The use of balanced cross-sections in the calculation of orogenic contraction: A review

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The use of balanced cross-sections in the calculation of orogenic contraction: A review

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SUMMARY: Balanced section calculations assume that the section has been deformed by a plane strain. If the section is underlain by a décollement, it can be restored to its pre-deformational length by dividing the area of the section by its original stratigraphic thickness. The thickness can be measured in undisturbed foreland beds or reconstructed by unstraining deformed beds. The restoration of the balanced section does not depend on any particular mechanism of deformation or folding, but in the special case of flexural slip folding the section can be unstrained by straightening out the sinuous bed lengths of the folds. However, in nature the area of the section may have decreased by 15–45%, but the assumption of plane strain always leads to a minimum estimate of shortening. Balanced section calculations suggest that the margins of orogenic belts have contracted by 35–54%.

Balanced geological cross-sections are drawn with the assumption that the area of section has not changed during deformation. It is assumed that an original sequence of flat-lying beds has been folded and faulted to form the present geological cross-section. If the section contains flexural slip folds with their axes normal to the section, the beds will suffer no shortening or elongation along the axes. Hence, the deformation in the plane of the section will be a plane strain. The area of the section before strain 'balances' the area after strain. However, it is not always necessary to assume that the folds in the plane of the section are flexural slip folds. Balanced section calculations can be made on geological sections simply by assuming that in the plane of the section there has been no change in area (Chamberlin 1910; Goguel 1962).

Many of the ideas which have been incorporated into balanced section construction have been developed by geologists working in the oil and gas fields of Alberta (Douglas 1950; Hunt 1957; Carey 1962; Bally *et al.* 1966) prior to the formal description by Dahlstrom (1969*a*). In the Alberta sections, most of the folds appear to have been formed by flexural slip and the thickness of the beds measured normal to the bedding remains constant. If the area of the cross-section remains constant, the bed lengths measured around the folds at different structural levels in the section must remain constant (Dahlstrom, *op. cit.*). The sinuous bed lengths around the folds should always be checked when a geological cross-section is constructed and if the bed lengths in any section do not 'balance' the section cannot be correct (Dahlstrom, *op. cit.*). The bed lengths should also be balanced across normal, reverse, and thrust faults. Dahlstrom (*op. cit.*) has shown that, at one level in the section, orogenic contraction can be accommodated by thrust faults, and at another level by folds. But the bed lengths in each of the levels must balance, and if they do not there must be a convincing structural explanation. Hence, the term balanced cross-section can be used in two ways. It can either be applied to sections where it is assumed that the areas of the original and the strained sections are the same, or, in the case of flexural slip folding, where the bed lengths in the original and the strained sections are the same.

Balanced sections have one very useful property; they can be unstrained to restore the beds of the section back to their depositional position (Bally *et al.* 1966; Dahlstrom 1969*a*). Therefore, these sections can be used to carry out finite strain analyses on large sections of the Earth's crust. So far, the technique has only been used on gently folded rocks (Chamberlin 1919) or in geological cross-sections along the margins of orogenic belts (Chamberlin 1910; Bucher 1933; Laubscher 1962; Dennison & Woodward 1963; Bally *et al.* 1966; Dahlstrom 1970; Gwinn 1970).

Balanced section calculations

Balanced section calculations were first used to estimate the depth to the décollement underlying concentric folds (Chamberlin 1910, 1919; Bucher 1933; Goguel 1962; Laubscher 1962; Dahlstrom 1969*b*). Consider a series of beds of thickness t_0 in their depositional position which initially overlies a plane destined to become a thrust or décollement surface (Fig. 1). The initial section has an arbitary length l_0 but it is important in section balancing that the ends of the section are fixed at points where there is no interbed slip. These points can be chosen as the planes which will become axial surfaces of folds or planes normal to the regional dip in the undisturbed beds of the foreland (Dahlstrom 1969*a*). The bed AB of length l_0 (Fig. 1) is transformed by folding into the strained state $A'B'$ with axial surfaces $A'D$ and $B'C'$ at each end of the section. The final length of the section after folding is AO or l_1. During the folding,

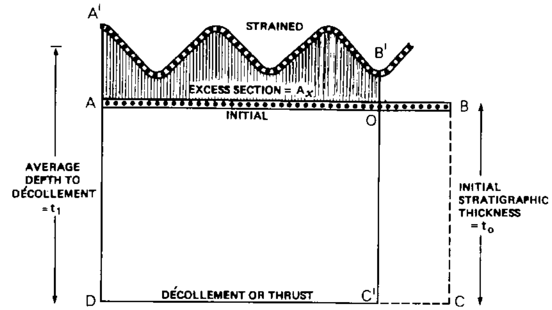

FIG. 1. Chamberlin's (1910, 1919) equal area calculation. The bed AB, originally at height BC above the décollement surface, is folded into a new position $A'B'$, $AB = l_0$, $AO = l_1$, $OB =$ shortening. During plane strain, $A'B'C'D = ABCD$, therefore excess section $A_x = OBCC'$. Initial depth to décollement $t_0 = A_x/OB$, or shortening $OB = A_x/t_0$. Within the area $A'B'C'D$ the rocks may be deformed by any style of folding or faulting.

the reference bed AB is uplifted to its strained position and the area A_x between the folded and the initial positions is described as the excess section (Gwinn 1970). The area of A_x can be measured with a planimeter or by Simpson's rule. The initial stratigraphic thickness t_0 is transformed to an average structural thickness t_1 (Fig. 1). If the deformation is a plane strain, the area of section $ABCD$ before strain will equal the area $A'B'C'D$ after strain. Because the area $AOC'D$ is common to both sections, A_x, the excess section, is equal to the area $OBCC'$. The latter is the product of the shortening and the initial depth to décollement (Fig. 1). If the folds have flexural slip geometry, the shortening can be estimated by unfolding the sinuous bed length around the folds. Hence the initial depth to décollement is A_x divided by the shortening (Chamberlin 1910, 1919; Goguel 1962; Dahlstrom 1969*b*). The initial height of the reference bed AB is usually estimated from the height of the same bed in the undisturbed beds of the foreland, although this assumes that there are no thickness or facies changes between the folded and undisturbed zones.

Chamberlin (1910) used a balanced section calculation to estimate that the depth to décollement in the Pennsylvania Appalachians lay between 9 and 52 km. He later estimated that the depth to décollement beneath the folds in the Colorado Rockies lay between 21 and 172 km (Chamberlin 1919). This latter result suggests that the surface folds die out in the Earth's mantle, clearly an absurd result. There are several possible sources of error in such a calculation. Firstly, the estimate of the initial height of the reference bed may be in error because of thickness or facies changes towards the undisturbed belt. Another possible error could arise from the assumption of flexural slip folding. The folds could have similar geometry in which case the unfolded bed length will not give an accurate estimate of total shortening. However, recent applications of the technique by Laubscher (1962) in the Jura and by Dahlstrom (1969*b*) in Alberta have given satisfactory estimates of depth to décollement.

Another source of error was described by Bucher (1933). Chamberlin (1910, 1919) divided his geological sections into small segments and carried out depth to décollement calculations in each segment. Bucher (*op. cit.*) suggested that a more accurate estimate could be made if the calculation was carried out over the whole area of the section. He re-estimated that the depth to décollement in the Appalachians was 21 km and, in addition, that the depth to décollement in the Jura was 852 m. It is not clear why Bucher's technique should give a better estimate. However, it is likely that the division of the section into segments may produce serious over- and under-estimates of the depth to décollement which cancel out over the whole section.

The logic of the Chamberlin technique can be reversed to calculate the amount of shortening if the depth to décollement is known (Dennison & Woodward 1963; Gwinn 1970). This is potentially more useful because the depth to décollement may be known from borehole data or from seismic sections (Bally *et al.* 1966; Dahlstrom 1970). Also, it is not necessary to assume flexural slip folding in the plane of the section. The beds beneath $A'B'$ (Fig. 1) can be deformed internally by concentric or similar folds, faults, imbricate thrusts, or solution transfer mechanisms. All that is necessary to calculate the shortening is to assume plane strain. The calculations have not been described in detail by Dennison & Woodward (1963) or by Gwinn (1970) but the method has been described by Elliott (1977). The original length of the deformed geological cross-section l_0 is given by

$$l_0 = (A_x/t_0) + l_1 \text{ (Elliott 1977, after Gwinn 1970).}$$

Gwinn (1970) used this calculation to estimate that the Valley and Ridge Province of the Appalachians had undergone 43% orogenic contraction (where the contraction is measured by the conventional engineering strain $\varepsilon = (l_1 - l_0)/l_0$).

The original length of a geological cross-section can also be estimated from the total area of the section (Dennison & Woodward 1963; Kiefer & Dennison 1972). Imagine a section of layered rocks of original length l_0 and original stratigraphic thickness t_0 (Fig. 2). The section becomes shortened by folding and thrusting to a final length l_1 and an average structural thickness of t_1. Irrespective of the style of folding, if the deformation is a plane strain,

$$\text{area of section} = l_0 t_0 = l_1 t_1$$

$$\therefore \quad l_0 = l_1 t_1 / t_0$$

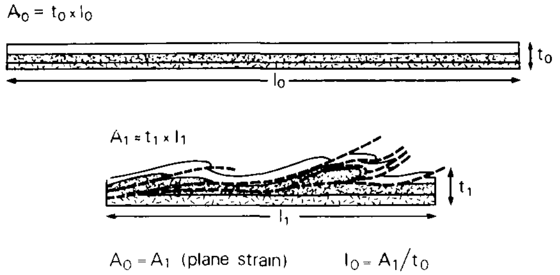

FIG. 2. Balanced section calculation using the total area of the section. A section of crust of original length and thickness l_0 and t_0 is contracted to l_1 with an average structural thickness of t_1. During plane strain $A_0 = A_1$; therefore $l_0 = A_1/t_0$. A_1 is measured more accurately with a planimeter.

This, of course, is simply a rearrangement of the Elliott equation above. The area of the excess section or the geological cross-section can be measured more accurately with a planimeter. Using the latter equation, Kiefer & Dennison (*op. cit.*) estimated that the Devonian of Alabama and Georgia had an average orogenic contraction of 35%.

I have applied this calculation to the rocks of the Etnedal nappes of the southern Norwegian Caledonides (Fig. 3). The original thickness of the nappe rocks (t_0) cannot be measured directly because the rocks of equivalent age in the foreland are much thinner and have a completely different stratigraphy. However, t_0 can be estimated by unstraining the distorted stratigraphic section using finite strain data (Cloos 1947; Borradaile & Johnson 1973; Tobisch *et al.* 1977; Hossack 1978). Calculations by these authors indicate that stratigraphic sections are frequently either halved or doubled in thickness during deformation. The stratigraphy of the Etnedal nappes has been measured carefully by R. P. Nickelsen (*pers. comm.*). The Eocambrian to Middle Ordovician sequence has a deformed thickness of 332 m averaged over several sections (Fig. 3). The rocks contain a slaty cleavage which is nearly parallel to bedding and deformed pebbles in conglomerates suggest that there was 20–25% ductile flattening normal to the cleavage (C. Peach, *pers. comm.*). The maximum 25% flattening value has been chosen because it leads to a minimum result in subsequent calculations. If all the beds of the section have undergone the same amount of flattening then the removal of the ductile strain gives an estimated original stratigraphic thickness (t_0) of 443 m (Fig. 3). The flattening strain has been measured in the most competent rocks of the section and hence this t_0 value must at least be a minimum estimate. The present cross-sectional area of the nappes is 4.15×10^7 m^2. Dividing this area by the estimated t_0 suggests that the original length of this section between Mellane and the thrust front (Fig. 3) was 94 km. The

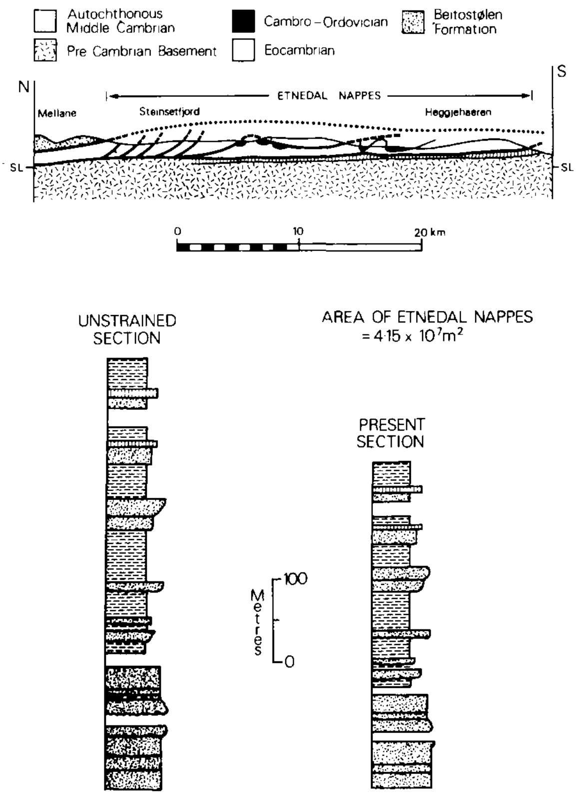

FIG. 3. Geological cross-section, with ×2 vertical exaggeration, through the Etnedal nappes between Mellane and the thrust front to the S. Beneath is the stratigraphy of one of the nappes in the Steinsetfjord district averaged over several localities (R. P. Nickelsen, *pers. comm.*). Removal of 25% cleavage-forming strain from the present day thickness of 332 m restores the beds to a pre-deformational thickness (t_0) of 443 m. The area of the Etnedal nappes from Steinsetfjord southwards to the thrust front is estimated to be 4.15×10^7 m^2.

present length of the section is 54 km, giving a contraction of 43%. It is fortunate that this calculation is insensitive to errors in the estimate of the original stratigraphic thickness (t_0). Dennison & Woodward (1963) showed in their calculations that variations of up to 365 m (1000 ft) in t_0 affected the calculated shortening by only 2–3%. Unfortunately, it is not possible to estimate from their figures what percentage 365 m represents in the total stratigraphic section. However, I have calculated from the Etnedal data that a 50% underestimate in the thickness t_0 would only reduce the original length of the nappes to 83 km.

The most sophisticated restorations of geological cross-sections have been made by geologists in the Alberta fold and thrust belt (Douglas 1950; Bally *et al.* 1966; Dahlstrom 1969*a*, 1970). In addition to sinuous bed length and area balancing calculations, they have also had to take into account the stratigraphic separation along numerous major and imbricate fault zones of the region. Thus, the total shortening is the sum of both the folding and thrusting, and the orogenic contraction in Alberta is estimated to have been 54% (Price & Mountjoy 1970). In addition, the restored sections of Bally *et al.* (*op. cit.*) indicate the area loss by erosion from the frontal toes and risers of the nappes.

Reduction of area in balanced cross-sections

Geiser (1978) criticized the assumption of plane strain in balanced section calculations. For instance, Ramsay & Wood (1973) suggested that slates may undergo a volume decrease of 10–20% during finite strain. Many of the rocks in the marginal parts of orogenic belts show evidence of solution transfer (von Plessman 1964) and this could also lead to a volume loss during finite strain. Therefore it is pertinent to consider the effect of volume loss on balanced section calculations. The processes by which volume may be lost are compaction accompanying lithification, tectonic compaction, and pressure solution. In addition, the area of the section can decrease because of elongation along the orogenic strike.

Compaction during lithification

During burial, wet sediments may lose up to 50% of their volume because of the removal of pore water (Wood 1974). This type of compaction has uniaxial symmetry and merely reduces the stratigraphic thickness normal to the bedding. Oertel (1970) estimated that lapilli tuffs in the Lake District had been compacted by 52% before they suffered tectonic finite strain. Sanderson (1976) estimated that undeformed chalk may undergo 32–44% compaction during lithification. However, lithification volume loss should not affect balanced section calculations. These calculations normally compare the area of a sedimentary sequence in the cross-section with the area of the same *compacted* beds in the undisturbed foreland. Therefore, the lithification compaction strain has already been accounted for and does not enter the calculation.

Tectonic compaction

There is some evidence that previously lithified sediments may be compacted further during deformation. Wood (1974) described an increase in density as mudstones are deformed to form slates. The density may increase from 2.5 up to 2.7–2.85 g/cm^3, equivalent to a volume reduction of 10%. In contrast, Siddans (1977) found that there was no density increase between undeformed and deformed sediments along a section in the French Alps. Hence, the little evidence that exists in the literature suggests that this kind of volume loss is less than 10%. It is likely that this loss will have biaxial or triaxial symmetry, but in order to simplify subsequent calculations (Fig. 5) the simple case of isotropic volume loss has been assumed. A 10% volume loss produces a 6.7% area reduction in a cross-section (Fig. 5).

Pressure solution

Von Plessman (1964) described the effects of tectonic pressure solution in cleaved rocks. Contractions of 20–30% are usual at right angles to the pressure solution cleavage and may go up to 50% (von Plessman, *op. cit.*; Alvarez *et al.* 1978). If the dissolved material is completely removed from the rock there will be a corresponding decrease in volume. However, much of the quartz and calcite which is dissolved away may be deposited locally within the rock in areas of lower pressure (Durney & Ramsay 1973; Elliott 1973; Alvarez *et al.* 1978). Veins and fibres of quartz and calcite are common, but not always present, in rocks showing pressure solution. It will be assumed in subsequent calculations that volume losses of 20–30% normal to the cleavage are possible by pressure solution and that this volume loss is likely to have uniaxial symmetry.

Elongations along orogenic strike

Most orogenic strains which have been measured lie within the flattening field of a Flinn plot (Ramsay & Wood 1973). An apparent flattening strain can be produced by the superposition of a volume loss on a plane strain (Ramsay & Wood, *op. cit.*). However, many of the natural strains would require volume losses of up to 60% to account for their symmetry by this mechanism. Therefore, most slates would appear to have undergone real flattening which involves extension along the Y finite strain axis.

Slaty cleavage is normally parallel or sub-parallel to the axial surfaces of folds. In the margins of orogenic belts, such as the Appalachians and Scandinavia, the axial surfaces and main cleavage are parallel to the strike of the orogenic belt (Cloos 1947). Geological cross-sections are usually drawn normal to the orogenic strike in the XZ finite strain plane (Cloos, *op. cit.*) with the Y axis normal to the plane of section. If Y suffers an elongation during the tectonic strain, there must be an area decrease in the XZ plane, even during constant volume deformation because

$$(1+\varepsilon_1)(1+\varepsilon_2)(1+\varepsilon_3)=1+\Delta$$

where Δ = the volume strain $\delta V/V$.

By assuming no volume strain, it is possible to calculate the average strike elongation of an area in an orogenic belt by integrating the measured finite strains along the Y finite strain trajectories (Hossack 1978). The integrated elongation along the Caledonian strike in the Bygdin area, Norway was found to be between 11 and 15%. I have used Cloos's data (1947) to determine the strain integration around the South Mountain Fold (Fig. 4) and here the strike elongation

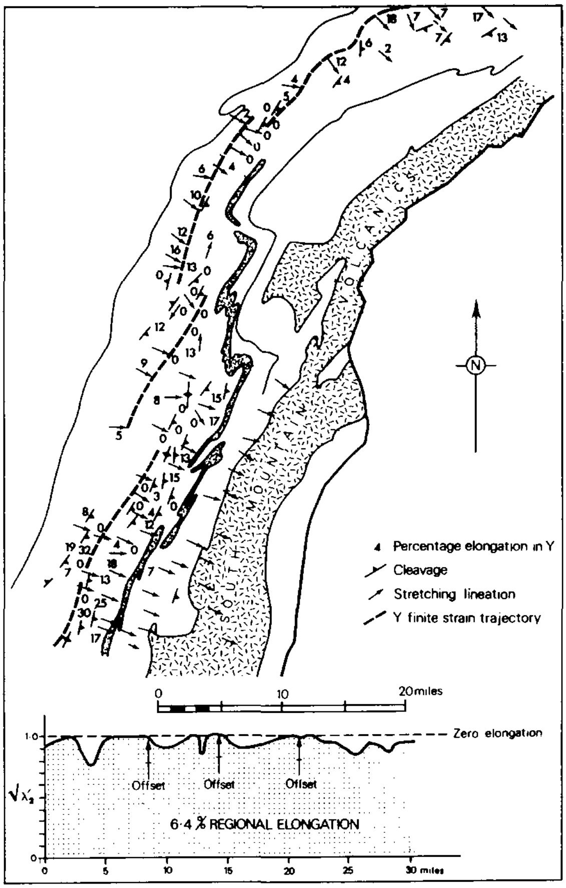

FIG. 4. Regional strain integration (Hossack 1978) around the South Mountain fold, Maryland (Cloos 1947). Y-finite strain trajectories drawn using the orientation of the cleavage traces and the stretching lineations. To complete the integration around the fold, several off-set trajectories have been chosen. Positions of off-sets are indicated on the integration curve. Total regional elongation, assuming no volume change, is 6.4%.

is 6.4%. This probably represents a real strike elongation, because Nickelsen (1966) described *Lingula* shells from the Allegheny foreland plateau which have extended by fracturing in the Y direction by an average of 4.4%. Borradaile (1979) suggested that the average strike elongation in the Dalradian of Islay was 8%. Hence, the average strike elongation from localities in the Scandinavian and Scottish Caledonides and the Appalachians suggests that strike elongations of 10% might be typical. This will produce an area change of 9% in an XZ balanced section where there is no volume change (Fig. 5). Strike elongations of up to 30% can produce area reductions of just over 20% in a section at right angles to the strike. Note, however, that if the strike direction of an orogenic belt suffers a shortening, there will be an area increase in the balanced section.

Errors in balanced section calculations

Strike elongation and tectonic compaction may each produce area reductions in geological cross-sections. Within the margins of orogenic belts each may typically have values of around 10%. Both kinds of area reduction are likely to have biaxial symmetry. Unfortunately there are an infinite number of ways by which two strain tensors can be combined. To simplify calculations, I have assumed that both kinds of area decrease have isotropic symmetry and that reductions in length are the same in all directions in the section. Fig. 5 has been prepared by combining varying values of strike elongation and tectonic compaction. Strike elongation combined with 10% tectonic compaction produces a total area decrease of 15% in a geological

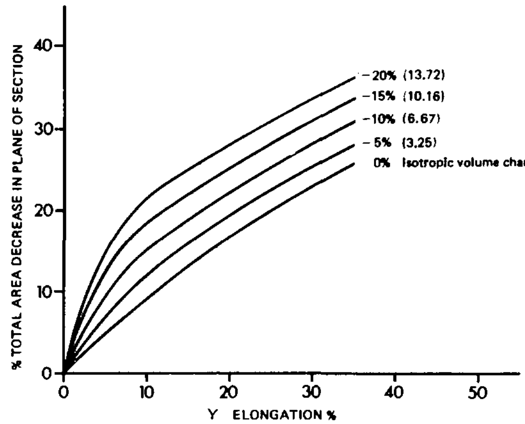

FIG. 5. Total area changes in geological cross-sections with strike (Y) elongation combined with isotropic tectonic compaction. Percentage isotropic volume decrease indicated against each curve. Figures in brackets represent the corresponding area decrease in the section brought about by the volume decrease.

section at right angles to the strike. This area reduction will affect the estimate of original length (l_0) and thickness (t_0) of the undistorted section. By assuming plane strain with no volume reduction, the balanced area calculation produces original lengths and thicknesses that are 15% too low. However, the technique at least provides a minimum result. It is likely that the original stratigraphic thickness of the Etnedal nappes was closer to 510 m, if there had been 15% area reduction, rather than the 443 m calculated with the plane strain assumption. Similarly the original length of the section could have been closer to 108 km rather than the calculated 94 km. The former figure gives an orogenic contraction of 50% compared to the 43% calculated by assuming plane strain.

The effects of pressure solution are much more difficult to estimate. There is not much evidence in the literature describing what happens to material that is dissolved away from the solution seams. Some authors (Durney & Ramsay 1973; Elliott 1973, Alvarez *et al.* 1978) assumed that the material is precipitated locally in areas of lower pressure and that the total volume of rock is preserved. With no volume change, the pressure solution would produce an additional plane strain of 20 or 30% which could be added by tensor multiplication (Elliott 1972) to the ductile strain used to restore beds back to their pre-deformational thickness (Fig. 3). However, if the dissolved material is completely removed out of the plane of the section, there will be an area decrease of 20–30%. This decrease does not have isotropic symmetry but occurs in one direction normal to the solution seams. Many orogenic belts seem to fall into 2 possible geometric models. Firstly, many belts have crudely horizontal beds with near-vertical slaty cleavage (e.g. the N Wales slate belt, Wood 1973). In this case, if there is 20–30% pressure solution without precipitation, the calculated original length of the belt (l_0) will be 20–30% too small.

The Etnedal nappes seem to fall into the second geometric model where there are crudely horizontal beds with a near-horizontal slaty cleavage. Any loss of material in a vertical direction will obviously affect the original stratigraphic thickness (t_0), which may be 20–30% too small. Hence, allowing for possible unidirectional shortening as a result of pressure solution, the original beds of the Etnedal nappes may have been up to 610–660 m thick rather than the 510 m calculated by allowing for isotropic area changes. Once again, the assumption of plane strain with no area change provides a minimum estimate.

There is an obvious need for mass transfer estimates to be carried out on rocks containing evidence of pressure solution to find out how much material actually leaves the rock, how much remains behind to be precipitated, and how far the removed material can be transported. When such estimates are available, more refined balanced section calculations will be possible.

Discussion

Balanced section calculations are an important tool for geologists because it is possible to estimate in a simple manner the orogenic contraction across large areas of the Earth's crust. The most simple calculation assumes that the area of a geological cross-section is the same before and after strain. Hence, the original length of the section is given by the area of the section divided by the original stratigraphic thickness of the beds in the section. Fortunately, the calculation is insensitive to errors or variations in the assumed original stratigraphic thickness. In spite of probable area decreases in the plane of the section during tectonic strain, balanced section calculations always lead to a minimum estimate of the amount of orogenic contraction. Balanced section calculations have been carried out in the marginal parts of several orogenic belts. The marginal thrust belt of the Appalachians has contracted by 35–43% (Dennison & Woodward 1963; Gwinn 1970; Kiefer & Dennison 1972), the Alberta thrust belt by 54% (Price & Mountjoy 1970) and the Etnedal nappes of S Norway by 43% (this paper). It is unlikely that balanced section calculations can be used in the internal parts of orogenic belts, but here the Eulerian destraining technique of Schwerdtner (1977) and Borradaile (1979) can be used to remove tectonic finite strain. It may eventually be possible to unstrain complete orogenic belts by using balanced section calculations and the Schwerdtner technique.

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