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Strain variation in thrust sheets across the Sevier fold-and-thrust belt (Idaho–Utah–Wyoming): implications for section restoration and wedge taper evolution

GAUTAM MITRA

Department of Geological Sciences, University of Rochester, Rochester, NY 14627, U.S.A.

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Abstract—Although strain in fold-and-thrust belts is typically inhomogeneous, on the largest scale strain within thrust sheets shows regular patterns. The strain, which is a quantitative measure of the deformation of any thrust sheet is independent of the translation of the sheet. The internal deformation can be represented by a deformation profile (the change in shape during thrusting of a fault perpendicular line) which can then be used to unstrain the sheet during restoration in cross-section balancing. Deformation profiles in well-studied thrust sheets show regular geometric patterns, and compatibility constraints can be used to minimize the data needed to define the profiles in other sheets. From the geometry of deformation profiles and direct field observations, the most important component of internal deformation of thrust sheets in many fold-and-thrust belts is early layer parallel shortening (LPS). This can be removed in a straightforward manner during restoration, and significantly affects the geometry (i.e. taper) of the restored section. Original basin wedge taper may have important implications for the mechanical evolution of fold-and-thrust belts. Thus changes in original basin taper resulting from strain removal may affect significantly the way in which we interpret the evolution of a fold-and-thrust belt in an area.

INTRODUCTION

THE large-scale geometry of many fold-and-thrust belts (FTBs), which are an important component of most major convergent orogens, is fairly well known (Bally *et al.* 1966, Dahlstrom 1970, Boyer & Elliott 1982, Woodward 1985). The concept of balanced cross-sections (Dahlstrom 1969, Elliott 1983, Woodward *et al.* 1989), together with surface and subsurface (well and seismic) data, has been used to draw fairly well constrained pictures of the three-dimensional geometry of FTBs and the sedimentary basins from which they evolved. However, most balanced sections and their restorations are inaccurate in detail, because they fail to incorporate information on strain within thrust sheets and detailed kinematic information that would allow stepwise retrodeformation and removal of strain increments (McNaught 1990, Protzman & Mitra 1990, McNaught & Mitra in press).

In recent years, available balanced sections and restorations have been used (e.g. Boyer 1991, Coogan 1992a,b) to test mechanical models (e.g. wedge mechanics of Davis *et al.* 1983) of FTB evolution, and of the effects of pre-orogenic basin configuration (i.e. initial wedge shape) on FTB evolution. It is clear that incorporating strain information would not only improve restorations of balanced sections, but would also clarify our understanding of the mechanical evolution of FTBs. However, the paucity of strain data in most FTBs, and the daunting task of gathering such information across an entire FTB, has discouraged workers from incorporating such information into their analyses.

In this paper I will address two main questions. First, what are the typical patterns of strain variation within thrust sheets and is it possible to simplify the process of incorporating strain into a balanced section by using these patterns? Answering this question will allow us to set up the methodology to incorporate strain into sections. Second, does ignoring the penetrative strain in thrust sheets significantly affect our interpretation of the kinematics and mechanics of emplacement of thrusts? To answer this question I will use as an example part of the Sevier thrust belt in the Cordilleras of the western U.S.A.; in this FTB the large-scale geometry is well constrained (Royse *et al.* 1975, Dixon 1982, Lamerson 1983, Coogan and Royse 1990) and sufficient strain data are available to allow a preliminary restoration incorporating strain across the entire FTB (Mitra & Yonkee 1985, Craddock *et al.* 1988, Mitra & Protzman 1988, McNaught 1990, Yonkee 1990, Craddock 1992).

STRAIN VARIATION WITHIN THRUST SHEETS

The fundamental theorem of deformation geometry says that any deformation can be decomposed into a strain, a rotation and a translation (Means 1990). For the emplacement of any thrust sheet the total displacement vector field can be conveniently broken down into two components (Fig. 1) (Elliott 1977).

(1) Bulk translation along the thrust fault, which can be measured as the mean slip of the middle of the sheet in a transport-parallel section. In practical terms the slip is conveniently measured from offset ramp cutoffs; flats or décollement horizons give us no indication of the total slip on a fault.

(2) Internal deformation of the thrust sheet, which includes both rigid body rotation and pure strain. This

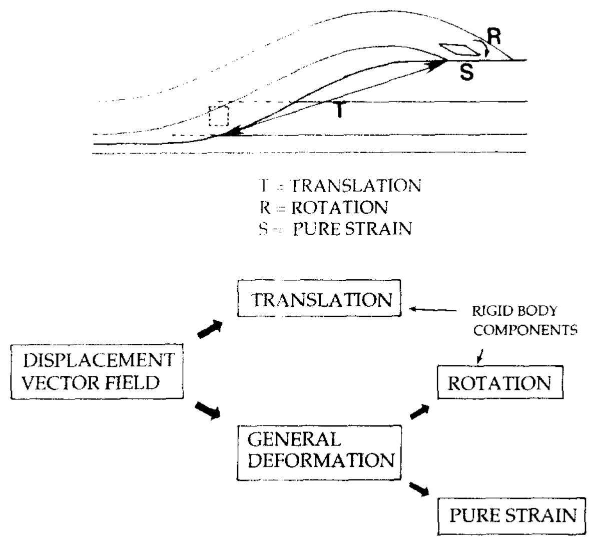

Fig. 1. The main components of the total displacement vector field shown for the case of emplacement of a thrust sheet.

can be measured as a finite strain which may include pre-thrust layer parallel shortening as well as any internal deformation that takes place during thrusting.

The translation and deformation components are independent, and it is possible to have a far-traveled thrust sheet with little deformation, or a sheet with large finite strains that has not moved very far (Elliott 1977). This has been documented in many thrust belts and I cite only a few examples to illustrate the point. The Lewis thrust is one of the largest in the North American Cordillera, and has a minimum displacement of 35 km (Dahlstrom *et al.* 1962, Dahlstrom 1970), yet the thrust sheet shows very little internal deformation and contains perfectly preserved undeformed sedimentary structures and stromatolites in Proterozoic Belt rocks. On the other hand, the Morcles thrust sheet which is part of a large antiformal stack in the western Helvetic Alps (Boyer & Elliott 1982) has a translation of only 13 km (Trumpy 1980) yet shows large amounts of internal deformation (Ramsay & Huber 1983). The classic work of Cloos (1971) using deformed oolites to show regional strain variations in the Appalachians indicates significantly larger strains in the hanging wall than in the footwall of small (1–2 km slip) thrusts (e.g. Wiliamsport thrust) in the Great Valley Province of Maryland (Edwards 1978).

Most balanced sections are restored by removing the translation component (the slip) on the fault and the rigid body rotation component represented mainly by large-scale dip changes due to fault bend or fault propagation folding. However, the pure strain component of deformation is rarely accounted for in section restoration. Pure strain within the thrust sheet can be accommodated by different combinations of small (cm to 10 m)-scale folding and faulting, fracture and vein systems, and penetrative fabrics formed by plastic or diffusional deformation processes; the structures that develop depend on the lithologies involved and the conditions of deformation (e.g. Mitra 1979, 1987). Mesoscopic folding strains can be measured by the sinuous bed length method, and several methods exist for determining strain from mesoscopic fault–fracture systems (e.g. Wojtal, 1982, 1986). Penetrative strains can be measured by using various strain markers (fossils, oolites, pebbles, etc.) and grain center-to-center techniques (Ramsay & Huber 1983, Erslev 1988, Erslev & Ge 1990). Pressure solution strains can be determined from toothed stylolites (Stockdale 1926) or from volume of insolubles in pressure solution seams vs host rocks (Yonkee 1983).

The internal deformation component can be a significant part of the total displacement field. Its inclusion in section restoration ideally requires a complete description of the strain variation in a thrust sheet. Only a few thrust sheets (from different FTBs) have had their strains adequately studied to allow such a description: these include the Morcles nappe in the Swiss Alps (Ramsay *et al.* 1983, Ramsay & Huber 1983), the Moine thrust sheet in Scotland (Coward & Kim 1981, Coward *et al.* 1992), the Valdres nappe in the Bygdin area of Norway (Hossack 1968, 1978) and the North Mountain thrust sheet in the Central Appalachians (Cloos 1971, Woodward 1985). When such detailed strain information is available it is possible to do a complete section restoration incorporating strain; this is done by integrating strain along strain trajectories and unstraining appropriate portions of the section by appropriate amounts (Hossack 1978).

The problem can be simplified by applying compatibility constraints (Cutler & Elliott 1983). If we assume that a thrust sheet is a continuum with smoothly varying strains, then compatibility is satisfied within the sheet; in such a case, only two complete strain measures (magnitude and orientation) along any given strain trajectory are sufficient information to calculate strain magnitude and orientation anywhere else along that trajectory, and therefore to completely define the variation in strain along that trajectory. Unfortunately, in most thrust sheets, major lithologic boundaries may behave as 'discontinuities' across which there are large strain gradients; thus additional data may be required to completely define strain variation within the sheet. For a continuous thrust sheet, the problem can be further simplified if the displacement geometry of the thrust sheet can be approximated by a simple model, such as inhomogeneous simple shear parallel to the fault; for this particular displacement geometry there is no strain variation in the transport direction and only two strain measures can completely define strain within the entire sheet (Cutler & Cobbold 1985). Most natural thrust sheets show more complex strain patterns than this, with an overall decrease in strain from the back to the front of the sheet; nevertheless, compatibility constraints can be used to minimize the amount of data needed to characterize strain variation within the sheet.

In order to use the compatability constraint to characterize strain variation within a sheet we need to know the finite strain trajectories. These can be defined by the variation in orientation of penetrative cleavage in internal sheets, but are more difficult to define in external sheets. Unfortunately, such information is typically lacking from most FTBs making it difficult to apply

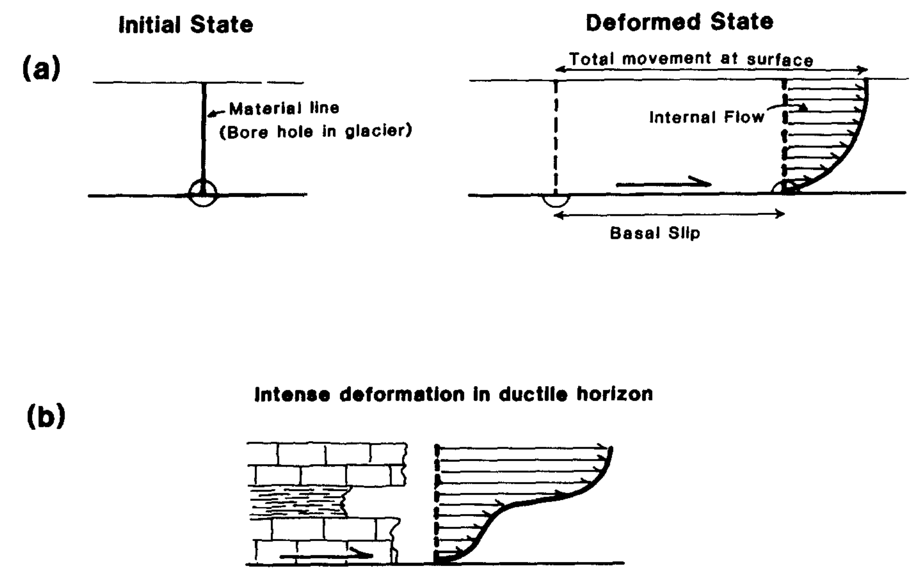

Fig. 2. (a) The deformation profile is defined by the change in shape of a passive marker perpendicular to the basal slip surface. The marker is also translated in the direction of motion due to basal slip. (b) The deformation profile may be complex in a thrust sheet with varying lithologies.

compatibility constraints without first gathering large amounts of cleavage orientation data. Where such information is available, strain trajectories in thrust sheets typically show a concave-up pattern, asymptotic to the thrusts bounding the base and the top of the thrust sheet; examples include the Morcles nappe (Ramsay *et al.* 1983, Ramsay & Huber 1983) and the Glarus nappe (Groshong *et al.* 1984) from the Swiss Alps, the Gavernie nappe from the Pyrennees (Choukroune & Seguret 1973), the North Mountain and Blue Ridge thrust sheets of the Appalachians (Mitra & Elliott 1980, Boyer & Mitra 1988, Boyer 1992), and the Willard thrust sheet in the Sevier thrust belt in Utah (Yonkee 1990).

Based on the information available from such well-developed thrust sheets it is clear that thrust sheets, as a rule, show a systematic strain distribution. Therefore, the variation in displacement due to internal deformation of the sheet should also show some regular patterns that could be accounted for in section restoration. One of the simplest ways of achieving this is by the use of deformation profiles within thrust sheets.

DEFORMATION PROFILES

The internal deformation of any moving sheet (e.g. a glacier or a thrust sheet) can be represented by a deformation profile (Fig. 2), which shows the relative displacement of points within the sheets due to the internal deformation. The profile is defined by the final shape of a passive marker that was originally perpendicular to the slip surface at the base of the sheet. For example, a vertical bore-hole in a glacier is progressively bent over a period of years into a concave-up curve as the ice close to the surface flows faster than the ice near the base of the glacier (Sharp 1960, Paterson 1969). Displacement vectors connecting the initial to the final shape (of the marker) define the deformation profile. If the translation vector or slip at the base of the sheet is added in we can obtain the displacement profile for the sheet.

Within a thrust sheet we can imagine a similar marker that was originally fault-perpendicular before movement on the fault. Because most thrust faults are bed-parallel over much of their extent and because they cut through originally horizontal beds, the imaginary marker would be vertical before fault movement. The marker is similar to an undeformed state loose line (Geiser 1988). Loose lines are used to keep track of relative movement of beds in forward or reverse models of thrust emplacement, and serve to point out obvious inconsistencies in the movement picture that arise in certain balanced sections. A deformation profile, on the other hand, is obtained from strain data within the thrust sheet and, therefore, describes the actual movement picture for the sheet; it thus places constraints on how a proper restoration should be done. To completely account for all of the internal deformation in a sheet we need to bring the deformation profile back to its original fault-perpendicular position during restoration.

Much of the internal deformation of a thrust sheet takes place by fault parallel (and therefore bed-parallel) shear, and we can define three simple end-case geometries for the deformed marker which describes the deformation profile (Fig. 3) in cross-sections drawn parallel to the transport direction.

(1) If there is no internal deformation of the sheet or if all the internal deformation is achieved by layer-parallel shortening, the marker will remain vertical and will be displaced due to slip (translation) on the fault.

(2) If there is homogeneous simple shear within the sheet (with or without layer parallel shortening or extension) the marker will remain straight but will be tilted in the transport direction, with displacement of individual points on the marker being proportional to their elevation above the fault.

(3) If there is inhomogeneous simple shear within the

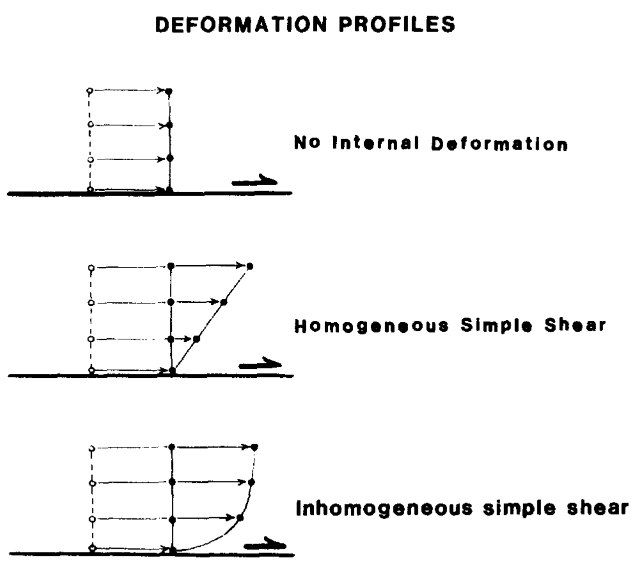

Fig. 3. Three end-case geometries of deformation profiles based on possible strain variations within thrust sheets (see text for details).

sheet (with or without layer parallel shortening or extension) the marker will become curved, convex toward the transport direction (i.e. concave-up), with points higher in the sheet having traveled farther than points near the fault.

In natural examples, there may be complete and continuous variation between these three end-member types, so that a particular example may fall somewhere within a triangle (Fig. 4). Natural examples will also generally show second-order shape changes in the deformation profile due to variations in lithology (and hence mechanical behavior) within a sheet (Fig. 2). For example, the presence of weak horizons can result in zones of intense slip within a thrust sheet such as those seen in the Pequop Mountains in the Basin and Range Province in Nevada (Fig. 5) (Mitra & Protzman 1988)

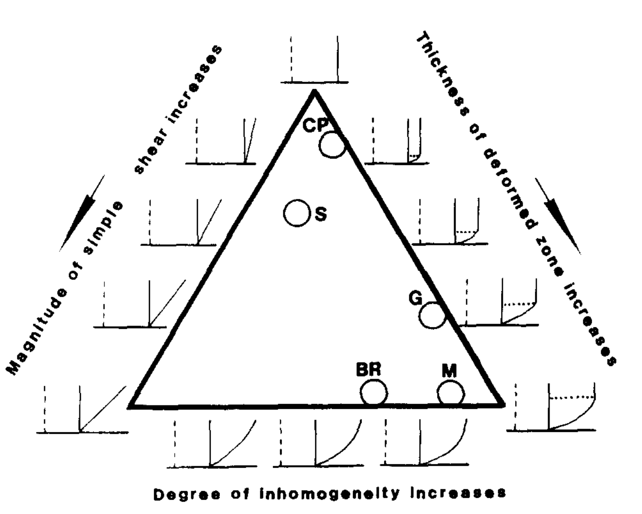

Fig. 4. Triangle diagram showing possible variations in shape of deformation profiles due to variations in magnitude of simple shear, thickness of simple shear zone at base of sheet, and degree of inhomogeneity in sheet movement. Deformation profiles from Fig. 6 are roughly plotted to show range of variations seen in natural examples: Cumberland Plateau (CP), Glarus (G), Särv (S), Morcles (M) and Blue Ridge (BR) thrust sheets.

giving rise to a 'bundled profile' (Protzman 1990). But, by and large, thrust sheets have fairly simple deformation profiles when viewed at the largest scale—asymptotic to the fault at the base, becoming steep within the body of the sheet. For inclusion of strain in regional scale restorations such 'first-order' deformation profiles provide adequate information.

The deformation profile may be recorded by original vertical markers, such as dikes, within the sheet. The Särv thrust sheet in the Scandinavian Caledonides (Gilotti & Kumpulainen 1985) is an excellent example, with vertical (bed perpendicular) dikes showing very sharp curvature in the lower 15 m of the sheet (Gilotti 1989) and becoming asymptotic to the thrust at the base (Fig. 6a). More commonly the profile must be obtained by integrating strain and strain trajectory information in internal thrust sheets—the Morcles (Ramsay *et al.* 1983) and Glarus (Groshong *et al.* 1984) sheets in cover rocks, and the Blue Ridge (Mitra 1979) sheet in basement rocks provide sufficient information to obtain deformation profiles (Fig. 6a). In external thrust sheets, internal deformation of the sheet can be measured from slip on networks of mesoscopic faults (Wojtal 1982, 1986), which directly provide a deformation profile—the Cumberland Plateau sheet in the Southern Appalachians (Wojtal 1986) is an excellent example (Figs. 6a & b).

All the deformation profiles obtained show similar concave-up shapes. Strain trajectories in many other internal sheets and displacement information in a few other external sheets (Wojtal 1982) suggest that they would have similar deformation profiles. The deformation profiles obtained vary over a wide range of scales both in the thickness of the basal zone of fault parallel shear and the total displacements related to internal deformation of the sheets (Fig. 6a). Typically, the zone of basal shear is thicker for sheets emplaced at higher metamorphic grade and there is more displacement related to the internal deformation. When the profiles are roughly plotted on Fig. 4 based on their shapes, the resulting plot suggests two generalizations: (1) natural deformation profiles tend to lie along the right-hand and lower edges of the triangle diagram; and (2) external thrust sheets tend to plot near the apex of the triangle, while more internal sheets plot lower down on the triangle.

The observed deformation profiles agree with theoretical deformation profiles calculated assuming laminar flow parallel to the fault (Wojtal 1992). The shapes of the calculated profiles depend on the stress exponent in the flow equation, with the curvature decreasing for lower values of the stress exponent. In natural thrust sheets emplaced at high metamorphic grades, the observed microtextures suggest deformation mechanisms that would lead to low stress exponent flows; the observed profiles (e.g. Morcles and Blue Ridge sheets) with low curvature and a thick basal zone of shear show very good agreement with theoretical predictions. External sheets where much of the internal deformation takes place by cataclastic flow typically show sharply curved profiles with a relatively thin basal zone of shear

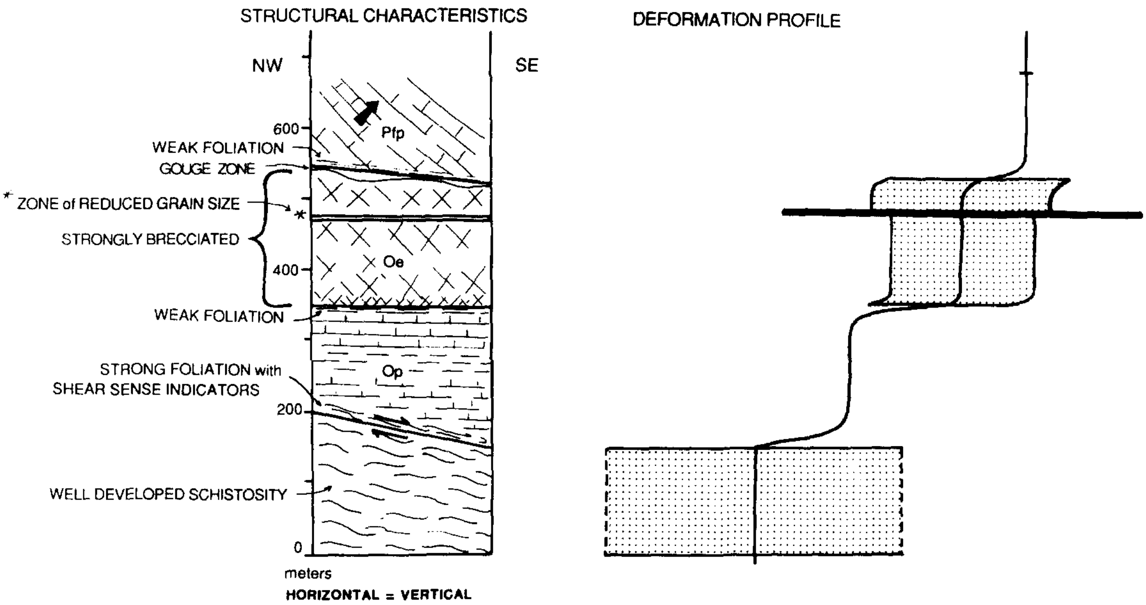

Fig. 5. An example of 'bundled deformation profile' with several weak zones of intense shearing from the Pequop Mountains, Nevada.

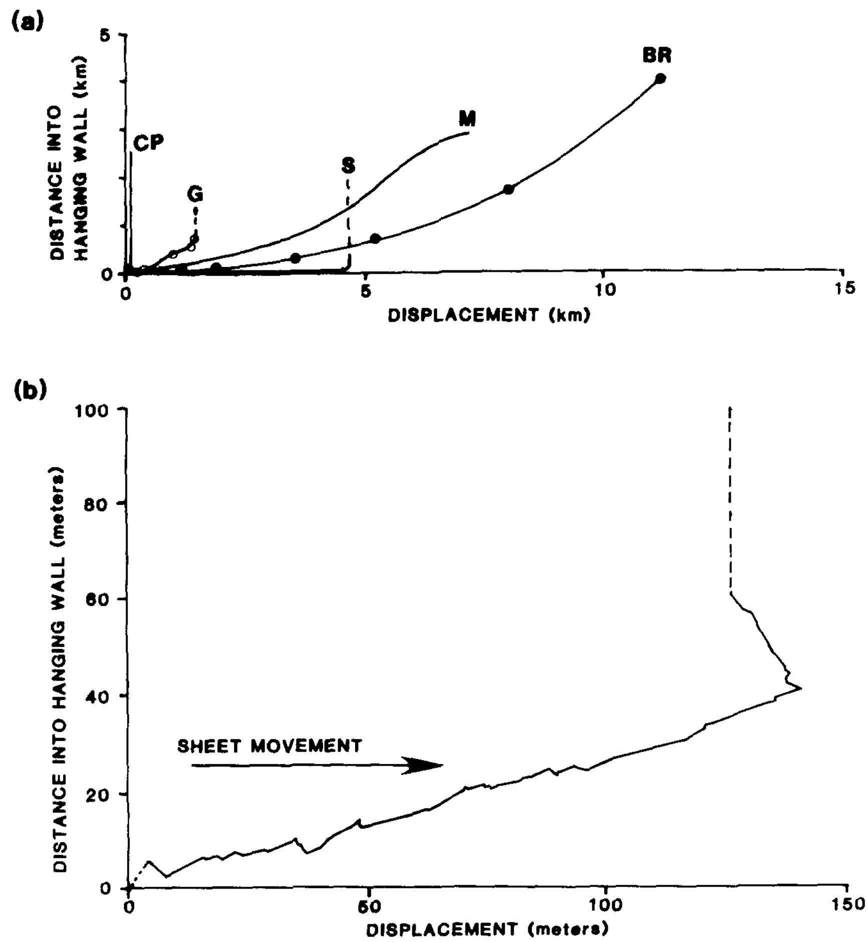

Fig. 6. (a) Deformation profiles of the Cumberland Plateau (CP), Glarus (G), Särv (S), Morcles (M) and Blue Ridge (BR) sheets compared at the same scale. Note the general pattern of increasing thickness of basal shear zone from foreland to hinterland. (b) Composite deformation profile of the Cumberland Plateau thrust sheet, Southern Appalachians (after Wojtal, 1982). The basal shear zone extends ~60 m into the hanging wall.

(e.g. Cumberland Plateau sheet), suggesting that the deformation can be approximated by high exponent flow. The exact shape of the deformation profile, of course, also depends on the total internal deformation in the sheet.

The curvature of any deformation profile is a reflection of the large amounts of inhomogeneous simple shear at the base of the sheet, with pure shear or bulk flattening type of layer parallel shortening (LPS) becoming increasingly important higher up in the main body of the sheet (Boyer & Mitra 1988). The simple shear dominated zone at the base of the sheet can also include a component of LPS to satisfy compatibility constraints (Sanderson 1982). The simple shear displacement in the basal zone simply adds to the translation (slip) on the basal fault, and is, in many cases, only a small fraction of the total slip (particularly in the external portions of FTBs); thus ignoring this simple shear component in such regions would not generally cause major errors in cross-section restorations where fault slip had been accounted for. However, in sheets where the simple shear zone at the base is thick (e.g. Morcles sheet), the displacement in the zone may be significant and must be accounted for during restoration (Ramsay *et al.* 1983).

The LPS component in the main body of the sheet records internal shortening of the sheet and is completely independent of the slip on the basal fault. Far away from the fault (high up within the sheet) the deformation involves no internal rotation and strains record only the LPS component. This shortening component is usually ignored in section restorations and is often large enough that it may result in significant errors in the geometry of the restored sedimentary wedge. The LPS component may be recorded in the form of different structures in different lithotectonic units depending on deformation mechanisms active under the conditions of emplacement; but the structures should record equivalent amounts of shortening in the different lithotectonic units unless there are major observable weak zones in the stratigraphic package. LPS strain can be quantified in different stratigraphic units by appropriate methods and can be incorporated into cross-sections in a straightforward manner.

Layer parallel shortening typically occurs early in the deformation history during thrusting. In the New York Plateau area of the Central Appalachians LPS has been shown to extend for more than 100 km in front of the most external Valley and Ridge (FTB) structures (Engelder & Geiser 1979). Shortening of 10–15% is recorded in Devonian rocks above a Silurian salt detachment, suggesting that a LPS front migrated out in front of an advancing thrust system. In the Pennsylvania Valley and Ridge, cross-cutting relationships indicate the presence of several generations of small-scale structures (Nickelsen 1979, Gray 1991) that formed during blind thrusting; the earliest and most predominant of these structures is a penetrative LPS fabric that was modified by later deformation (Gray & Mitra 1993). In the Idaho–Wyoming–Utah salient of the Sevier FTB cleaved clasts from successive synorogenic conglomerates can be used to show that LPS fabric was developed in successive thrust sheets immediately before each sheet was emplaced (Mitra *et al.* 1984). In the external sheets (e.g. Darby, Absaroka, Crawford) the early LPS fabric was modified by later, simple structures resulting in fanning of bed-perpendicular cleavage around folds (Mitra & Yonkee 1985). In more internal sheets where the structures are more complex (e.g. Meade sheet), detailed structural information from both the footwall (Protzman & Mitra 1990) and hanging wall (McNaught 1990, McNaught & Mitra in press) allow stepwise restoration which indicates the presence of an early LPS fabric (10–35% shortening, Protzman & Mitra 1990) that was completely modified by later structures.

In summary, while LPS is not the only internal deformation in most thrust sheets (particularly internal sheets), in many cases it may be the largest component of internal shortening affecting sedimentary wedge taper. The shape of the deformation profile is a useful guide to determining the relative importance of basal shear vs LPS in the evolution of the geometry of the sheet. In a thrust sheet with a gently curved deformation profile and a thick basal zone of shearing, the dominant component of deformation affecting the shape of the sheet may be fault parallel shear (Ramsay *et al.* 1983). In such a case a complete restoration should include removal of the strain by unshearing along interpreted (or assumed) displacement paths (Coward 1980, Ramsay *et al.* 1983, Casey & Huggenberger 1985, Ramsay & Huber 1987). If, on the other hand, a thrust sheet has a sharply curved deformation profile (in some cases modified by later deformation) with a thin zone of shearing at the base, LPS most likely played a significant role in wedge taper evolution and should be removed from restorations. Since LPS is typically developed early in the deformation history of each thrust sheet, removal of this shortening from a cross-section is fairly straightforward because it can be done as a last step in the restoration and requires minimal strain information gathered from weakly deformed rocks high up within the sheet. Thus LPS can be easily removed (as described later) from regional scale restorations. To test the effect of strain removal on regional-scale restorations, this method was used to restore the Idaho–Utah–Wyoming fold-and-thrust belt and the results compared with earlier restorations (that did not account for strain).

REGIONAL GEOLOGY

The Idaho–Utah–Wyoming fold-and-thrust belt (FTB) is part of the Sevier orogenic belt in the North American Cordillera (Fig. 7). A series of W-dipping thrust faults transported parts of the pre-existing Paleozoic and early Mesozoic miogeocline eastward during late Cretaceous and early Tertiary times (Armstrong & Oriel 1965). To the east the thrust belt impinges on the foreland basin and locally interacts with adjacent Laramide uplifts (Dorr *et al.* 1977, Hunter 1987). Westward the hinterland is more difficult to interpret because

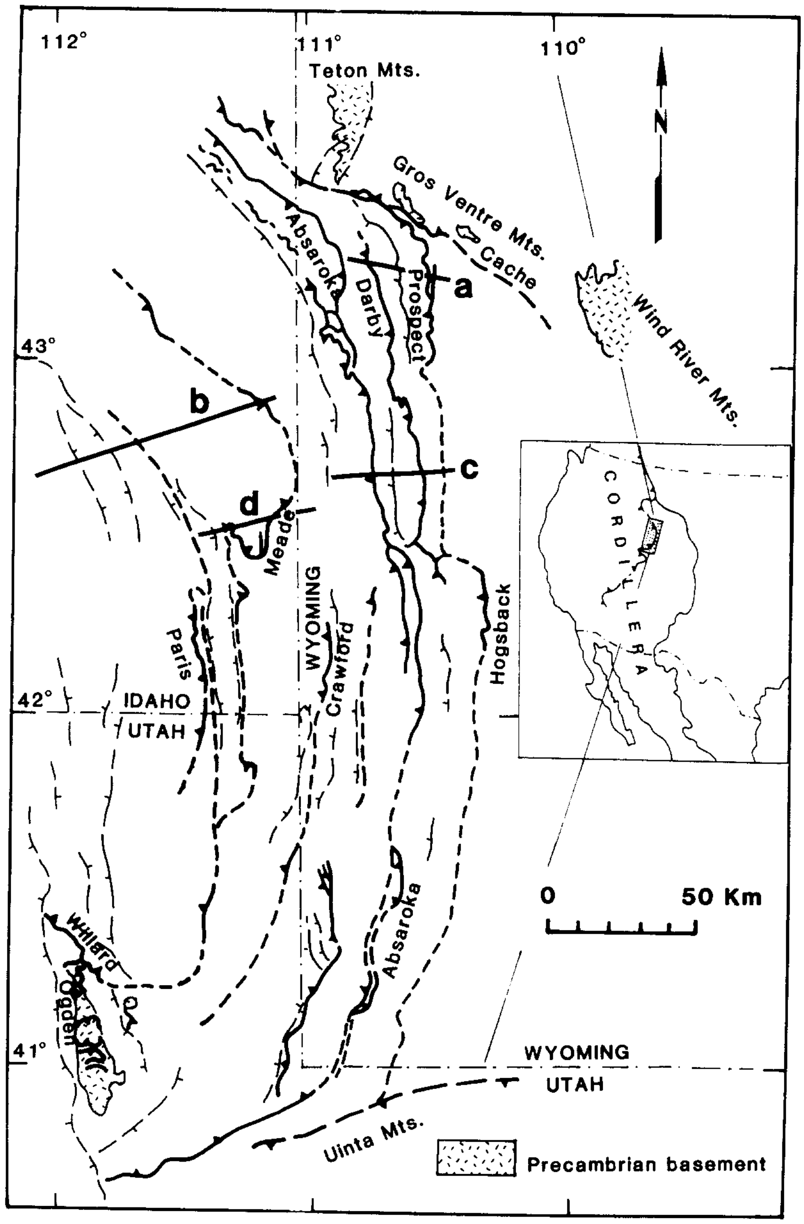

Fig. 7. Tectonic map of Idaho–Utah–Wyoming fold-and-thrust belt showing the main thrust faults and listric normal faults (after Coogan & Royse 1990). Locations of cross-sections (Figs. 9a–d) are shown.

Tertiary normal faults of the Basin and Range Province obscure earlier Mesozoic deformation.

The structures of the eastern part of the Idaho–Utah–Wyoming FTB are typical of the "foothills family of structures" (Dahlstrom 1970). Major thrust faults are listric in shape and join at depth to a low-angle décollement (Royse *et al.* 1975, Dixon 1982, Lamerson 1983) within the Cambrian section. This detachment drops into the Precambrian section toward the hinterland. In detail the major thrusts have a ramp–flat geometry with the flats along preferred glide horizons which are stratigraphically controlled (Royse *et al.* 1975, Coogan & Yonkee 1985). Large-scale folding in the thrust belt is the result of movement of hanging wall ramps onto footwall flats. Smaller-scale folds develop above imbricate fans that occur along some of the major thrusts. Listric normal faults, bounding a number of major strike-parallel valleys, merge into thrusts, usually at footwall ramps. Farther west normal faults become more common but their relationship to the thrust belt structures is less clear (Royse *et al.* 1975, Burgel *et al.* 1987).

Eight major thrusts make up the Idaho–Utah–Wyoming FTB. From west to east they are the Willard, Paris, Ogden, Meade, Crawford, Absaroka, Darby and Prospect thrusts (Fig. 7). The timing of movement of these thrusts is known from cross-cutting relationships and synorogenic conglomerates (Armstrong & Oriel 1965, Oriel & Armstrong 1966, Royse *et al.* 1975, Wiltschko & Dorr 1983, Heller *et al.* 1986, DeCelles 1988). These dates indicate that the westernmost thrusts moved first and subsequent movement proceeded sequentially eastward; the older thrusts continued to be active periodically through much of the thrusting history (Yonkee *et al.* 1992, DeCelles in press). Although fault zones themselves are usually not well exposed, field observations suggest that the zones of shearing associated with the major faults are relatively thin compared to the thickness of the thrust sheets. Thus, these thrust sheets have sharply curved deformation profiles with thin basal shear zones, and any internal shortening higher up in the sheets would have significantly affected the geometry of the sheets and regional wedge taper.

The exposed rocks in the external portion of the FTB were deformed at low metamorphic grade ($P < 5$ kb, $T < 200°C$) and show the development of a spaced cleavage in the Jurassic Twin Creek Formation (Imlay 1967, Mitra *et al.* 1984). Studies of cleaved Twin Creek pebbles from synorogenic conglomerates indicate that the cleavage developed before the conglomerates were shed off their respective sheets, and cleavage developed progressively from west to east as successive thrust sheets were emplaced (Mitra *et al.* 1984). Our own detailed studies of the fabrics within the Meade, Crawford and Absaroka thrust sheets have shown that the cleavage developed early in the deformation history of each thrust sheet (Mitra & Yonkee 1985, Protzman & Mitra 1990). The relationship of cleavage to structural geometry supports this conclusion. Cleavage is axial planar to small folds and formed concurrently with small-scale buckle folds and contraction faults just prior to or during the earliest movement on each thrust indicating an early layer parallel shortening (LPS) (Mitra & Yonkee 1985). Cleavage strongly fans about large-scale folds formed during later movement, suggesting passive rotation of cleavage with bedding during the later stages of thrust history. The rotated cleavage may also be folded and overprinted by a second weaker cleavage during a second ramping, or movement on a later, lower thrust. Cross folding may also develop as a result of lateral ramping. Thus, complex patterns of interfering folds, multiple cleavages and associated fiber filled fractures may be produced during motion on a single thrust fault (Mitra & Yonkee 1985).

The early LPS represented by cleavage within the thrust sheets is also recorded by other markers (deformed fossils, weak penetrative strain fabrics in oolites and sandstones, and small-scale folding and faulting). The internal shortening of sheets can be quantified, and because the sheets have the appropriate (sharply curved) deformation profiles the strains can be removed in a simple manner during section restoration.

STRAIN VARIATION IN THRUST SHEETS

In the Idaho–Utah–Wyoming FTB cleavage in the Twin Creek Formation represents significant amounts of layer-parallel shortening within the thrust sheets. In the absence of major detachments within the stratigraphic section, strain compatibility requires equivalent strains within other horizons or smoothly changing strains through the stratigraphic package. Preliminary studies indicate that, depending on lithology, the strain in other units can be accommodated by mesoscopic fracturing, faulting and folding and/or microscopic penetrative strains. The penetrative strain is recorded by different strain markers that are present at intervals throughout the Paleozoic–Mesozoic section (McNaught 1990) (Fig. 8) in the ID–UT–WY FTB; the most useful strain markers are Pentacrinus fossils, as well as oolites and sandstones on which the Fry center-to-center technique can be used. All these strain measures yield strain axial ratios in the plane of measurement, rather than absolute shortening or extension values. If we assume plane strain in sections parallel to the transport direction, the appropriate shortening values in the plane of section can be calculated.

Strains from different lithologic units within the same thrust sheet can be compared with one another to check for consistency in strain values. Typically the long axis (X) of the strain ellipsoid is bed-perpendicular, the intermediate (Y) and short (Z axes) are bed-parallel with Z being parallel to transport; for convenience and ease of comparison I have kept the orientations of these axes (relative to bedding) the same even where the shape of the ellipsoid does not follow the typical pattern. In the frontal part of the Prospect thrust sheet Jurassic Nugget sandstones and Jurassic Twin Creek oolites yield XZ strain ratios of 0.87 (i.e. bedding plane fabric) to 1.22 corresponding to maximum LPS values of ~9% (Fig. 9, section a). Pentacrinus fossils (which typically lie on bedding) in Twin Creek limestones show YZ (bedding plane) strain ratios of 1.15 to 1.17 corresponding to plane strain shortening of 13–15% (Fig. 10, cross-section line a). In the back part of the same sheet, shortening due to early mesoscopic (outcrop scale) wedging is 28.6% (an equal-area axial ratio of 1.8) in folded Mississippian Madison limestone that shows little or no grain-scale deformation (Fig. 9, section a). In the middle of the Absaroka thrust sheet (near the southern part of Star Valley, Wyoming), Nugget sandstones yield XZ strain ratios of 1.6 corresponding to 21% shortening (Fig. 9, section c); YZ bedding plane strains from Pentacrinus fossils in Twin Creek limestones average 1.32 corresponding to plane strain shortening of ~25% (Fig. 10 north of lines c and d). In the Sheep Creek slice of the Crawford sheet XZ strain ratios in the Nugget, Timothy and Thaynes sandstones vary from 1.24 to 1.27 corresponding to shortening values of 11–12% (Fig. 9, section d). The strains reported here also include shortening due to small (cm to 10 m)-scale folding and wedging where such structures are observed. In all of these cases there is little variation in shortening from one unit to the next in the stratigraphic sequence, and there is generally an increase in LPS from the front to the back of each thrust sheet.

Micritic limestone of the Jurassic Twin Creek Formation deform dominantly by pressure solution, forming a well-defined LPS fabric. Shortening can be estimated from the density of pressure solution seams and the proportion of insolubles in host rocks vs seams (Yonkee 1983). Alternatively the intensity of cleavage can be used to obtain a rough estimate of shortening using a scaling similar to Alvarez *et al.* (1978) but calibrated to the Twin Creek micrites. Layer parallel shortening can

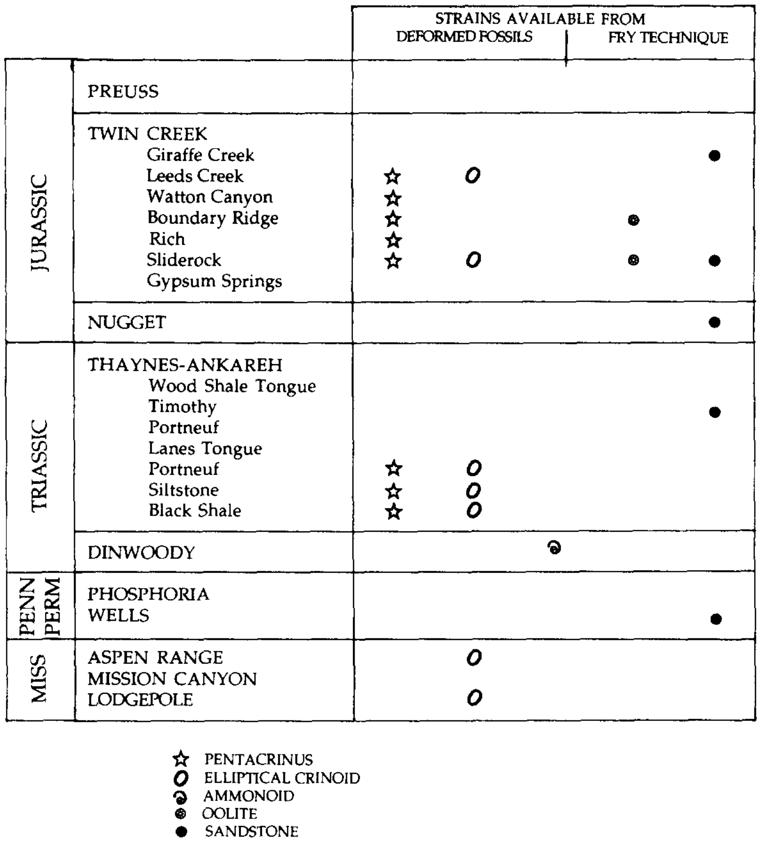

Fig. 8. Part of the Paleozoic–Mesozoic section in the Idaho–Utah–Wyoming fold-and-thrust belt showing identified horizons of oolites and sandstones suitable for Fry analysis and horizons with good fossil strain markers (McNaught personal communication).

be estimated in this way over a wide area by mapping cleavage intensity (Mitra & Yonkee 1985), and these LPS values agree with those obtained from other strain markers (Fig. 10). The strain obtained by the various methods can then be used to obtain a picture of the regional variation in strain within the thrust sheets.

For this analysis, the shortening within individual thrust sheets due to twinning (Craddock *et al.* 1988, Craddock 1992), pressure solution (Yonkee 1983, Mitra & Yonkee 1985) and plastic deformation (deformed fossils) (Protzman 1985, McNaught 1990) were compiled (where available) and determined (in other areas). The strain is partitioned differently for different lithologies, and the appropriate components have to be added to obtain the total strain at a particular location. Since strain varies from the back to the front of each sheet, strains were determined for the back and front of each sheet and the mean calculated (assuming that strain decreases linearly from the back to the front of the sheet).

The mean strain due to internal deformation within successive thrust sheets shows an interesting pattern of variation (Fig. 11). The strains progressively increase from the front to the back of the FTB, with internal thrust sheets showing as much as 35% mean internal shortening (Fig. 12). In the Ogden thrust system, which involves basement, both basement and cover rocks show large amounts of shortening (Yonkee 1990); the somewhat larger shortening values in the cover compared to the basement (Fig. 12) may be due to scale-problems in measuring strain in basement or due to the presence of small-scale detachments between basement and cover. The Willard and Paris sheets (farthest west) have very little internal strain, although information on these sheets is very sketchy and detailed work is currently being carried out (Yonkee personal communication).

To compare the internal shortening of thrust sheets to shortening due to slip on the faults the latter were recalculated as shortening strains by comparing the final length of each thrust sheet to its restored length. The strain estimate for each thrust obviously depends on how a base length is chosen for calculating the strain. In each case, the base length was measured from the trailing edge of the thrust sheet to the leading edge of the next (future) thrust sheet; e.g. for the Absaroka sheet the base length for strain determination is measured from the trailing edge of the Absaroka sheet to the future trace of the Darby sheet. Thus these strains are averaged over long base-lengths that overlap with one another for successive thrust sheets. While the actual slip on individual faults varies, the resulting shortening strains for successive thrust sheets are remarkably consistent. These strains are typically larger than those due to internal deformation within the sheets (Fig. 12), although thrust sheets such as the Meade and the Crawford which are closer to the hinterland show internal strains comparable to the fault slip strains. As mentioned before, the fault slip strains are the ones typically

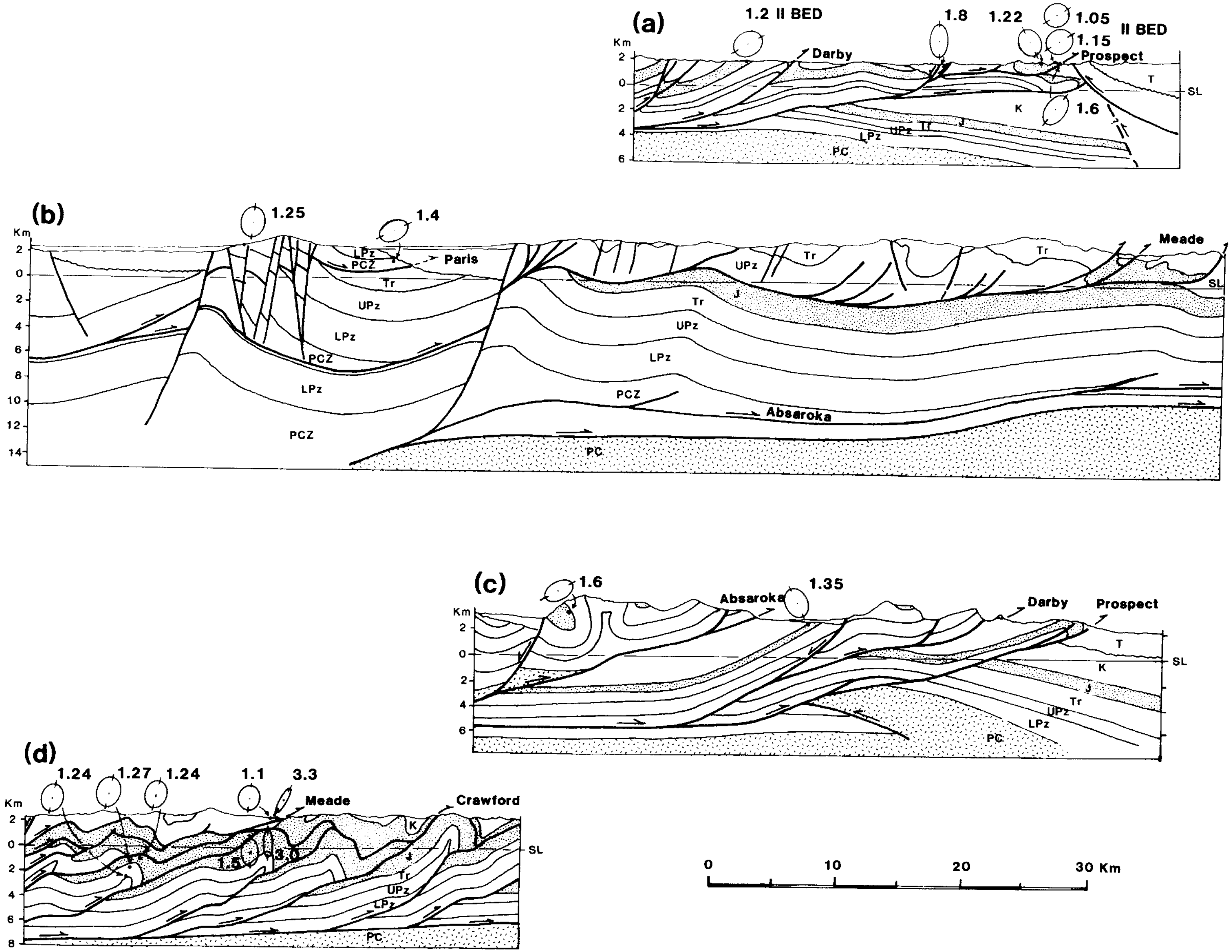

Fig. 9. Cross-sections of parts of the ID–WY FTB showing typical XZ strain ratios determined from Fry plots on oolites and sandstones. Locations of cross-sections are shown on Figs. 7 and 10. (a) Strain variation in frontal part of the Prospect thrust sheet. (b) Strains within the Paris sheet in SE Idaho. (c) Strains in the Absaroka and Darby sheets, southeast of Star Valley, Wyoming. (d) Strains in the Meade sheet and the Sheep Creek slice of the Crawford system.

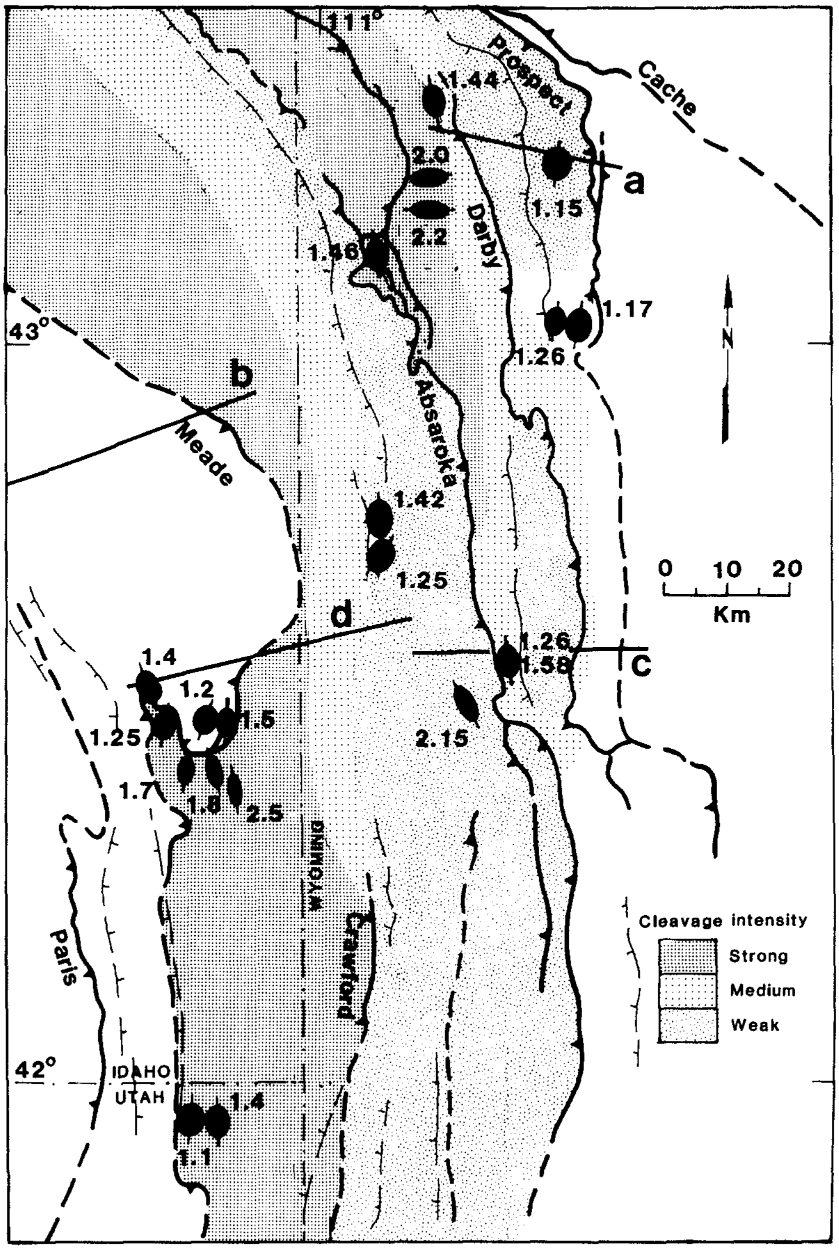

Fig. 10. Map of part of the northern ID-WY FTB showing typical bedding plane strains (YZ) determined from deformed Pentacrinus fossils, and from Fry plots on sandstones and oolites. The map also shows intensity of cleavage development in the Twin Creek Formation for parts of the area. Locations of cross-sections (Fig. 9) are shown.

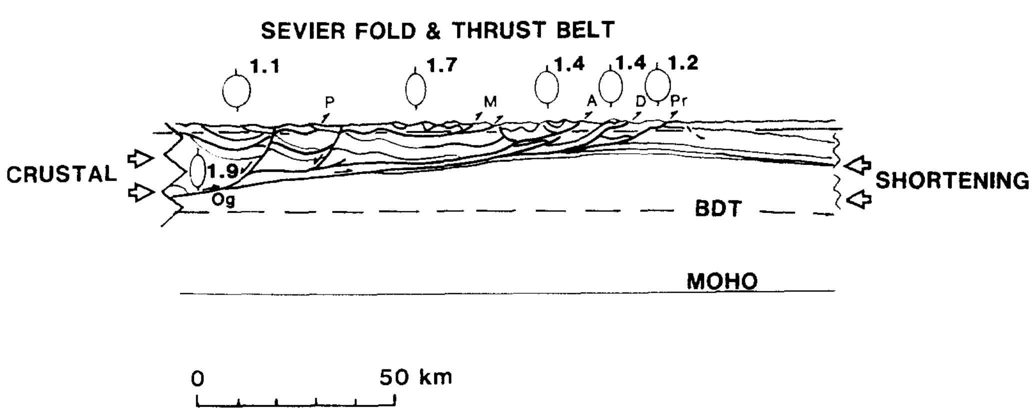

Fig. 11. Mean XZ strain axial ratio due to layer parallel shortening in the major thrust sheets of the ID–UT–WY FTB. Thrusts shown are Prospect (Pr), Darby (D), Absaroka (A), Meade (M), Paris (P) and Ogden (Og).

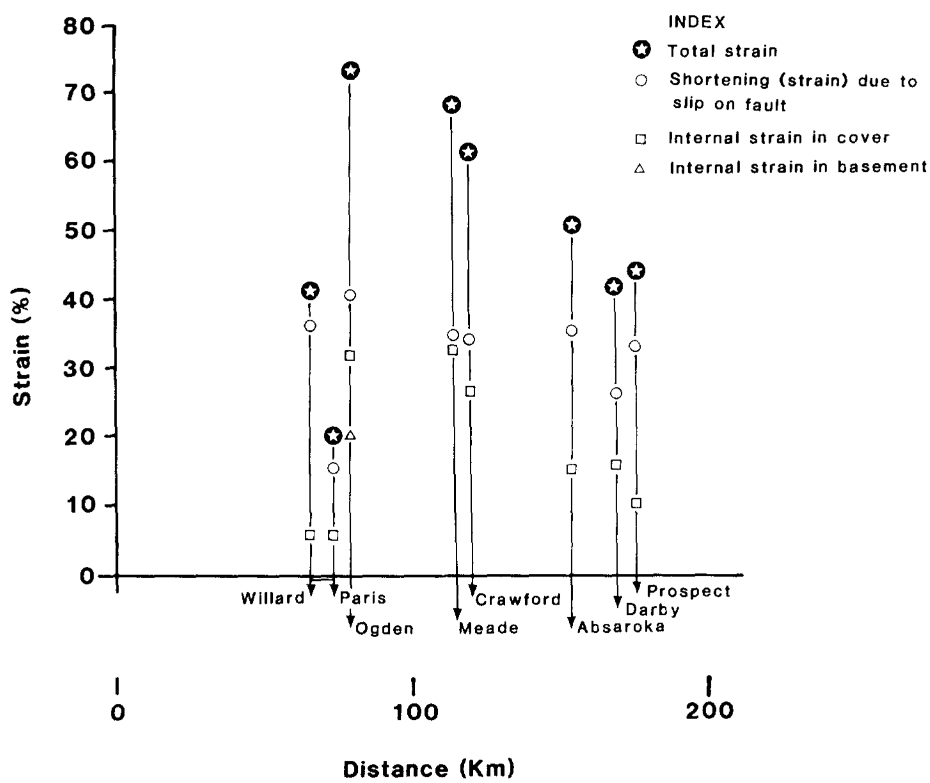

Fig. 12. Variations in internal strain, strain due to slip on major faults, and total strain (all expressed as % shortening) in the main thrust sheets of the ID–UT–WY FTB. Note the dramatic drop in internal strain in the Willard and Paris thrust sheets.

accounted for in cross-section restorations (e.g. Royse *et al.* 1975, plate IV).

STRAIN IN CROSS-SECTIONS

The problems of finding an appropriate method for cross-section balancing have been addressed by many authors (e.g. Geiser 1988, Woodward *et al.* 1989, Mitra & Namson 1989). Both line length and equal-area balancing methods have their own problems depending on the data available. A combination of equal-area balancing and line-balancing a stiff layer usually produces the best results (Geiser 1988, Mitra & Namson 1989). However, it is important to know the original thickness of the 'stiff layer' as clearly pointed out by Mitra & Namson (1989, fig. 3), because this can lead to major variations in the restoration. The only way of independently checking the original thickness of a marker layer is to determine its internal strain (typically LPS) and take that away to arrive at the original thickness (Fig. 13).

Strains as small as 10% can result in significant changes in restored lengths (Fig. 14). Depending on the conditions of deformation, the strain can be equal-area plane strain which results in thinning and lengthening the section in restorations (Fig. 14), or it can be LPS with volume loss which simply results in lengthening of the section on restorations. In the ID–UT–WY FTB, none of the rocks show evidence for large amounts of volume loss; while some limestones (e.g. certain members of the Jurassic Twin Creek Formation) show large amounts of pressure solution, nearby layers within the same formation show large volumes of calcite veins. Because of this, strains were incorporated into restorations using the assumption of equal-area plane strain along the line of section. While the error produced by not including strain in the restoration of a single external thrust sheet may be small, the errors are cumulative as we proceed to the back of the thrust belt (Fig. 14).

To test the effect of including strain in cross-section restoration the northern cross-section of the Idaho–Wyoming FTB (Royse *et al.* 1975, plate IV) was used as a base. The sheet lengths and thicknesses were recalculated using the internal strain. Figure 15 shows step-wise restoration of successive thrust sheets to their pre-deformation state. Markers spaced 25 miles (40 km) apart in the deformed state are shown in the successive restorations to give a visual impression; markers for restorations not including internal strain (Royse *et al.* 1975) are shown for comparison.

To look at the effect of removing internal strain on the taper of the initial wedge, the tapers for successive sheets were recalculated (Fig. 16) and compared with tapers shown on the northern cross-section of the ID–WY FTB (Royse *et al.* 1975). For each thrust sheet the pre-deformation taper was directly measured from the surface and basal slopes. The taper was also calculated from the length of the sheet and its thickness in front and in back. Assuming that strain decreases linearly from the back to the front of each sheet, sheet lengths were calculated using the mean internal strain within each sheet. The front and back thicknesses were recalculated using internal strains determined for the front and back of the sheet, respectively. Using these new restored

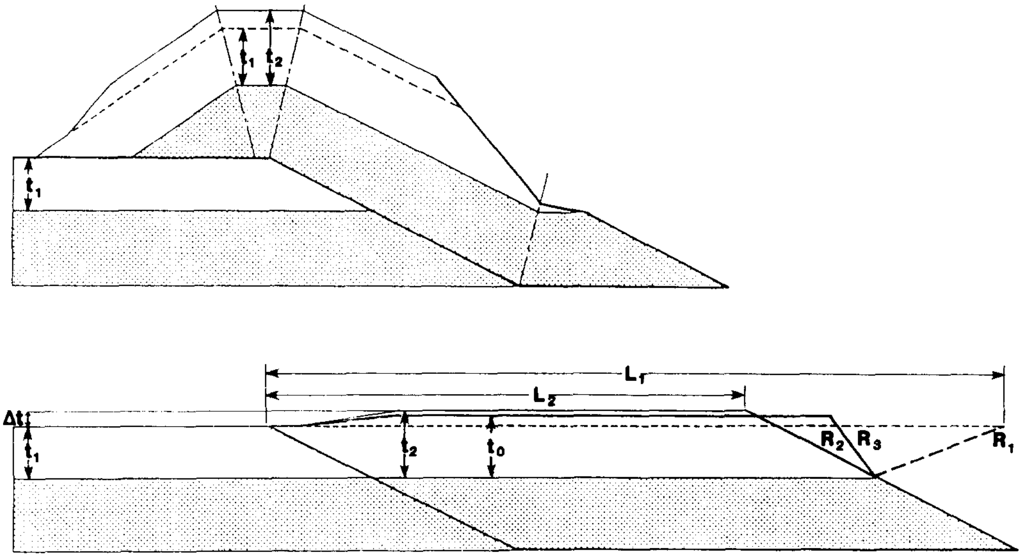

Fig. 13. Effects of variation in original thickness on the combined equal area and key-bed restoration method (after Mitra & Namson 1989). Incorrect estimation of the original thickness (t_1) leads to an overestimate of the bed length (L_1) and an unlikely ramp geometry (R_1). A correct original thickness (t_2) (based on 10% shortening and thickening) yields bed length L_2 and a reasonable ramp geometry R_2. Note that other strain values (e.g. 18%) would yield a different original thickness (t_0), leading to another reasonable ramp geometry (R_3); there is considerable leeway in the slope of the ramp dependent on strain in the key-bed.

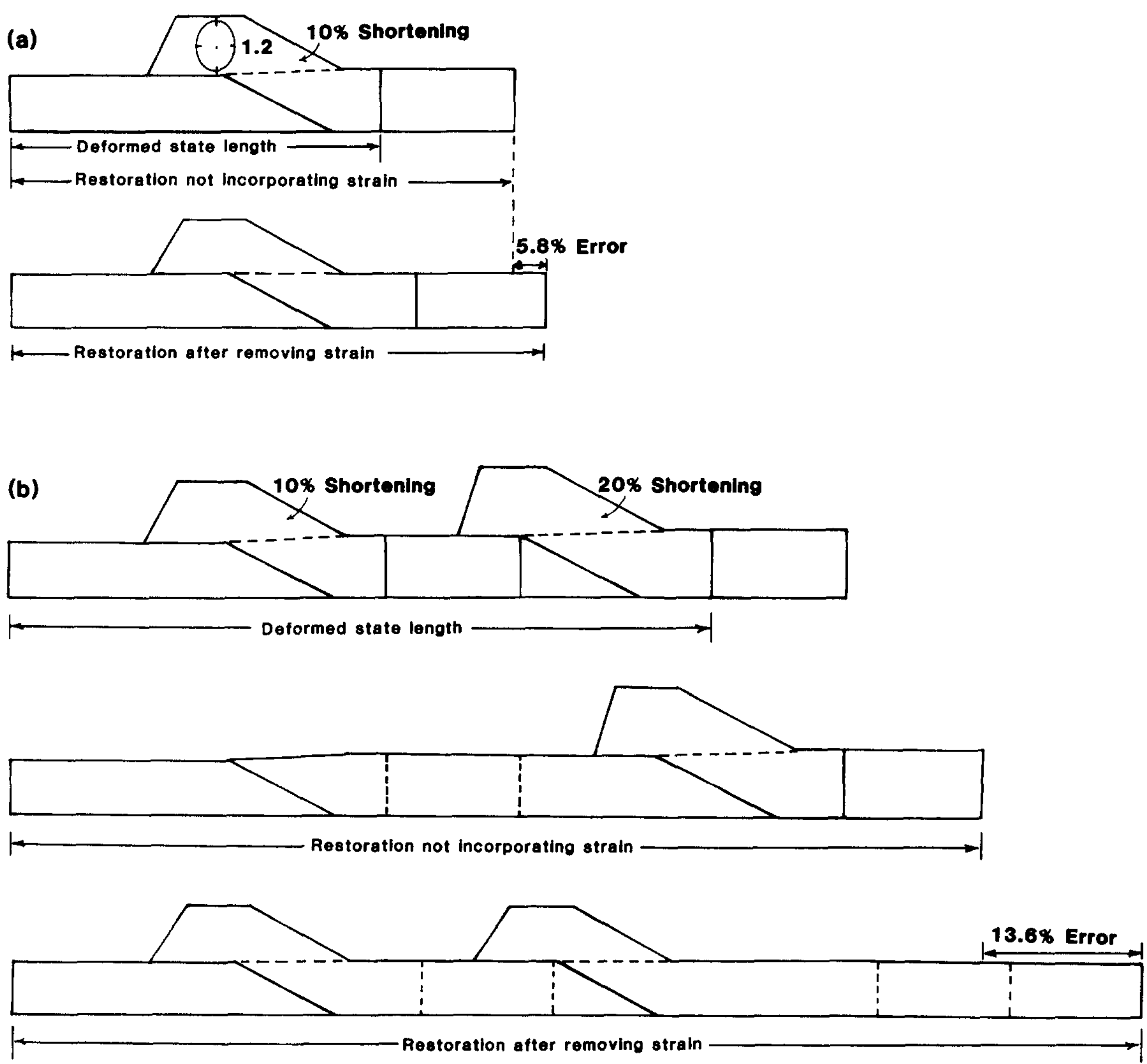

Fig. 14. (a) 10% plane strain internal shortening results in a 5.8% error in line-length balanced equal-area restoration (b) Balancing error is cumulative. Two sheets with 10 and 20% plane strain internal shortening results in 13.6% error at the trailing edge of the second sheet in line-length balanced equal-area restoration.

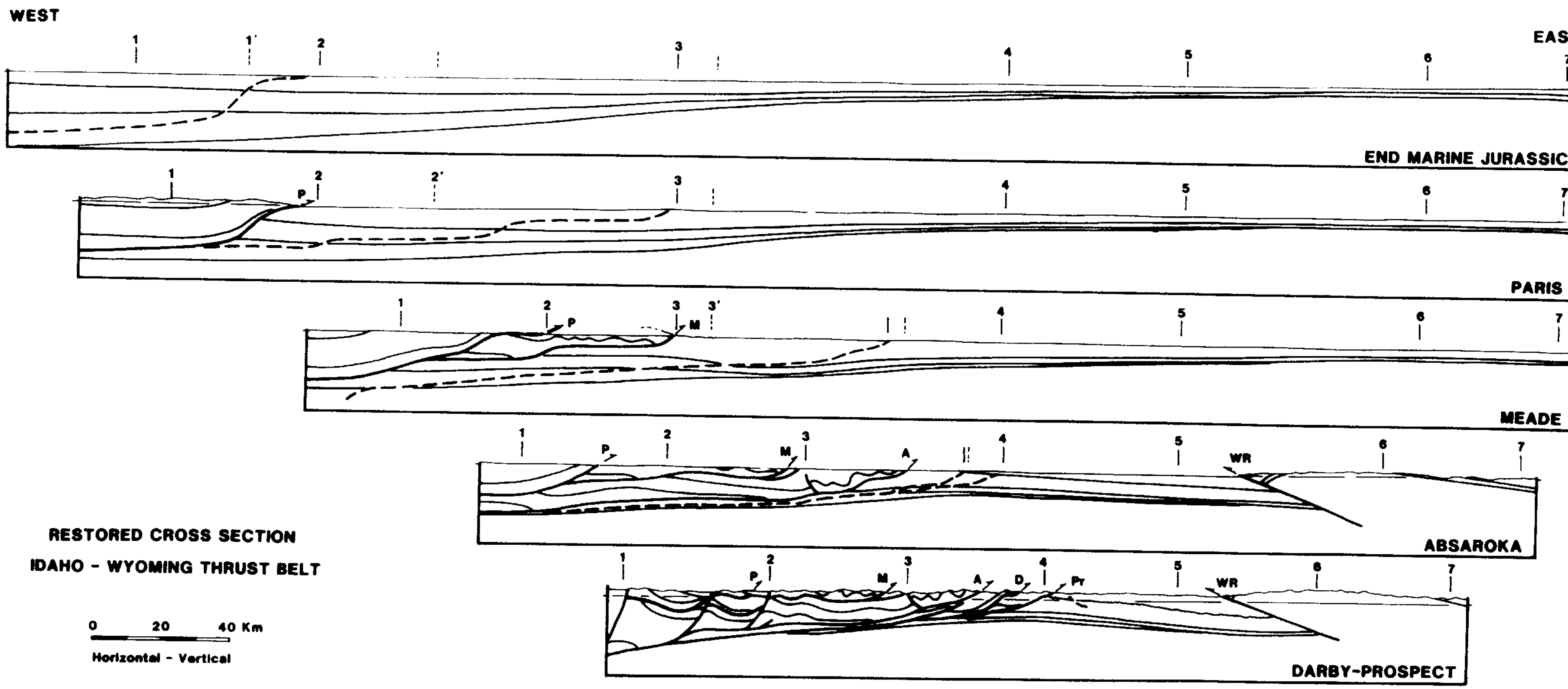

Fig. 15. Restoration of the northern ID–WY FTB cross-section (Royse *et al.* 1975) incorporating mean strains within individual sheets. Markers set up 25 miles (40 km) apart in the deformed state are used in the restoration. Note the difference in location of markers when strain is included in restoration (1, 2, 3, ...) vs when strain is not included (1′, 2′, 3′, ... after Royse *et al.* 1975). Thrust faults shown are Prospect (Pr), Darby (D), Absaroka (A), Meade (M) and Paris (P).

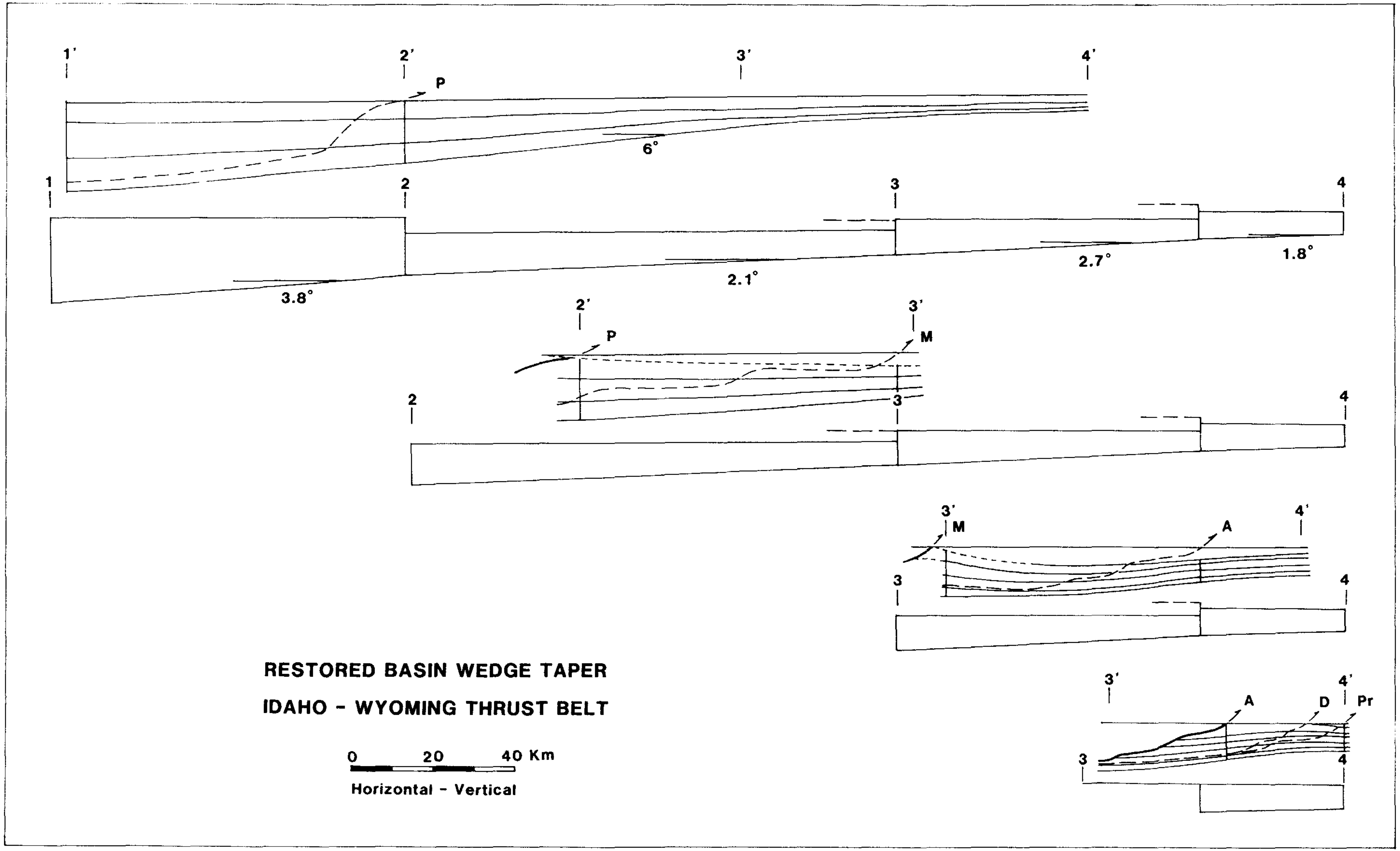

Fig. 16. Recalculated taper angles of original sedimentary prism from different thrust sheets of the ID–UT–WY FTB during stepwise restoration. Compare these with the 6° taper on the Royse *et al.* (1975) restoration. For each sheet, length is recalculated using mean strain, and thicknesses at the leading and trailing edges are calculated using strains at those locations. Reference markers (1, 2, . . .) originally 40 km apart are used in each restoration, and the restoration of Royse *et al.* (1975) with markers (1', 2', . . .) are shown for comparison.

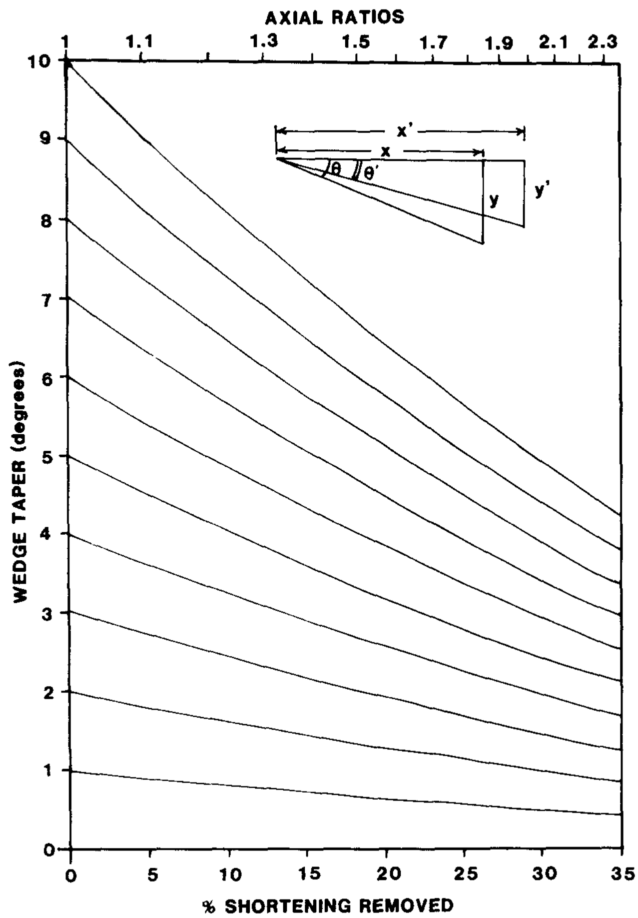

Fig. 17. Change in taper of wedges due to removal of different amounts of strain (see text for details).

lengths and thicknesses the initial taper was recalculated (Fig. 16); in most cases there is a significant decrease in the initial taper when internal strain is removed from the sheet.

We can look at the effect of removing strain on taper reduction in a more general way. Assuming plane strain, wedge taper can be recalculated for any given strain (Fig. 17). For a taper angle θ, $\tan \theta = y/z$, where y is the thickness of the wedge at its thick end and x is its length. If shortening is removed from the wedge, it becomes longer and thinner. For equal-area plane strain, $T_1 T_2 = 1$, i.e. $T_1 = 1/T_2$, where T_1 and T_2 are the principal stretches.

The new wedge length $x' = T_1 x = x/T_2$.
The new thickness $y' = T_2 y$.
The new taper angle (θ') is given by

$$\tan \theta' = y'/x' = \tan \theta .\ T_2^2.$$

The results of removal of different amounts of internal strain from wedges of different initial taper (i.e. no strain removed) are shown in Fig. 17. For example, if a wedge has a taper of 8°, as shortening is removed the taper is reduced along the sloping line; for 10% shortening strain removal the taper is reduced to 6.5°.

DISCUSSION

Recent work (Boyer 1991) has suggested a relationship between initial basin taper and mechanics of thrusting. A basin with low initial taper requires more internal strain to build up wedge taper. Most miogeoclinal wedges from which FTBs develop show a hinge line with the outboard part of the basin showing significantly larger taper. Thus initial thrusting (near the hinterland) would require little internal shortening of the wedge and internal shortening should increase as thrusting progresses onto the foreland. Unfortunately, although the hypothesis is elegant, it has only been tested on available restored basin geometries which are based on balanced restorations that did not incorporate internal strain. As shown earlier, when strain is incorporated into a restoration, the initial basin geometry may look quite different from a restoration that did not incorporate strain. Thus, even if Boyer's hypothesis works to a first order, interpretations for most thrust belts will tend to be inaccurate in detail unless strain is incorporated into the restoration. The restored tapers in Fig. 16 show a much larger taper for the westernmost (Paris–Willard) sheet than the other sheets; the westernmost sheets also show a noticeably smaller internal strain (Fig. 11) than the Meade and Crawford. This agrees with Boyer's hypothesis and suggests that the original basin hingeline lay just east of the westernmost thrusts.

Lateral variations in basin geometry (and therefore, initial taper) may result in different amounts of thrusting and different thrust geometries in different parts of a thrust belt (Boyer 1991). For example, an initial large taper in part of a basin may cause a single large thrust sheet to start moving while the adjoining parts of the basin are still building up taper before thrusting can start (Boyer 1991). The large sheet thus forms a salient providing a large proportion of detritus to the synorogenic foreland basin and building up the thickest foreland basin deposits in front of the salient. Once again, while this is reasonable to a first-order approximation, the interpretation of development of salients may be flawed in detail without strain data from the thrust sheets involved in adjoining salients and reentrants.

During progressive thrust development within a sedimentary wedge there is complex interaction between wedge shape, surface erosion, deposition of synorogenic sediments and thrusting (Coogan 1992a,b). Detailed studies on synorogenic deposits (DeCelles personal communication) are starting to yield timing and source information on these deposits that suggest a complex history of wedge shape evolution, with an overall forelandward sequence of thrusting modified by many episodes of out of sequence deformation. While it might be tempting to interpret such synorogenic information in terms of movement on thrusts alone, the deformation could also include internal strain resulting in shortening and thickening of individual thrust sheets. Strain and strain history information would need to be incorporated into the structural development history to completely and correctly interpret the information available from synorogenic deposits.

The method for incorporating strain in cross-sections suggested in this paper is obviously only a first-order approach to the problem. It takes into account the initial LPS fabric seen in the external portions of many thrust belts, and shows that including the LPS strain in restorations can make significant differences in the restoration. While LPS fabrics probably represent the single largest component of strain within thrust sheets in the external portions of FTBs, there can be many other components of strain in internal thrust sheets (e.g. Gray & Mitra 1993) that need to be taken into account in a complete restoration. While this has been done for single thrust sheets (e.g. McNaught & Mitra 1988, 1990, in press, Protzman & Mitra 1990), considerably more data need to be collected before complete restorations can be attempted across an entire FTB.

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