the kinematics of this part of California by varying some of the other initial estimates. In particular, I examine the relation between tectonic rotations both in the Mojave and Western Transverse ranges and the kinematics elsewhere in the deforming zone by fixing ω1 and ω2 to various a priori values.

Solutions A and F

If all available estimates of rates of deformation, excluding tectonic rotation in the Mojave and Western Transverse ranges, are included in the model, the total number of equations is 28, and matrix A in eq. (9) will be a 28×17 matrix, and the overdeterminacy (N−M) is 11. The weighted least-squares solutions (solutions A and F, Tables 1 and 2) predict: (i) the magnitude of the angular velocity about a vertical axis of the faults in the Western Transverse ranges is significantly greater than that for the Mojave region; (ii) the faults in the Western Transverse ranges are rotating clockwise about a vertical axis at 3.1°±0.4° Ma⁻¹ (solution A) or 2.7°±0.3° (solution F); (iii) the Mojave region is undergoing negligible anticlockwise rotation about a vertical axis at 0.7°±0.5° Ma⁻¹ (solution A) or 0.2°±0.4° (solution F); (iv) the dextral strike-slip rate on the San Andreas fault in the 'bend' region is 29–33 mm yr⁻¹. Solution F has a lower fit (see eq. 15) than solution A, with fits of 0.19° and 0.37° respectively.

In solutions E and J, eq. (24) which links the rotation rate about a vertical axis in the Western Transverse ranges with that in the Mojave region, is not included in the model. This is perhaps the most arbitrary of all the constraints. However, removing it has little effect on the solutions. Solutions E and J are essentially the same as solutions A and F.

The relative velocities derived in the models can easily be summed to determine a horizontal velocity field for this part of California. Fig. 10 shows a velocity field for solution F, relative to the Pacific plate (see Table 3).

Forced solutions

In solutions B, C, G and H, the rotation rates about a vertical axis in the Western Transverse ranges are investigated by fixing them to certain a priori values. The fits for these solutions can then be compared with those for

Figure 10. Horizontal velocity field for part of the San Andreas fault zone, relative to the Pacific Plate, based on Solution F (see Tables 2 and 3).
solutions A and F, where these rotation rates are left unconstrained in the model. In solutions B and G, the rotation rate in the Western Transverse ranges, and also the Mojave region, are constrained to be zero. It is clear that this has a marked effect on the fit of these solutions, which are up to 11 times worse than the fits for the solutions where these rotation rates were left unconstrained (solutions A and F). As the predicted rotation rate about a vertical axis in the Mojave region is nearly zero in solutions A and F, the poor fits for solutions B and C strongly suggest that the fault kinematics of this part of California require the Western Transverse ranges to rotate clockwise.

If the rotation rate for the Western Transverse ranges is constrained to be the average rate, determined from palaeomagnetic data (circa 6° Ma⁻¹ clockwise), then the solution fits are also poor (solutions C and H), with fits up to 10 times worse than those for solutions A and F. These solutions predict a high rate of internal shortening in the Western Transverse ranges (circa 22 mm yr⁻¹), high rates of deformation on their northern margin (u₂₆ and u₁₄), and high rates of dextral shear (circa 9 mm yr⁻¹) in the region south-west of the northern segment of the San Andreas fault. Also, solutions C and H predict a relatively low dextral strike-slip rate on the San Andreas fault itself in the 'bend' region of circa 23 mm yr⁻¹.

Solutions A, F, E and J predict rates of internal shortening in the Western Transverse ranges which are either in the middle or at the low end of the estimated range (10 ± 5 mm yr⁻¹). In solutions D and I, the internal rate of shortening here is fixed at the high end or 15 mm yr⁻¹. These solutions predict a slightly higher clockwise rotation rate about a vertical axis, compared to solutions A and F, of 3.9° ± 0.3° Ma⁻¹ (solution D) and 3.8° ± 0.3° Ma⁻¹ (solution I). Also, the Mojave region is predicted to be rotating anticlockwise at circa 1.5° Ma⁻¹. The fits for these solutions are only between 1.3 and 2.5 times worse than those for solutions A and F.

Long-term tectonic rotations

Palaeomagnetic data suggest an average rotation rate about a vertical axis in the Western Transverse ranges of circa 6° Ma⁻¹ over the last 15 Ma, and negligible anticlockwise rotation in the Mojave since 18 Ma (Luyendyk et al. 1985; Luyendyk 1989, 1991). The negligible rotation rate of the Mojave region suggested by the least-squares solutions (Solution A and F) is therefore consistent with the palaeomagnetic data, and also supports Jackson & Molnar's (1990) model of rotation in the Western Transverse ranges, which assumes that the San Andreas fault in the 'bend region' is not actively rotating, though it may have rotated anticlockwise in the past (Garfunkel 1974). However, the predicted rotation rates in the Western Transverse ranges for both solutions A and F are lower than the long-term rate, though they have the same sense.

There is no reason why the instantaneous rotation rate about a vertical axis in the Western Transverse ranges, on a time-scale of tens to hundreds of thousands of years, should be the same as the long-term average rate over millions of years, determined from palaeomagnetic studies. Lamb (1994) has analysed the Western Transverse ranges in terms of highly elongate rigid blocks which are 'floating' on a ductile lower part of the lithosphere. In this model, the rotation rate is strongly dependent on the orientation of the elongate blocks (Lamb 1987). Given the long-term rotation history, the model predicts that even if the overall ductile flow at depth has remained constant, the rotation rate of the overlying elongate blocks in the Western Transverse ranges has decreased to a present-day rate of circa 3° Ma⁻¹, in good agreement with solutions A and F.

Liddicoat (1992) has palaeomagnetically determined large clockwise tectonic rotations about a vertical axis in the Western Transverse ranges, in the western Ventura basin. Rotations of circa 20° have been measured in rocks which may be much younger than 3 Ma, though there is some uncertainty in the age. Taken at face value, these results suggest very young and rapid tectonic rotations (>7° Ma⁻¹) which are much faster than those in solutions A and F. However, the rate determined in solutions A and F is an average rate that applies to the entire length of the Western Transverse ranges, extending 250 km in an E-W direction. It is possible that the tectonic rotations measured in the Ventura basin are only local and do not apply to the whole Western Transverse ranges. This is also suggested by the fact that the high strain rates in this part of the Ventura basin region, observed over the last few years, are not typical of the rest of the Western Transverse ranges (Donnellan et al. 1993; Feigl et al. 1993).

Northern margin of the Western Transverse ranges

Deformation along the northern margin of the Western Transverse ranges in the vicinity of the Santa Ynez and Big Pine faults, is described by the relative velocities v₁₃ and v₁₄ (Fig. 9). A combination of these two orthogonal velocity vectors can describe relative motion in this region with any orientation, though the magnitude of each vector is estimated a priori to be low (<2 mm yr⁻¹).

All solutions predict positive rates for v₁₃ and v₁₄, implying components of E-W dextral shear and N-S extension between the northern margin of the Western Transverse ranges and the region further north. These predictions are in conflict with the available evidence for deformation in this region, which suggests components of N-S compression and E-W sinistral shear (Donnellan et al. 1993).
It is important to realise that the components of relative motion on the northern margin of the Western Transverse ranges \( (v_{43} \text{ and } v_{44}) \), predicted by solutions A to J, are really a consequence of the assumption that the region north of the Western Transverse ranges and west of the San Andreas fault is undergoing dextral simple shear parallel to the trend of the northern segment of the San Andreas fault. Though this assumption is probably reasonable in the vicinity of traverse 1 (Ekstrom et al. 1992), it may not be true slightly further south and immediately north of the Western Transverse ranges where earthquake fault-plane solutions suggest off-shore shortening in a direction roughly orthogonal to the trend of the northern segment of the San Andreas fault (Wallace 1990). In this case, the motion of the coastal and offshore part of the region immediately north of the Western Transverse ranges is in a SW direction, relative to the Pacific plate (see Fig. 10).

If all the orthogonal inland shortening along traverse 1 (circa 3 mm yr\(^{-1}\)), is taking place offshore and along the coast further south, then this SW-directed motion must be subtracted from the predicted value of circa 3 mm yr\(^{-1}\) of relative motion for solutions A and F in a NE direction \( (v_{23} \sim v_{44}) \) across the northern margin of the Western Transverse ranges. In other words, the true relative motion here, taking into account coastal or offshore shortening, is predicted to be very small and consistent with the low observed rates of active deformation here. It is interesting to note that solutions with high rates of clockwise rotation about a vertical axis of the Western Transverse ranges (solutions C and H), which also have poor fits, predict unrealistic E–W dextral shear and N–S extension in this region, even after a correction for offshore shortening is made.

The above discussion suggests that the orthogonal component of shortening (circa 3 mm yr\(^{-1}\)) in the northern sector of the San Andreas fault, switches from being accommodated inland and NE of the San Andreas fault in the north (Coalinga area), to offshore and SW of the San Andreas fault further south (Santa Maria valley area). If this is the case, then the northern segment of the San Andreas fault is rotating about a vertical axis clockwise at circa 1.5° Ma\(^{-1}\) into a direction more nearly parallel to the relative plate motion vector (Fig. 11). If this rotation has only been taking place since 4 Ma, when the orthogonal component of the relative plate motion vector increased (Harbert 1991), then the total finite rotation may be less than 6° and difficult to detect.

**CONCLUSIONS**

The analysis in this paper has been solely concerned with the kinematics of active deformation in the brittle crust. An attempt has been made to derive the velocity field for a wide zone of active deformation by searching for a self-consistent pattern of faulting in the brittle crust. However, the solution is only as good as the assumptions that went into the model. If the model predicts, within error, observations which were not estimated a priori, then these assumptions may be realistic. For instance, despite the simplifying assumptions used, the model predicts many features of the deformation in California.
The model can be used to test the available estimates of the deformation. By predicting a solution which is consistent with all the data, it is also a method of refining the errors in the estimates for any particular parameter. If the predicted errors in the solution are markedly less than the errors in the estimated parameters, and the agreement with additional observations, not used in the model, are good, one can be more confident that any additional information will not substantially alter the picture.

The description of a zone of deformation by a velocity field does not necessarily imply that the deformation is continuous. However, at large length-scales, small-scale faulting may be better characterized by a continuum description. Rapid slip on large faults cannot be described in this way. Thus, the velocity field of a zone of deformation may have discontinuities, across which there are abrupt changes in the velocities. Such discontinuities are particularly evident in California, where the slip rate on the San Andreas fault is a large fraction of the magnitude of relative plate motion. Haines’s (1982) method, using a smoothed distribution of the seismic-moment tensor, will tend to smooth out and ‘spread’ the deformation so that it conforms with a continuum, even if there are major discontinuities. This smearing of the velocity field will depend on the size of domain in which the seismic moment tensors are summed. Thus, Haines’s (1982) method, when indiscriminately used, is in danger of disguising real and significant discontinuities in the deformation of the brittle crust. However, the simple method described in this paper, though capable of describing continuums, will tend to emphasize the discontinuities. In addition, it can deal specifically with the rotation of fault blocks. It is therefore a powerful way of investigating how motion on individual faults effects the overall pattern of deformation.

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