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The solid-body tilt of deformed paleohorizontal planes: application to an Archean transpression zone, southern Canadian Shield

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Abstract—The solid-body tilt (SBT) of deformed paleohorizontal planes is equal to the angle between their unstrained top direction and an upward-pointing vertical axis. Kinematic models of important tectonic processes such as transpression and ductile thrusting have specific SBT requirements. These models can be tested, in ancient and modern orogens, by determining the actual SBT of strata, early magmatic sills and late horizontal intrusions with relict primary layering.

Classical transpression models predict that, except for sheath structures, all folds in bedding be upright and horizontal, i.e. that SBT values be close to zero within the hinge zones. Moreover, all passive folding is accomplished by amplification of asperities and caused by the pure-shear component of the transpression tensor.

Nine SBT analyses were made in the Archean crust of northwest Ontario to test the transpression hypothesis north of the Quetico–Wabigoon boundary, western Superior Province, Central Canada. The nine SBT values do not discredit the hypothesis that the Quetico Fault marks an Archean transpression zone which included the Wabigoon rocks adjacent to the Quetico metasediments.

INTRODUCTION

THE final inclination of originally horizontal surfaces (OHS, e.g. bedding planes, contacts of sills, unconformities) within ductilely deformed rock masses depends on the state of strain as well as the solid-body tilt (SBT), a component of the solid-body rotation (Ramsey & Huber 1983, p. 170). SBT values of ≤90° are equal to the dip of OHS after removal of the ductile strain (Ramsay 1969, Schwerdtner 1976, 1985). Large tilt angles can be found within tight upright folds and ductile fault zones, which may be classified according to their SBT patterns.

Solid-body rotations are caused by the combined effects of spin and shear-induced vorticity within ductile rocks (Means *et al.* 1980). The proportionate contribution of the two mechanisms to the local SBT (Lister & Williams 1983) cannot be determined, however, without knowing the actual path of tectonic deformation.

Kinematic models have been formulated recently for a variety of tectonic scenarios, and most models have particular SBT requirements. For example, no SBT occurs in ideal transpression zones (Sanderson & Marchini 1984), and the direction of SBT is perpendicular to the strike of inclined ductile-thrust zones (De Paor 1987).

Unfortunately, the mechanical anisotropy of deforming rock complexes and their analogue models leads to kinematic perturbations such as multi-order buckle folding or boudinage, which hamper the practical use of SBT patterns by the structural geologist. For example, the horizontal strata in transpression zones are prone to upright buckling (Harland 1971, fig. 3), which creates periodic strain perturbations and large SBT angles in the fold limbs. Passive folding, on the other hand, is governed strictly by the transpression tensor (Sanderson & Marchini 1984, equations 1 and 8), and therefore proceeds without SBT. More complicated models are available for oblique-collision orogens at continental margins (Ellis & Watkinson 1987, 1988, Girard *et al.* 1988), which again imply specific SBT conditions. These and other models can therefore be tested, in passively folded rocks of ancient and modern orogens, by determining the pattern of natural SBT. Depending on when a family of OHS was created in the geologic history of a rock mass, the SBT relates to the total tectonic deformation or merely a finite increment (Schwerdtner & Gapais 1983).

CALCULATION OF SOLID-BODY TILT (SBT)

In the strained state, paleovertical markers (e.g. *Scolithos* tubes) and top direction vectors are generally oblique to the OHS (Schwerdtner 1978, De Paor 1987). Strain removal re-establishes the perpendicularity between top direction vectors and OHS, which generally remain inclined or even overturned. The SBT angle (ρ) is therefore equal to the departure of unstrained top directions from an upward-pointing vertical reference axis (Fig. 1). The precision of SBT angles depends on the accuracy of compass measurements as well as errors in the strain magnitude ratios (Robin & Torrace 1987).

Let $\lambda_1 > \lambda_2 > \lambda_3$ be the principal quadratic elongations (Nadal 1950, Chap. 12). The orientation of the unit vector ($\mathbf{n}''$) normal to a deformed OHS may be specified,

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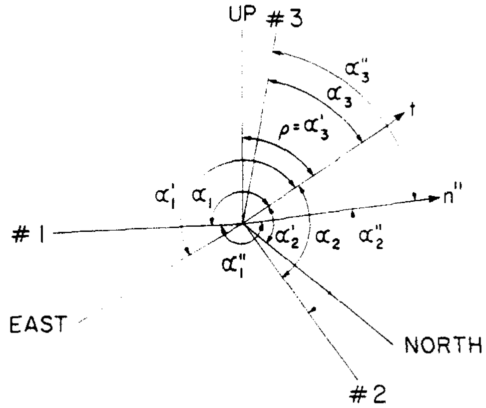

Fig. 1. Unit vector (n") normal to a deformed paleohorizontal plane (not shown) and unstrained top direction vector (t) in two reference frames. Geographic reference axes: east (No. 1'), north (No. 2') and upward (No. 3'). Nos 1, 2 and 3 are principal reference axes of the strain ellipsoid, corresponding to the directions of λ_1, λ_2, λ_3, respectively. Strained top direction vector not shown.

with respect to the principal axes of strain (Nos 1, 2, 3, Fig. 1), by the squared direction cosines a_1'', a_2'', a_3'' (Nye 1957). The orientation of the unstrained top direction, again with respect to the principal axes of strain, may be specified by the squared direction cosines a_1, a_2, a_3 (Nadai 1950, equations 12–47, Jaeger 1962, p. 36), whereby

$$\begin{aligned} a_1 &= \cos^2\alpha_1 = \lambda_1 a_1''/(\lambda_1 a_1'' + \lambda_2 a_2'' + \lambda_3 a_3'') \\ a_2 &= \cos^2\alpha_2 = \lambda_2 a_2''/(\lambda_1 a_1'' + \lambda_2 a_2'' + \lambda_3 a_3'') \\ a_3 &= \cos^2\alpha_3 = \lambda_3 a_3''/(\lambda_1 a_1'' + \lambda_2 a_2'' + \lambda_3 a_3''). \end{aligned} \quad (1)$$

Equations (1) perform the task of strain removal, but do not put the top-direction vector into a geographic reference frame. Consider a right-hand frame in which axis 1' points east, axis 2' points north and axis 3' upward (Fig. 1). The direction cosines of the angles α_1, α_2, α_3 are three perpendicular components of a top-direction unit vector that can be transformed into the geographic reference frame. Let $(\mathbf{l}_{ij})$ be the set of nine direction cosines that relate the principal strain directions to the geographic axes. Accordingly,

$$(\mathbf{l}_{ij}) = \begin{bmatrix} l_{11} & l_{12} & l_{13} \\ l_{21} & l_{22} & l_{23} \\ l_{31} & l_{32} & l_{33} \end{bmatrix}, \quad (2)$$

where i refers to a geographic axis and j to a principal axis of the strain ellipsoid. The components of the vector $(\mathbf{a}_j^{1/2})$ are transformed to (Nye 1957, Chap. 1)

$$a_i'^{1/2} = \mathbf{l}_{ij} a_j^{1/2}, \quad (3)$$

where $a_i^{1/2} = \cos\alpha_i$ and $a_i'^{1/2} = \cos\alpha_i'$.

Note that $\alpha_3' = \rho$, the SBT angle (Fig. 1), can be obtained from

$$\cos\alpha_3' = l_{31}\cos\alpha_1 + l_{32}\cos\alpha_2 + l_{33}\cos\alpha_3. \quad (4)$$

Finally, the geographic 'trend' (Ragan 1985, p. 366) of the unstrained top direction is

$$\beta = \tan^{-1}(\cos\alpha_1'/\cos\alpha_2'). \quad (5)$$

This means that $\pi + \beta$ is the azimuth of the horizontal tilt axis. Because of the low precision of field measurements and strain analyses, it seems reasonable to perform the transformation operation (equations 2–5) on the stereonet (Ragan 1985, Appendix B).

Sensitivity of mathematical equations

It is not always realized that angular changes of a few degrees can drastically alter the results of strain calculations. Equations (1)–(4) suggest that SBT estimates are affected in a similar fashion, and that the precision of ρ-values can be very low. This may be seen by considering a hypothetical example in which the deformation intensity is extremely high and the deformed top direction is nearly parallel to the horizontal axis No. 1. Strain removal (equation 1) produces a strong inclination of the top direction, which corresponds to a moderately large ρ-value (e.g. 45°). Had the deformed top direction been exactly parallel to axis No. 1, then the strain removal would not have rotated the top direction, leaving $\rho = 90°$. In other words, an error of ±2° in the angle between bedding and the 1–2 plane can double the ρ-value.

This problem was investigated quantitatively for the special case of plane strain ($\lambda_1 = 1/\lambda_3$, $\lambda_2 = 1$), whereby No. 1 was vertical and No. 2 normal to the unstrained top direction (t). Therefore, $\rho = \alpha_1$ (Fig. 1) and $a_2 = 0$ in equations (1) and (3). The variation of ρ with $a_1'' = a_3'' = \alpha''$ was determined for several values of λ_1 (Fig. 2). The α''/ρ curves show that, for plane strain with ≥400% maximum extension ($\lambda_1 \geq 25$), ρ changes by >50° between $\alpha'' = 90°$ (bedding strictly parallel to 1–2 plane) and $\alpha'' = 88°$. Small angles between bedding and planar strain fabrics are very common, which renders the present method useless at high deformation levels.

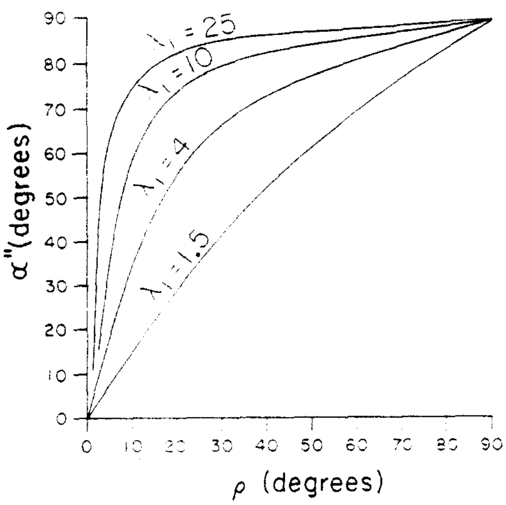

Fig. 2. Solid-body tilt angle (ρ) as a function of α'' (angle between n" and No. 3') for various values of λ_1.

TILT VALUES IN FOLDED STRATA

Except for overturned structures, the dip angle of horizontally folded OHS is equal to the SBT wherever the lines of strike and dip are parallel to principal axes of total strain. This is true at all localities within concentric buckle folds produced by longitudinal strain (Ramsay 1967, pp. 397–400). The equality also holds if horizontal folding is accomplished by mechanical failure of the hinges and rigid-body rotation of the straight limbs (Ramsay 1967, pp. 436–456).

The SBT of folded strata is part of their total deformation and therefore attainable along many kinematic paths. Like their total strain (Schwerdtner 1988), the SBT of strata need not be solely due to folding, especially within plunging upright structures. The plunge angle of such structures may represent a prefold SBT increment, which needs to be eliminated before using the SBT pattern as a guide to the fold mechanism. Alternatively, the plunge may be acquired after a folding, e.g. by rotation of rigid fault blocks. Because of this uncertainty, the following generalizations apply only to the passive folding of undeformed rocks and analogous model substances about horizontal axes (Schwerdtner 1988).

Tight passive folds can be generated by (1) homogeneous deformation of parallel surfaces with slight curvature (Ramberg 1964, Cobbold & Quinquis 1980), or (2) by heterogeneous deformation of perfectly planar markers (Ramsay 1967, pp. 423, 435 and 436). The initial curvature can be caused by low-amplitude buckle folding (Flinn 1962, Ramsay *et al.* 1983) or represent primary asperities in the OHS. A slight component of heterogeneous deformation (e.g. Ragan 1985, p. 246) can play the same role as an initial curvature.

Mechanism 1 leads to passive amplification of initial asperities, but depends greatly on the geometry of OHS at the start of passive folding. If the initial amplitude of asperities is infinitesimal and their finite deformation irrotational then the SBT is zero at all localities within tight folds. The SBT is finite but invariant if the deformation is clearly rotational, i.e. it includes large components of homogeneous simple shear. In either case, the intensity of fold-forming homogeneous strain must be very high to transform slight asperities into tight folds (Ramberg 1964, Cobbold & Quinquis 1980). Tight folding is apt to have been dominantly active if the strain intensity is low to moderately high within tight natural structures.

Mechanism 2 leads to folds with variable SBT values, especially if the total deformation varies greatly within a structure. Such heterogeneity involves large simple-shear components and associated SBT angles that vary greatly with position (e.g. Ramsay 1967, fig. 7-89). The SBT pattern of natural folds may therefore be used to discriminate between the two principal mechanisms of passive folding.

TEST OF TRANSPRESSION HYPOTHESIS IN THE ARCHEAN CRUST WEST OF LAKE SUPERIOR

In modern structural geology, 'progressive deformation' refers to states of total continuous deformation that are attained along *specified* kinematic paths (Flinn 1962, Ramsay 1967, pp. 326–332). Transpression (Harland 1971, Sanderson & Marchini 1984) is a progressive deformation which may be responsible for much of the large ductile strain accumulated in the Archean meta-

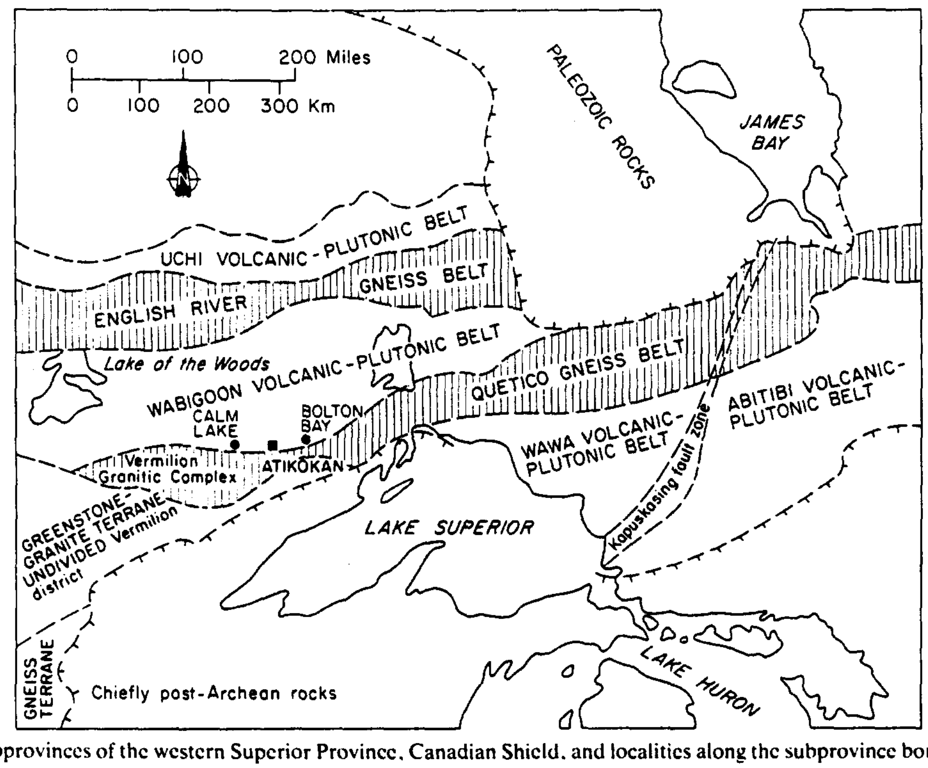

Fig. 3. Subprovinces of the western Superior Province, Canadian Shield, and localities along the subprovince boundaries.

volcanics, metasediments and migmatites near the faulted boundaries of the Quetico Subprovince (Fig. 3) west of Lake Superior (Bauer *et al.* 1986, Hudleston *et al.* 1986, 1988, Stott 1986, Borradaile *et al.* 1988, Sarvas 1988). The boundary regions of other subprovinces in the western Superior Province, Canadian Shield, may have experienced the same type of progressive deformation (Stott *et al.* 1987).

Near the northern boundary of the Quetico Subprovince (Fig. 3), upright F_1 folds are oblique to the boundary fault (Borradaile 1982, Borradaile & Schwerdtner 1984) and attributable to dextral ductile transpression (Borradaile *et al.* 1988, Sarvas 1988). Near the southern boundary of the subprovince, however, F_1 folds are generally recumbent (Sawyer 1983, Bauer 1985, Hudleston *et al.* 1986, Sawyer & Robin 1986), but the *subsequent* deformation is explicable by dextral transpression (Bauer *et al.* 1986, Hudleston *et al.* 1988). SBT values will be estimated in large fold structures just north of the Quetico Subprovince, using field data obtained by Borradaile in the southernmost Wabigoon Subprovince (Borradaile 1982, Borradaile & Schwerdtner 1984). These ENE-trending folds seem to be truncated by the late-stage Quetico Fault, which marks the northern boundary of the Quetico Subprovince as well as an inferred transpression zone (Borradaile personal communication, 1988). If the Quetico Fault was either created or reactivated at a late stage of dextral transpression (Borradaile *et al.* 1988, p. 1076) then it could have cut the oblique folds generated at an earlier stage of the same transpression. This possibility warrants an SBT-based test of the transpression model in the folded Wabigoon metavolcanics north of the Quetico Fault (Fig. 3).

The transpression model

Ideal transpression zones are narrow ductile domains between large rigid blocks (Harland 1971, Davies 1984, Sanderson & Marchini 1984). The progressive deformation in such transpression zones consists of two synchronous parts (i) a transcurrent simple shear parallel to the zone boundaries and (ii) a concordant pure shear, without longitudinal strain in the direction of simple shear (Fig. 4). The total deformation can be represented by the Cartesian tensor

$$(\mathbf{T}_{ij}) = \begin{bmatrix} 1 & \gamma/\sqrt{\lambda} & 0 \\ 0 & 1/\sqrt{\lambda} & 0 \\ 0 & 0 & \sqrt{\lambda} \end{bmatrix} \qquad (6)$$

obtained by combining a finite simple shear with a finite pure shear (Sanderson & Marchini 1984). This combination is not meant to simulate the actual deformation path of transpression, but γ is the total transcurrent unit shear and λ the vertical quadratic elongation of the total pure shear. Equation (6) implies that there is neither vertically-directed simple shear nor SBT (Fig. 4). The ductilely transpressed rocks slide upward without friction on the zone boundaries while perfect cohesion is

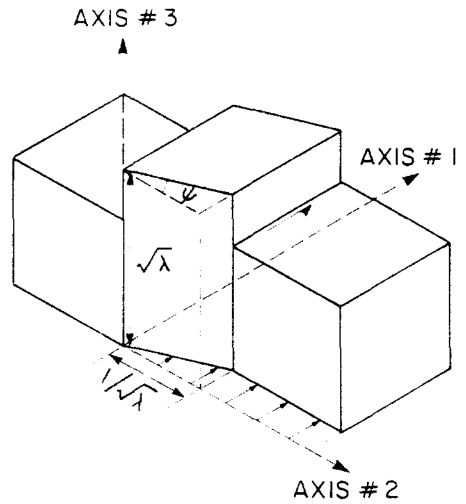

Fig. 4. Deformation of ideal transpression zones (redrawn from Sanderson & Marchini 1984, fig. 1). Axes 1 and 2 are non-principal, as opposed to those in Fig. 1. Arrows in basal plane represent component of transcurrent shear only.

retained along strike. Such extreme anisotropy in frictional behaviour could be justifiable on dynamic grounds, but may seem contradictory in a kinematic sense. If vertical slip on the zone boundaries is disallowed then the contact strain ellipsoids become discordant and can lose their oblateness (Sanderson & Marchini 1984). If propagated into transpression zones, this effect creates oblique stretching lineations and a total strain field with monoclinic symmetry:

$$(\mathbf{T}_{ij}) = \begin{bmatrix} 1 & \gamma(1 + \{\partial u_2/\partial x_2\}) & 0 \\ 0 & 1 + (\partial u_2/\partial x_2) & 0 \\ 0 & \partial u_3/\partial x_2 & \sqrt{\lambda} \end{bmatrix}. \qquad (7)$$

whereby γ is constant and $\lambda^{-1/2} \doteq 1 + (\partial u_2/\partial x_2)$. Furthermore, $\partial u_2/\partial x_2$ and $\partial u_3/\partial x_2$ (Cobbold 1977) are displacement gradients caused by vertical frictional drag along the zone boundaries. If the direction of transcurrent shear is slightly inclined then $T_{32} = \delta + (\partial u_3/\partial x_2)$, where δ is the component of vertical uniform shear.

The discussion shows that transpression zones need to be modelled as dynamical systems, and that available kinematic treatment (Davies 1984, Sanderson & Marchini 1984) may be insufficient. In particular, the degree of heterogeneity of the total deformation and the associated horizontal variation of vertical shear and longitudinal strain need to be evaluated under appropriate dynamic conditions, both for transpression at subduction zones (Harland 1971, Hudleston *et al.* 1988, Hoffman in press) and for transpression within continents (Tapponier *et al.* 1982).

Equation (6) predicts no spin in simple transpression zones, and a shear-induced finite rotation about the vertical axis. This implies that perfectly planar strata remain horizontal throughout progressive deformation, assuming that the transpression model is valid. Real strata are not perfectly coplanar and therefore respond to horizontal shortening by buckling or passive folding (Ramberg 1963, Ramsay 1967). Tight buckling about horizontal axes results in large SBT of sedimentary strata within natural transpression zones (Harland 1971, fig. 3). Tight passive folding, on the other hand, can be accomplished by homogeneous irrotational strain (Flinn 1962, Ramberg 1964). If the undeformed horizontal strata have slight geometric corrugations ('asperities') that resemble circular dome-and-basin structures, then transpression (equation 6) can amplify the domes and basins and distort their horizontal outlines by transcurrent simple shear without changing the upright character of the passive structures. Comparable folds are obtained if the initial asperities are oval or irregular in horizontal outline.

Depending on the magnitudes of transcurrent simple shear and vertical extension, the vertical axis coincides with the No. 1 or 2 axis of bulk strain (Fig. 1). If No. 2 is vertical, then the corrugated bedding becomes nearly cylindrical during progressive deformation (equation 6). This is because all tangents to bedding except those normal to axis No. 1 (Fig. 1) rotate passively toward the horizontal plane (Flinn 1962), in which the strain is non-coaxial. The amplitude of the cylindrical horizontal folds may therefore remain relatively small.

If No. 1 is vertical then the dome-and-basin corrugations are being amplified, but no passively rotating tangent to bedding can get past the vertical axis, i.e. there is no overturning of beds. This means that the axes of sheath folds generated in simple transpression zones will have very steep attitudes.

These are important constraints for sheath folding within ideal transpression zones (equation 6) as well as natural transpression regimes. Even appreciable dip-shear components within transpression zones may not succeed in producing sheath folds with subhorizontal axes and overturned hinge segments (Borradaile *et al.* 1988, fig. 3).

The original shape of upright asperities can be restored by removing the total strain (equation 6). This does not include the shear-induced solid-body rotation about the vertical reference axis (equation 6), but restores the boundaries of strata to a quasi-horizontal attitude.

Estimates of solid-body tilt (SBT)

The Bolton Bay Syncline occurs in the easternmost segment of the Atikokan greenstone belt (Fig. 3), just north of the Quetico Fault (Bau 1979, Borradaile & Schwerdtner 1984). In map view, two thirds of this subhorizontal fold are mafic metavolcanics, the remaining third is composed of metasediments, which are confined to the south limb (Fig. 5). Like many upright folds west of Lake Superior, the Bolton Bay Syncline seems to be a tight chevron structure containing suboblate vertical pillows and pebbles. Because the fold is virtually horizontal, its three-dimensional shape cannot be estimated by down-plunge projection.

A major lithologic boundary separating metavolcanics from metasediments occurs within the south limb and does not reappear in the north half of the structure

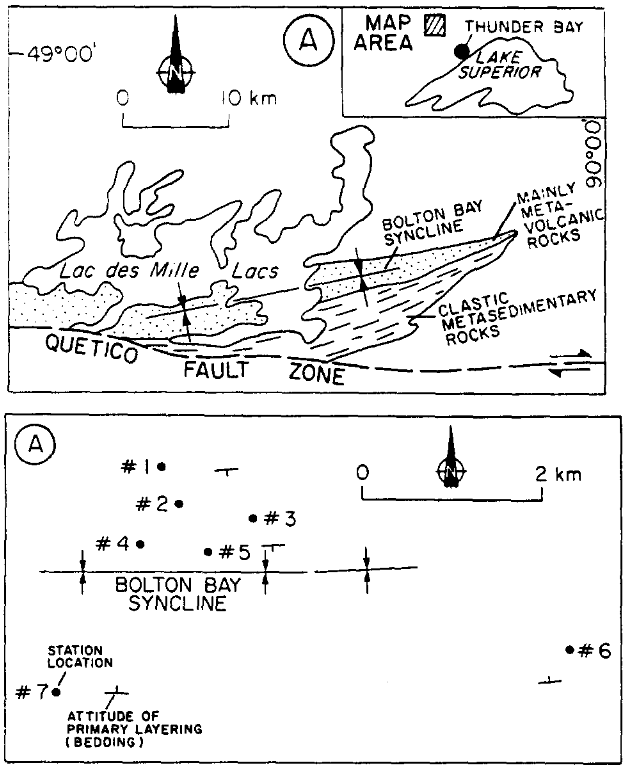

Fig. 5. Bolton Bay Syncline (a) and sites of strain analysis (b) (Borradaile & Schwerdtner 1984, Schwerdtner 1986).

(Fig. 5). The subsurficial terminus of the metasediments is unknown, but it may lie in the hinge zone of the syncline and possibly influence the siting of the hinges. The primary layering is rarely folded on a small scale, and seems to have been passive during the regional deformation. As in the metavolcanic and metasedimentary rocks of the Vermillion district of Minnesota (Fig. 3) (Hudleston 1976), lineation and schistosity are effectively vertical throughout the Bolton Bay Syncline (Kay 1967, Bau 1979).

The line of intersection between bedding and schistosity is effectively horizontal so that the ρ-values (SBT) can be estimated within the vertical cross-section (Schwerdtner 1985, fig. 8). As $\mathbf{n}''$ is normal to axis No. 2 (Fig. 1), $a_1'' = \cos^2\alpha''$, $a_2'' = 0$, $a_3'' = \sin^2\alpha''$ and equation (1) simplifies to

$$\begin{aligned}\cos^2\alpha_1 &= \cos^2\alpha = (1 + \lambda_3\tan^2\alpha''/\lambda_1)^{-1}\\ \cos^2\alpha_2 &= 0 \\ \cos^2\alpha_3 &= \sin^2\alpha = (1 + \lambda_1\cot^2\alpha''/\lambda_3)^{-1}.\end{aligned} \tag{8}$$

Similarly, equation (2) simplifies to

$$(\mathbf{l}_{ij}) = \begin{bmatrix} 0 & l_{12} & L_{13} \\ 0 & l_{22} & l_{23} \\ 1 & 0 & 0 \end{bmatrix} = \begin{bmatrix} 0 & \cos\phi & -\sin\varphi \\ 0 & \sin\phi & \cos\varphi \\ 1 & 0 & 0 \end{bmatrix}. \tag{9}$$

where $\phi = \alpha' - \alpha$ (Fig. 1).

Accordingly, equation (3) becomes

$$\begin{aligned}\cos\alpha_1' &= -\sin\phi\sin\alpha \\ \cos\alpha_2' &= \cos\phi\sin\alpha \\ \cos\rho = \cos\alpha_3' &= \cos\alpha.\end{aligned} \tag{10}$$

Finally using equations (8)–(10) one obtains:

$$\rho = \cos^{-1}\{(1 + \lambda_3 \tan^2 \alpha''/\lambda_1)^{-1/2}\}$$
$$\rho = \sin^{-1}\{(1 + \lambda_1 \cot^2 \alpha''/\lambda_3)^{-1/2}\}.$$

Table 1 lists values of ρ obtained at seven sites within the Bolton Bay Syncline by using Borradaile's structural data (Borradaile & Schwerdtner 1984) as well as earlier calculations (Schwerdtner 1985). Note that ρ values of $>50°$ are obtained at all sites, which is incompatible with the bulk deformation in ideal transpression zones (equation 6 and Fig. 4). If the Bolton Bay Syncline is a horizontal buckle fold, however, its high ρ-values can be explained by perturbation of a transpressive deformation. Indeed, the total strain ratios obtained in the syncline (Table 1) are too small for creating a tight upright fold (Schwerdtner 1985) (Table 1) by passive amplification of low-amplitude asperities. Other evidence such as the paucity of small-scale buckle folds point to passive folding, however.

Table 1. SBT values in Bolton Bay Syncline

Site No.	$\sqrt{\lambda_1}/\sqrt{\lambda_3}$ (minimum values)	(°)	Tilt direction
1	3.1	61	SE
2	2.0	74	SE
3	3.0	70	SE
4	2.6	75	SE
5	2.5	78	SE
6	1.3	65	NW
7	2.0	54	NW

Like the Bolton Bay Syncline, the Calm Lake folds (Fig. 3) are upright F_1 structures occurring just north of the Quetico Fault (Borradaile 1982, figs. 1–5). The folds are isoclinal, overturned chevron structures which have a hinge-plane schistosity or cleavage and steep non-axial stretching lineations. However, the state of total strain is highly oblate and the lineation very weak. It is not known to what degree the folding was passive, but the steep plunge of the fold hinges seem incompatible with transpression (equation 6).

Estimates of total rock strain are available at two localities, where the schistosity is subparallel to the overturned primary layering and the dip angle about 60° (Borradaile 1982, fig. 3). If the layering were strictly parallel to schistosity then the dip values would be equal to the supplement of ρ. In view of the moderately large axial ratios of the total-strain ellipsoid (2.85 and 4.00), the actual values of ρ cannot be much smaller than 110°, but seem too large even for *active* upright folding of horizontal strata in transpression zones.

Interpretation of SBT data

The angles of ρ estimated just north of the Quetico Fault (Fig. 5a) are incompatible with passive upright folding of horizontal strata in ideal transpression zones (Fig. 4 and equation 6). Either the transpression hypothesis is valid but the folding is predominantly active, or the transpression model (equation 6) is inadequate. Alternatively, the folded rock strata may have formed as recumbent structures and rotated subsequently into subvertical attitudes. This is particularly plausible for sideward-facing upright folds (Borradaile 1982, p. 185), which could have been rotated by adjacent granitoid plutons.

More definite conclusions could be reached by ascertaining that the upright folding was active or dominantly passive. This might be accomplished by means of the total-strain pattern as shown by Hudleston (1976) in the Wawa metavolcanics (Fig. 3). In addition, dynamic models of transpression zones are urgently needed which predict the patterns of total and progressive strain in passively-behaving rock assemblages.

SUMMARY AND CONCLUSIONS

The solid-body tilt (SBT) of deformed paleohorizontal planes is equal to the departure from verticality of the unstrained top directions. Kinematic models of important types of deformation regimes have specific SBT requirements, and these models can be tested by making SBT analyses in rocks that escaped severe straining.

Transpression is held to have affected the Archean rocks near subprovince boundaries west of Lake Superior (Stott 1986, Borradaile *et al.* 1988, Hudleston *et al.* 1988). Transpression models predict that SBT values be close to zero throughout upright passive folds, and that the SBT values be relatively small near the hinge lines of most folds.

Nine SBT analyses were made in Wabigoon metavolcanics and associated metasediments with minimal $\sqrt{\lambda_1/\lambda_3}$ of ≤ 4.0 to test the transpression hypothesis at the Quetico–Wabigoon boundary. The analytical sites are within upright to inclined structures that seem to be due largely to active folding. All SBT values are $>50°$ and possibly explicable by buckle perturbation of the transpressive deformation field. The steeply-plunging hinge lines of some folds do not accord with the classical transpression model, which may be inadequate for the present purpose.

No SBT values could be estimated in the Quetico metasediments, where strain data seem to be unavailable. These rocks were apparently subjected to dextral transpression (Borradaile *et al.* 1988), but the final geometry of passively folded beds is difficult to explain by transpression alone.

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