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Ramifications of four-dimensional progressive deformation in contractional mountain belts

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Abstract

Studies of progressive deformation aim to identify and establish temporal sequences of structural stages, each characterized by a suite of structures which form under specific rock and environmental conditions. Because of the natural spatial variation in rock type and environmental parameters in an orogen, several structural stages may operate concurrently in different parts of an orogenic wedge. At any given instant in time, rocks experiencing different stages of deformation are bounded by deformation fronts. As intrinsic and extrinsic conditions of deformation vary through time, spatial migration of deformation fronts causes rocks to record temporal overprinting of structural stages. In convergent orogens, deformation fronts, as mapped in their finite state, are typically forelandward-dipping due to the regional orogenic wedge geometry and thermal structure. Depending on the competing deformation processes, deformation rate changes on either side of the front may cause deformation fronts to change orientation, relief and surface area with time. Differences in migration rates of different deformation fronts could cause rearrangement of the sequence of structural stages with depth and laterally across the same mountain belt. © 1999 Elsevier Science Ltd. All rights reserved.

1. Introduction

The sequence of progressive deformation that we establish at the surface has an impact on the way in which we draw our cross-sections, how we might expect rock permeability to have changed through time, petroleum prospectivity, our understanding of rock kinematics, seismic hazard assessment and tectonic phenomena like the progressive development of accretionary prisms. Given the importance of progressive deformation studies, sophisticated approaches to regional studies are required. This paper examines some of the consequences of diachronous progressive deformation in mountain belts and presents some scenarios that justify further study.

Progressive deformation takes place in a series of temporal stages, each of which is defined by a particular family of mesoscopic and microscopic structures. In much the same way as metamorphic isograds separate rocks of different degrees of metamorphism in mountain belts, deformation fronts represent the boundary between rocks that have undergone (or are in the process of undergoing) different stages of deformation (Gray and Mitra, 1993) or the first occurrence of any deformation mechanism (Groshong, 1988). Early studies in metamorphic terranes often assumed that the lower-grade metamorphism recorded on one side of an isograd was also incurred on the other side of the boundary prior to the higher-grade metamorphism. Similarly, it was also commonly assumed that deformation fronts in mountain belts merely represented boundaries between rocks that had undergone different degrees of deformation (e.g. Cloos, 1953). More complexly deformed rocks were commonly assumed to have undergone all of the same stages of deformation as their less deformed neighbors (e.g. Arndt and Wood, 1960).

den Tex (1963) and Means (1976) pointed out that it

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is dangerous to make such spatial to temporal correlations. Deformation mechanisms and the structural stages they facilitate are dependent on variables such as temperature, pressure, grain size, and strain rate (Elliott, 1973). These parameters can vary with location at any instant in time as well as with position through time (Wojtal and Mitra, 1986). For example, at a folding deformation front, unfolded rocks on one side of the front may exhibit spaced, pressure solution cleavage. It is not safe to assume that the same cleavage is also present in the folded rocks, or if present, that the cleavage formed prior to the folds. Indeed, it is even difficult to try to correlate the same cleavage on the basis of orientation, morphology, and mineralogy from one location to the next (Mitra and Elliott, 1980; Tobisch and Paterson, 1988; Williams, 1988). What then are the implications of potentially diachronous development of a structural stage or multiple stages in an evolving mountain belt? When examining a deformation front at its final position, bounding two families of structures or different 'deformation-mechanism associations' (Groshong, 1988), it is necessary to distinguish between the possibility of once coeval but different deformation on the two sides of the boundary versus a finite deformation front that represents a boundary between times of deformation.

Studies of active mountain belts demonstrate that at any given point in time, different parts of the belts are subjected to different mechanisms, styles, and rates of deformation. The causes of this phenomenon are several. The internal architecture of the belt may vary dramatically or subtly with position. Topography and depth of burial also play a role in determining the character of deformation. Finally, environmental parameters (e.g. temperature) that exert controls on deformation are non-uniform and set up gradients in the belt. Development of the structure itself can significantly perturb these gradients (Oxburgh and Turcotte, 1974; Mitra and Yonkee, 1985). These factors taken together create a 'stratigraphy of deformation' both vertically and laterally across the belt.

Structural studies optimally can establish one-dimensional relative sequences of progressive deformation at any given outcrop. Extensive geochronological studies are required to unambiguously determine the absolute temporal relationships of all stages at each of many outcrops across a region. This approach may not be feasible in many cases and many geochronological methods have error ranges that exceed the time span required for the development of more than one stage. Even if we had access to such detailed and abundant geochronological data, would we know how to interpret it? For example, if the data indicated that adjacent areas in a mountain belt had opposite sequences of overprinting structures, what would that imply about the tectonic evolution of the mountain belt?

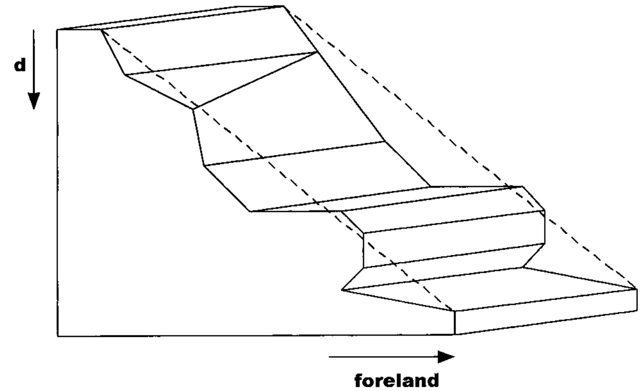

Fig. 1. A conceptual illustration of relief on a deformation front. As stratigraphy changes with depth and laterally, deformation progresses at different rates, producing slope changes in the deformation front. The average instantaneous orientation of the front may dip gently toward the foreland if the operating deformation mechanisms are strongly temperature-controlled.

Given that most geochronological data are derived from samples gathered at the surface, how can we project these sequential relationships to rocks below the surface (a necessary step in drawing a cross-section)? A convenient way to discuss the temporal–spatial relationships between structural stages is to focus on the behavior of deformation fronts. This paper examines the conceptual aspects of deformation fronts, explores the parameters that may control their behavior and relative geometries with time, and develops some predictive models. The potential impact of deformation front orientations and migration on thrust wedge development is also explored.

2. Deformation front orientations and geometries

The description of deformation fronts has mainly been limited to mapping their position at the surface. For example, Fellows (1943) mapped a curvilinear front that marks the transition from tectonites to non-tectonites in the central Appalachians. However, it is important to map deformation fronts in three dimensions because deformation fronts may have exerted controls on the gross architecture of the orogen as it developed (Gray and Stamatakos, 1997). Rather than linear boundaries, deformation fronts should be thought of as surfaces.

The simultaneous development of two different structural stages separated by a deformation front is largely an expression of different deformation mechanisms operating at the grain scale. Each deformation mechanism has its own flow law dependent on variables such as temperature, pressure, grain size and aspect ratio (e.g. Elliott, 1973; Rutter, 1976; White, 1976; Ashby and Verrall, 1978; Schmid, 1983).

Variables such as grain size, composition and permeability control deformation mechanisms and their strain rates, lateral and vertical variations in stratigraphy are likely to cause instantaneous deformation fronts to be saw-toothed in detail (Fig. 1). Some lithologies would undergo deformation more readily than others. This phenomenon is also expressed in the deformation profiles of deforming thrust sheets. Deformation profiles of large thrust sheets display first order strain gradients related to proximity to the thrust fault as well as second order perturbations in strain related to lithologic heterogeneity (Elliott, 1976; Mitra and Protzman, 1988; Mitra, 1994). In thrust sheets which exhibit strong lithologic controls on strain accumulation, the result is a 'bundled' deformation profile (Protzman, 1990; Nickelsen, 1993; Mitra, 1994).

The angle of inclination of a deformation front is dependent on the spatial variation of deformation mechanisms in the orogenic wedge. As temperature is a key parameter in most flow laws (Groshong, 1988), the thermal structure of an orogenic wedge will exert a control on the spatial distribution and orientation of deformation fronts within it. Holl and Anastasio (1995) have demonstrated that pressure solution cleavage fronts in the southern Pyrenees are parallel to paleoisotherms. Hwang and Wang (1993) have shown a direct correspondence between the spatial limit of deformation in the Taiwan fold and thrust belt and a downward deflection of isotherms.

Grain-scale finite strain studies in the central Appalachians have delineated tectonite fronts in their finite state (Cloos, 1971; Mitra, 1987; Evans and Dunne, 1991). A tectonite front is defined by Fellows (1943) as a boundary marking the first appearance of a crystallographic preferred orientation in a mountain belt. Evans and Dunne (1991) have noted that, in the central Appalachians, this level of fabric development coincides with rocks with X/Z finite strain ratios greater than 1.35, which also marks the threshold between low and high temperature deformation mechanisms in that locality. Based on their cross-sections and strain analyses in the North Mountain thrust sheet of Virginia, they show a tectonite front in its finite state that dips approximately 8° toward the foreland. Tectonite fronts, however, have less value in studies of progressive deformation because they are finite-strain-related surfaces that mark finite transitions in *amounts* of deformation; they are analogous to the transition from sedimentary to metamorphic rocks in the metamorphism of an orogen, and are, therefore, not abundant. Tectonite fronts do not carry incremental deformation information that is available from deformation-stage (mechanism)-related surfaces, which we refer to as deformation fronts. The latter, which are analogous to metamorphic isograds in an orogen, should be more abundant. Like deformation fronts a tectonite front is probably parallel to paleoisotherms (Evans and Dunne, 1991) and its orientation may give some indication of the general orientation of deformation fronts in the same orogen.

Magmatic and hydrothermal activity, characteristic of the internal portions of most mountain belts, set up horizontal thermal gradients by conduction and advection of hot fluids from the core of the orogen to the external portions of the orogen. Uplift and translation of deep-seated, hotter rocks by thrusting also produces lateral thermal gradients. Horizontal thermal gradients during regional metamorphism can potentially reach the same order of magnitude as vertical geothermal gradients (15–40°C/km) in the internal portions of mountain belts. It is likely, however, that horizontal thermal gradients are non-linear and that temperatures decrease rapidly away from heat sources in the internal mountain belt (Barr and Dahlen, 1989; Hwang and Wang, 1993; Ruppel and Hodges, 1994). While horizontal heat transfer by conduction is small to negligible in mountain belts (Sonder and Chamberlain, 1992), advective heat transfer across mountain belts may be accomplished by fluid migration (Oliver, 1986; Daniels et al., 1990) and emplacement of thrust sheets. Fluid temperature and flux can affect activation energies of deformation mechanisms, creating additional relief on deformation fronts and changes in migration rates of deformation fronts (Koons et al., 1998).

Topographic slopes, principally controlled by climate, in active mountain belts can also cause horizontal thermal gradients. Even small surface slopes can potentially generate different ambient temperatures at a given horizontal datum in the belt. Active fold and thrust belts feature surface slopes in the order of 4° (Davis et al., 1983). Over the width of a fold and thrust belt such as the Utah–Wyoming Sevier belt ($\sim$100 km), the surface slope alone would generate a horizontal gradient of 210°C at a given horizontal datum at depth, assuming a geothermal gradient of 30°C/km.

The above lateral variations taken together with vertical thermal gradients produce an instantaneous thermal structure in orogens consisting of isotherms gently inclined toward the foreland at angles greater than or equal to the surface slope (Barr and Dahlen, 1989; Hwang and Wang, 1993). These findings are consistent with the finite orientations of metamorphic isograds in some ancient mountain belts (e.g. Groshong et al., 1984).

Since many grain-scale deformation mechanisms are partially dependent on temperature, it follows that deformation fronts would mimic the contours of the isotherms to the degree that the deformation mechanisms they delineate are dependent on temperature. If isotherms are inclined toward the foreland, one might

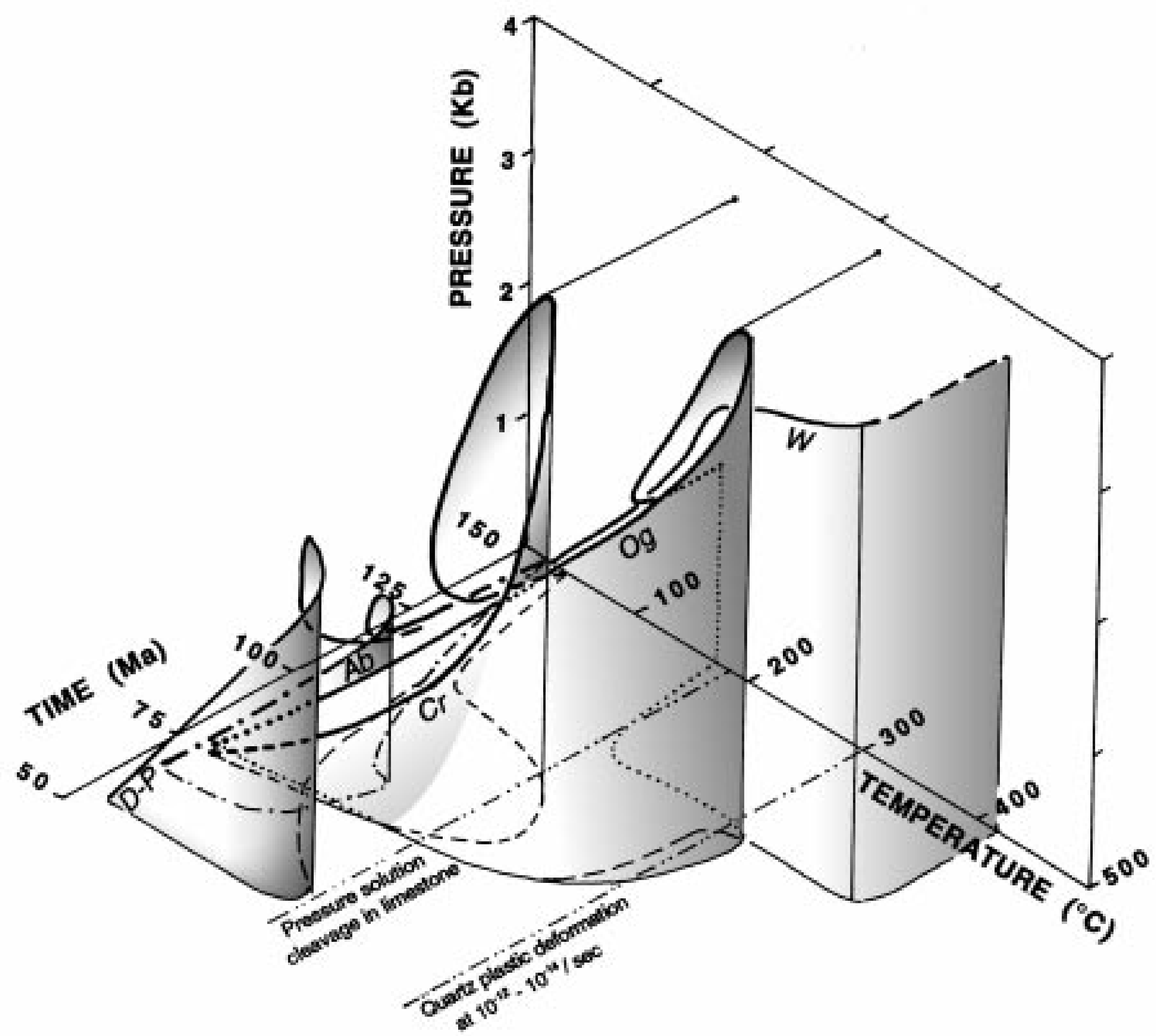

Fig. 2. Progressive thrusting in the external portion of the Sevier fold-and-thrust belt in the Idaho–Utah–Wyoming salient. The spiral *P–T–t* paths of specific locations within individual thrust sheets point to the spatial and temporal changes in deformation conditions within the fold-and-thrust belt. Threshold conditions for certain deformation mechanisms are only met at certain times within individual thrust sheets. Pressure solution cleavage in limestones is developed in different thrust sheets at different times, and time-transgressively toward the foreland; some sheets never reach threshold conditions and have poorly developed cleavage. Conditions for the development of penetrative foliation in quartzites are also reached at different times in different sheets, and time-transgressively toward the foreland. The thrust sheets shown are *W*: Willard, *Og*: Ogden, *Cr*: Crawford, *Ab*: Absaroka, and *D-P*: Darby-Prospect.

expect deformation fronts to share that general orientation (Fig. 1).

3. Deformation front migration

Since most mountain belts widen progressively as they evolve, it follows that deformation fronts should spatially migrate with time. Many previous workers have documented temporal changes in the location of thrust fronts in contractional mountain belts. The relative and absolute age relationships between thrust faults is based on the ages of synorogenic sediment preserved in the belts, geometric arguments and geochronology (e.g. Armstrong and Oriel, 1965; Woodward and Beets, 1988; DeCelles, 1994). Thrust wedges grow toward the foreland with time, with episodes of thrust reactivation and nucleation near the rear of the wedge to maintain critical taper (DeCelles and Mitra, 1997). Similarly, the orogenic wedge is most likely to begin the first increment of bulk-rock deformation at the rear of the wedge and deformation is time transgressive across the belt (Gray and Foster, 1997).

A single orogeny can produce a complex array of overprinting structures resulting in numerous deformation fronts. Based on a detailed study of spatial and temporal variations in progressive deformation in the

Valley and Ridge Province of Pennsylvania, Gray and Mitra (1993) recognized five regionally significant stages of progressive deformation and were able to demonstrate that at least two of the stages of progressive deformation were time-transgressive across the thrust belt. This finding is corroborated by paleomagnetic evidence of time-transgressive folding across the belt (Stamatakos et al., 1996). Both datasets demonstrate that deformation fronts migrated forelandward with time during the Alleghanian orogeny. Others have been able to demonstrate the migration of deformation fronts in other belts (e.g. Sevier fold-and-thrust belt, Mitra and Yonkee, 1985; Colorado Plateau, Nickelsen, 1993; Helvetic Alps, Groshong et al., 1984; Lachlan Fold Belt, Collins and Vernon, 1992, Gray and Foster, 1997).

Depending on the threshold environmental and rock parameters that are required for a deformation mechanism to activate, some deformation fronts may migrate into their final position (e.g. Groshong et al., 1984) while others may develop at a fixed location with respect to the autochthonous rocks in mountain belts. The latter type of deformation front may be rare relative to the former type and may form under a very restricted set of rock and environmental parameters, such that the conditions for this mechanism to operate are met at only one location in the belt during the history of the orogeny. These special conditions might be met at stress risers such as basement highs over which allochthonous rocks are translated, at footwall ramps or in folded rocks with rolling hinges. The rocks might experience localized, temporary changes in stress and other conditions as they are translated over these features.

It is perhaps easier to conceptualize deformation fronts that spatially migrate over time as the orogen evolves and the threshold conditions for a deformation mechanism to operate are met at different locations in the orogen through time. Fig. 2 shows an example of how P,T conditions change with time in successive thrust sheets in the external portion of the Sevier fold-and-thrust belt in Idaho–Utah–Wyoming. Each point within a thrust sheet cycles through its own P–T–t loop, reaching threshold conditions for certain deformation mechanisms at certain times. For example, the conditions for the development of pressure solution cleavage in limestones are reached time-transgressively across the belt within individual thrust sheets as they are emplaced.

Deformation fronts move because of stress, thermal and other gradients in orogenic wedges and changes in environmental parameters within the wedge with time (den Tex, 1963). For example, prograde metamorphism or changes in convergence rates in a convergent mountain belt may drive deformation front migration (e.g. Collins and Vernon, 1992). Gross mechanical

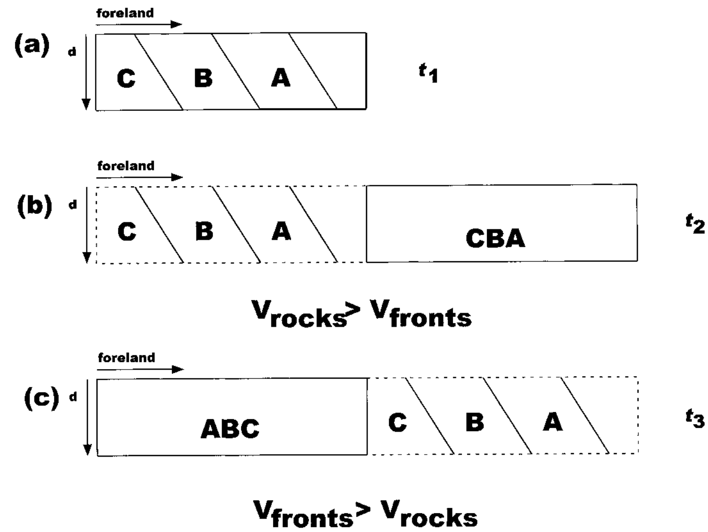

Fig. 3. A simple cross-section illustration of complications in progressive deformation histories produced by the relative velocity of deformation fronts and rocks. Cross-sections are drawn normal to the structural grain of the orogen. (a) Three different families of structures (Stages A, B, and C) are operational in different parts of the belt at t_1. One area farthest toward the foreland remains undeformed. (b) In the case where the fronts migrate more slowly than the rocks are being translated forelandward or where allochthonous rocks are translated through deformation fronts which remain in a fixed position with respect to authochthonous rocks, C is the oldest stage and A is the youngest stage recorded in the allochthonous rocks. (c) In the case where deformation fronts migrate faster than the allochthonous rocks are translated, Stage A will be the oldest stage of deformation and C will be the youngest stage recorded in the rocks that have undergone all three stages of deformation.

instabilities develop in a wedge that exceeds critical taper and as the wedge grows to decrease taper, deformation fronts can be expected to propagate farther out into the wedge. Finally, the wedge changes material properties as it deforms (DeCelles and Mitra, 1997). This may cause strain hardening or softening and the subsequent migration of deformation fronts.

On a mountain belt scale, differential stress distributions are controlled in large part by basal and topographic slope (Davis et al., 1983). In belts such as the central Appalachians, the sole thrust (near the basement–cover contact) is inclined toward the hinterland (internal belt) and the paleotectonic surface slope is unknown but can be assumed to have been inclined toward the foreland (external belt) based on consistent stratigraphic coarsening and thickening toward the hinterland. This geometry is commonly observed in convergent orogens and models predict that shear and normal stress within the wedge will decrease toward the foreland in the wedge (e.g. Hafner, 1951). Liu and Ranalli (1992) have shown that within the limits of the boundary conditions imposed on their model, the state of stress differs with position on the sole thrust. This results in a 'dislocation-type mechanism' for displacements on discrete patches of the fault and provides an additional reason for an uneven stress distribution within the wedge (Price, 1988; Liu and Ranalli, 1992).

A single orogeny can continuously deform a wedge by either increasing the size of the deforming wedge through time by growth at the toe of the wedge (e.g. Davis et al., 1983), deforming a wedge of constant size, or reducing the size of the actively deforming wedge through time as the orogeny wanes. In fact, all of these phenomena may be repeated cyclically throughout the growth of a single Coulomb wedge as it cycles through periods of supercritical, critical and sub-critical development (respectively) (DeCelles and Mitra, 1995). A critical or supercritical wedge (or orogen of constant or expanding size) may deform by propagating toward the foreland and as it does so, deformation fronts can be expected to migrate forelandward (Fig. 2; e.g. Mitra and Yonkee, 1985). When the wedge becomes sub-critical (or the active orogen contracts), the hinterland of the wedge undergoes internal deformation and the stages of deformation may be out-of-sequence as a result (e.g. 'OOSTs' of Morley, 1988).

Whatever the cause of migration of deformation fronts, their signature sequence of progressive deformation should be distinguishable from deformation fronts that are fixed as rocks are translated forelandward (Fig. 3). As rocks are translated toward the foreland through fixed deformation fronts, they should record an overprinting sequence beginning with deformation that occurred in the hinterland (Fig. 3b). On the other hand, if deformation fronts migrate at a faster rate than the rocks, the opposite sequence of progressive deformation will be recorded in the rocks (Fig. 3c).

4. Deformation front migration rates

Orogenic wedges generally increase in size with time and as they do, distinctive stages of deformation will be diachronous and migrate forelandward with time (e.g. cleavage formation in limestones of the Sevier fold-and-thrust belt; Fig. 2). This forelandward propagation of deformation fronts is analogous to forelandward thrust front propagation predicted by critical wedge theory and documented by field studies. The rate of migration of a deformation front is independent of the rate at which deformation proceeds on either side of it; the front simply serves as a switch that turns on and off the mechanisms that are active on either side of the front. Cleavage development in the Robertson Mountain Terrane in Northern Victoria Land, Antarctica was diachronous and the cleavage front migrated away from the core of the orogen at 0.4 cm/y, a rate that may be suggestive of the convergence rate (Dallmeyer and Wright, 1992). However, convergence rates do not directly control grain-scale flow laws. Ultimately, the rheology of the orogenic wedge may control the migration rates of deformation fronts. This rheology is dependent not only on its changing surface and body forces and thermal structure but also on its intrinsic mechanical properties.

The rate at which deformation fronts migrate is less well documented than the strain rates which define the mechanisms on either side of the front. Sometimes deformation initially lags behind the onset of prograde metamorphism, only to outpace the thermal pulse as it migrates forelandward (den Tex, 1963; Schuiling, 1963; Collins and Vernon, 1992). The result is a classic display of pre-kinematic metamorphic textures in the core of the belt and syn- or post-kinematic metamorphic textures in the external parts of the belt. This phenomenon illustrates that while operation of deformation mechanisms may be strongly temperature controlled (Groshong, 1988), the migration of deformation fronts may be controlled by parameters other than temperature (e.g. differential stress).

Although we have discussed some allogenic causes (stress distributions, thermal structure, changes in heat flow, climate (affecting erosion rates), and convergence rates with time) for deformation front migration, there are several autogenic features in mountain belts that may also produce rate changes in deformation front migration and subsequently affect the final structure of the mountain belt.

Once a body of rock has undergone a particular stage of deformation, its susceptibility to other deformation mechanisms may be modified (either enhanced or diminished) or in some cases, remain the same. Strain softening or hardening may occur as grains become deformed. For instance, in a coarse-grained body of rock, the first increment of deformation may involve elastico–frictional processes (perhaps forming small fault arrays, e.g. Wojtal, 1986) while deformation by pressure solution (perhaps involving the development of a cleavage) would not be favored because of the large grain size (Rutter, 1976; White, 1976; Wojtal and Mitra, 1986). As grain size is reduced, Mohr–Coulomb failure is increasingly difficult and pressure solution may become the dominant, rate-controlling mechanism of deformation (Mitra, 1984, 1994).

This example illustrates the dependence of subsequent stages of deformation on the development of previous stages. In a mountain belt, it would be possible to locate the present deformation front between the rocks that had undergone only the cataclastic deformation and those that had undergone additional pressure solution. It is tempting, then, to make an argument for time–space equivalence. However, these deformation mechanisms are dependent on different parameters. Crack propagation stress is inversely related to grain size; Mohr–Coulomb fracture is directly related to differential stress and pore fluid pressure and inversely related to lithostatic pressure. Pressure solution is strongly dependent on temperature, grain size and a variety of fluid characteristics (e.g. flux, solubility limit). It is certainly reasonable to expect that if permeabilities varied laterally, the sequence of structural stages might have been different. In addition, foreland basins are often characterized by fining toward the foreland. All other variables held constant, lateral grain-fining would enhance pressure solution rates toward the foreland since pressure solution rates depend inversely upon grain diameter.

Reaction softening or hardening either enhances or inhibits strain rates as the rocks undergo metamorphic reactions (Mitra, 1978; Gillotti, 1992). Holland and Lambert (1969) suggested that mineral transformations that occur with prograde metamorphism cause both increases and decreases in strain rates with time. An example of reaction softening would be the reduction of potassium feldspar grains to weaker micas or clay minerals (e.g. Mitra, 1978). Finally, rocks undergo geometric softening/hardening as the mechanical layering and fabric anisotropies are modified or reoriented with respect to the principal stress trajectories in the belt. Rocks become weaker (i.e. geometric softening) as mechanical fabrics become aligned parallel to orientations of high shear (Knipe, 1989). This is true at the grain scale and may also be true at the scale of the mountain belt. The roughly horizontal stratigraphy of the pre-orogenic miogeocline–shelf sequence represents a major anisotropy that becomes reoriented as the strata are folded. As folds become more pronounced (smaller interlimb angles, higher amplitudes, reduced wavelengths), thrust sheets are expected to undergo geometric hardening.

All of these strain hardening and strain softening effects associated with progressive deformation will influence the rates at which deformation fronts migrate within the mountain belt. Strain and reaction softening/hardening are, not surprisingly, dependent on the lithologies involved in the deformation. Saw-toothed deformation fronts may dramatically change shape as strain hardening/softening effects take place in different units. This possibility highlights the need to distinguish between instantaneous and finite geometries of deformation fronts.

In addition, successive fronts may begin to converge, diverge or even pass one another as the material properties of rocks are changed as a result of deformation. This is one reason why it is inappropriate to assume that the spatial distribution of deformation fronts is related to the relative ages or degrees of deformation in a mountain belt. The final mapped configuration of deformation fronts may consist of crossing, anastomosing or otherwise linear but utterly out-of-sequence deformation fronts. In an area which has experienced

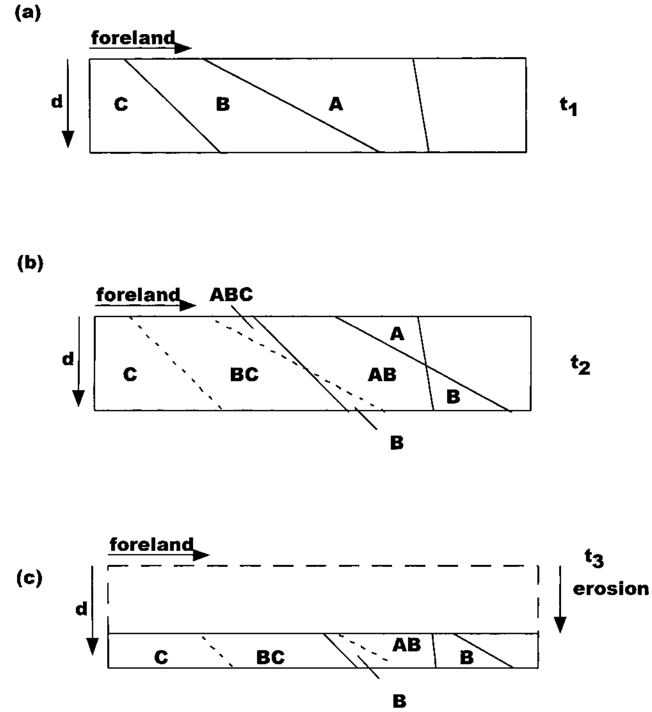

Fig. 4. A simple cross-section illustration of complications in progressive deformation histories produced by deformation fronts which migrate at different rates. Cross-sections are drawn normal to the structural grain of the orogen. (a) Three different families of structures (Stages A, B, and C) are operational in different parts of the belt at t_1. One area farthest toward the foreland remains undeformed. These families of structures (which define discrete structural stages) are bounded by deformation fronts of different slopes depending on the degree of dependence of rate-controlling deformation mechanisms with change in rock and environmental parameters with depth. (b) At t_2, the deformation fronts that mark the onset of Stages B and C have migrated forelandward with respect to the deformation front that marks the onset of Stage A. The result is a rich spectrum of deformation histories at different positions in the belt. This simple illustration suggests that progressive deformation histories are strongly position-dependent and that one may expect that deformation histories may be incomplete at many outcrops and/or significantly different in different parts of the same belt. Note that while it is possible to produce out of sequence (e.g. B then A) progressive deformation in the case of deformation fronts that pass each other as they migrate at different rates, it is not possible to produce sequence reversals if one of the deformation fronts is fixed and does not migrate (as illustrated in this figure). (c) Progressive deformation histories may be significantly different at depth or at different erosion levels. Compare this pattern with that at the surface level of the cross-section in (b).

passing, forelandward migrating deformation fronts delimiting different deformation mechanism associations A and B, the hinterland area may record a progressive deformation history of A followed by B and the forelandward area would record a temporal sequence of B followed by A. Clearly, space–time equivalence fails in this instance.

Consider a mountain belt simultaneously experiencing three different stages of deformation: A, B, and C (Fig. 4a). If stages B and C migrate forelandward at the same rate and stage A does not spatially migrate, or migrates at a slower rate relative to B and C, a complicated array of different progressive deformation histories should be recorded at different positions both laterally and with depth in the belt (Fig. 4b and c). A 'complete' deformation history (A then B then C) is observed in only one area in the middle of the belt. It is possible that this kind of rich diversity of structural expression has been misinterpreted in the past or may have been passed off as the result of incomplete or poor exposures rather than the simple result of migrating deformation fronts.

In the event that one stage of deformation is dependent upon the preceding stage of deformation, it is likely that while the second deformation front may begin to migrate at a faster rate and actually catch up with the first deformation front, the second front would be unable to pass through the first deformation front. Indeed, several fronts may all be dependent on one preceding stage of deformation and the end result may be a pile up of deformation fronts in a mountain belt. Such an area would be characterized by rock exhibiting a complex multi-stage deformation history on one side of the front and relatively undeformed rocks with a less complex progressive deformation history in detail on the other side.

5. Conclusions

This paper has attempted to address the impact of the processes involved in progressive deformation on the distribution of structures in a mountain belt. Deformation fronts of all kinds (cleavage fronts, thrust fronts, fold fronts, etc.) are probably not static, permanent markers of the limit of deformation in mountain belts. Rather, they are three-dimensional surfaces, probably highly irregular in detail, which probably had a long and complex history of formation and relative migration during the evolution of the mountain belt. Deformation fronts are most likely inclined toward the foreland due to the thermal structure of mountain belts and undergo morphologic changes as they migrate relative to the rocks. Some deformation fronts may remain fixed relative to autochthonous rocks. The earliest stages of development of a convergent orogen involves prograde metamorphism, development of topography and applied surface and body forces within the orogenic wedge that may drive successive deformation fronts forelandward with time. Deformation fronts migrate at rates controlled by external factors such as convergence rates but are also controlled by the flow laws that govern the constituent deformation mechanisms associated with each front. The intrinsic character of the wedge as well as the changes in rock parameters that take place during deformation may also place controls on the migration and migration rate changes of deformation fronts. It is therefore probable that deformation fronts migrate at varying rates through time and that they may converge, diverge or pass each other as they migrate at different rates. Because certain stages of deformation are dependent on previous stages of deformation, it is also conceivable that deformation fronts would pile up at the location of the controlling deformation front in the mountain belt. The end result would be a dramatic structural boundary between less deformed, less progressed rocks on one side and more structurally evolved rocks on the other side.

Incorporating these ideas on deformation fronts into the study of mountain belts may change the way in which orogens are perceived. Exploration of mountain belts may be enhanced by the knowledge that the mapped limit of a certain deformation front need not lie close to the limit of deformation at depth (even within the same formation). The expectation that deformation fronts are highly irregular in detail and may be gently inclined demands that assumptions about the aerial extent of deformation in mountain belts be reexamined at depth. Our work highlights the importance of independently establishing the sequence of progressive deformation at every outcrop examined in a regional structural study. Variations in the rates of migration of deformation fronts may ultimately result in a rock record that contains the opposite or at least different sequence of structural stages from that observed in other areas of the region.

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