DOE/ER/14886-1

Final Report

To

DEPARTMENT OF ENERGY

**Mechanical Models of Fault-Related
Folding**

DOE Grant No. DE FG02 98ER14886

by

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PURDUE UNIVERSITY

11/14/2002

DOE Patent Clearance Granted

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3-15-04

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Final Report

to

**Geosciences Research Program
Acquisition & Assistance Management Division
Office of Energy Research
ER-64/GTN**

Department of Energy

Washington, D.C., 20585

Mechanical Models of Fault-Related Folding

DOE Grant No. DE FG02 98ER14886

November 2002

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47907-1397

DESCRIPTION/ABSTRACT

DEPARTMENT OF ENERGY GRANT NO. DE FG02 98ER14886

The subject of the proposed research is fault-related folding and ground deformation. The results are relevant to oil-producing structures throughout the world, to understanding of damage that has been observed along and near earthquake ruptures, and to earthquake-producing structures in California and other tectonically-active areas. The objectives of the proposed research were to provide both a unified, mechanical infrastructure for studies of fault-related folding and to present the results in computer programs that have graphical users interfaces (*GUIs*) so that structural geologists and geophysicists can model a wide variety of fault-related folds (*FaRFs*).

The objectives were ambitious, but we have made considerable progress: We have developed fault-related folding theory in four areas: (1) folds associated with listric and antilistric faults and with gang faults; (2) growth of fault-related folds that form adjacent to faults that propagate or are stationary; (3) forced folding above a rigid basement; and décollement-ramp folding in isotropic and anisotropic materials. We have written *GUIs* for each of the theoretical developments so that we and other people can use the theory to study fault-related folding processes. One *GUI*, *SeDeRa*, allows one to study folds produced by décollement ramps that are active sequentially. Another *GUI*, *SeGaLi*, will model sequential or simultaneous activation of three-dimensional dislocation/fault elements and the folds, strains and displacement fields produced by such activity. Another *GUI*, *FoFo*, was designed to allow one to simulate forced folds and tri-shear deformation patterns. Finally, two *GUI*'s have been written, so far, to study the growth of fault-related folds in a flowing medium in which the faults are represented not by dislocations but by boundary elements. We introduced our *GUI* research at the annual 2000 GSA meeting in Reno.

Our computer programs have been copyrighted by the Purdue Research Foundation. The documentation with the computer programs makes clear that the programs were developed with DOE funding. Karen White (kfwhite@purdue.edu), of the Purdue Research Foundation, has contacted management of many petroleum companies in order to inform them of the availability of the software. We provide trial versions of the software on a Purdue website: www.eas.purdue.edu/fauxpli, but the trial versions have several components that are nonfunctional. However, people can get the full versions of the software, free of charge, simply by contacting Karen White or me (gotesson@purdue.edu) and completing a use agreement. All the manuals are available for downloading at the *fauxpli* website. Many individuals at companies and universities have been provided with the full software packages. A list may be obtained from Karen White, who maintains the use agreements.

We completed three field projects as part of the research supported by DOE. One of the field projects was based on mapping and analysis of structures within the San Rafael Swell, Utah. The research led to nine manuscripts, three of which have been published in the *Journal of Structural Geology* and six of which, so far, have published only on our web site. Another field project focused on some fault-related folds in the Bighorn Basin of Wyoming. The project led to two manuscripts, published on our web site plus a poster session at the 2003 national GSA meeting in Seattle. The third field

PAPERS/ABSTRACTS/WEB DOCUMENTS

DEPARTMENT OF ENERGY GRANT NO. DE FG02 98ER14886

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- Johnson, K.M., 2000. Types of internal deformation in the San Rafael Monocline, Utah. 12 p.
www.eas.purdue.edu/physproc/small_structures.htm
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www.eas.purdue.edu/physproc/stress_states.htm
- Durdella, M.J., and Johnson, A.M., 2000. Geometry and folding mechanisms of the Pitchfork anticline, Bighorn Basin Wyoming. www.eas.purdue.edu/physproc/kink-wedge.htm
- Johnson, A.M., and M.J. Durdella, 2000. The kink-wedge fold. www.eas.purdue.edu/physproc/kink-wedge.htm
- Johnson, A.M., 2001, Yielding instability and propagation of kink bands. 50 p.
www.eas.purdue.edu/physproc/propagation_of_kinks.htm
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www.eas.purdue.edu/physproc/pdf%20Files/AtillaFaults.pdf
- Johnson, K.M., Manuszak, J.D., and Johnson, A.M., 2000. How Kink Bands Work and How to Work with Kink Bands. 21 p.
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- Johnson, K.M., and Manuszak, J.D., 2000. How to Analyze Kink Folds, 10 p.
www.eas.purdue.edu/physproc/easy_guide_to_kinks.htm
- Johnson, K.M., 2000. Some Comments on high-amplitude kink folding, 12 p.
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- Johnson, A.M., Johnson, K.M., Durdella, J., Sozen, M., and Gur, T., 2001. Main rupture and adjacent belt of right-lateral distortion, Kaynasli, Turkey 12 November 1999 Duzce earthquake.
www.eas.purdue.edu/physproc/turkey_viaduct_study.htm

project was carried out in conjunction with faculty and students in earthquake engineering at Purdue University and with faculty and students in earthquake engineering at the Middle East Technical University in Turkey. We mapped several stretches of the November 1999, Düzce earthquake rupture and measured the deformation of a large viaduct, the Kaynasli Viaduct of the Trans-European highway between Istanbul and Ankara, Turkey. The viaduct passed over the eastern end of the earthquake rupture, so it presented a wonderful opportunity to analyze ground deformation in the vicinity of the North Anatolian fault. Some of the results of this study have been recently published in the Journal of Seismology. The results were also presented at the NSF-Tubitak, Turkey-Taiwan Grantee Workshop, March 2002 in Antalya, Turkey. Many of the participants indicated that our results were among the more exciting presented at the workshop.

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SUMMARY OF THE RESEARCH PROPOSED IN 1998

Grant No. DE FG02 98ER14886

"MECHANICAL MODELS OF FAULT-RELATED FOLDING"

GEOSCIENCES RESEARCH PROGRAM

RELATIONSHIP OF GENERAL TOPIC TO DOE'S RESEARCH PROGRAM

In several ways the research funded by **DOE Grant No. DE FG02 98ER14886** is relevant to the stated goals of the Department of Energy Geosciences Research Program. The research is fundamental in nature, and is of long-term relevance to one, or more, energy technology or safety objectives of the Department of Energy, including three areas in Geology,

Geophysics and Earth Dynamics: Evolution of Geologic Structures (faults and fault-related structures).

Properties of Earth Materials (properties of rocks and consolidated alluvium being faulted).

Rock Flow, Fracture or Failure (response of soft alluvium, consolidated alluvium and rock to faulting).

The topic of the proposed research is fundamental to our understanding of the forms of fault-related folds. Many structural geologists working in government, universities and private industry use and develop geometric models that they believe to represent the ideal forms of fault-related folds. In general there is no theoretical basis for a geometric model, except that it produces a balanced cross section, and one can imagine a kinematic process that would allow the structure to grow. Although the geometric models have provided considerable guidance in reconstructing only partly observable folds at depth, they have been untested by theory so it is unclear whether they are actually viable. There is no theory extant that will describe the typical fault-related fold. A reason this is important to the Department of Energy is that the geometric models are used—perhaps exclusively—in the petroleum industry in order to extract petroleum. A second, perhaps more important purpose of the theoretical research is to predict forms of fault-related folds that are omitted from the geometric models. There are numerous theoretical examples where the major fault-related fold is described, roughly, by a geometric model, but where minor structures are missing. In the theoretical analysis, one reproduces, approximately, the major fault-related fold, but as a bonus one can observe a variety of minor structures. I presented an example in previous pages, where a blind thrust fault produced a minor anticline below the fault tip while forming the major anticline above the fault tip. It is these minor structures that will provide completely new targets in exploration and especially development of oil fields.

INTRODUCTION TO FAULT-RELATED FOLDING

Although early investigators, including J. W. Powell, recognized the existence of fault-related folds, during the last 20 years it has become clear to structural geologists that most folds in petroleum-producing areas are fault related. As a result, a whole series of geometric and kinematic models of fault-related folds, including fault-arrest and fault-propagation folds, fault-bend folds and décollement folds, have been proposed to describe the shape or formation of fault-related folds. A problem is that, although the geometric models certainly describe important fold forms, they cannot explain uniquely how the forms develop. Another problem is that the geometric models cannot predict new fold forms or minor fold forms associated with major ones. This, of course, is where theory shines. Theory programmed to reproduce the essential features of a major fault-related fold automatically balances cross sections, shows how minor folds are associated with the major fold. Also, theory can predict major folds that have remained obscure or unrecognized. But theory has not kept pace with the geometric analyses. To date, only the theory of ramp syncline and ramp anticline have been developed adequately to relate to actual folds.

In the proposed research, we will develop the theory. For folds that involve the terminations of detachment surfaces, faulting theory will be used, but for materials that flow. By fine-tuning folding theory we can more and more closely model the shapes of folds and the conditions responsible for their formation. For folds that involve long detachment surfaces and ramps, folding theory of layers will be used. In this way we can use theory to predict fold forms. Such knowledge is important in the exploration for petroleum resources but especially for the development of petroleum in secondary, only partly visible structures.

The goal of the proposed research is to provide a unified, mechanical infrastructure for studies of fault-related folding. The mechanical theory will be presented via programs that have graphical users interfaces (GUI) so that structural geologists may model a wide variety of fault-related folds. The research will develop the algorithms behind the software, but the software will allow structural geologists to solve problems themselves. When the infrastructure is in place we will be able to determine the causes of the different forms of fault-related structures.

In a Penrose conference on fault-related folding five years ago, some 82 participants from throughout the world gathered to discuss fault-related folding and balancing of cross sections. Of these, 23 papers from the conference were published in a special issue of *Journal of Structural Geology* (Anastasio and others, 1997). A reason for the interest is that fault-related folding has diverse applications such as seismic-hazard evaluation and petroleum exploration and development. Restoration of cross sections through a balancing procedure is important because it places geometric constraints on interpretations of the structures. Also, if one could describe the process by means of which a structure formed, one could follow the evolution of the paleostructure and relate the evolution to the migration of oil and gas.

Although much of the interest in fault-related folding has developed during the last 20 years, the idea of fault-related folding began much earlier. John W. Powell (1873) recognized that monoclinial flexures are folds that form in front of a slipping fault, so that they are fault-propagation or fault-arrest folds. Fifty years later John L. Rich (1934) recognized that some folds in the Appalachians are a result of duplication of strata. His revolutionary concept was that anticlines formed where strata are repeated by low-angle faulting. The repetition occurs during thrusting, by movement of the thrust block from a lower décollement to a higher décollement through a diagonal connecting ramp. Thus was born the notion of the ramp anticline, one form of which is called the fault-bend fold. The ramp anticline, in various forms, has come to dominate analyses of subsurface structures in fold/thrust belts throughout the world. Prior to 1934, most folds were interpreted to be a result of layer-parallel shortening and buckling, or of density instability. Now most folds are considered to be fault-related.

- ? An important part of the research proposed here is to develop fault-related folding theory that can describe the essential features of the folds observed in seismic profiles.

The infrastructure of descriptive knowledge of fault-related folding in the intervening years has been due largely to Dahlstrom (1969, 1970); Wernicke and Burchfiel (1982); Bally (1983); Stone (1983a, 1983b, 1983c, 1985, 1995); von Heune (1986); von Heune and others (1983); and Wilson and Shumaker (1992). By far the most important baseline descriptions of fault-related folds in sedimentary rocks, and in some cases in underlying basement rocks, have been the published seismic profiles of structures in oil fields (Bally, 1983). There has been a virtual flood of such publications in the petroleum geology literature since the classic compilation by Bally, who engineered collaboration among a group of oil-company geologists, consultants, government geologists and academics in the production of a four-volume folio on interpretation of seismic profiles in a host of structural settings throughout the world.

METHODS OF STUDYING FAULT-RELATED FOLDING

Dahlstrom (1970, p. 333) demonstrated the value of balanced cross-sections and contributed in other important ways to geometric analyses of folding. His approach was to classify structural types in a fold/thrust belt and to define geometric rules that govern their behavior. He focused on the geometric level, carefully differentiating between what can be observed to exist and what can reasonably be inferred to exist. He stated that, "...[In this way] one establishes the models that the academic must attempt to explain and the explorationist must attempt to exploit."

According to Anastasio and others (1997) the topics in fault-related folding of current interest to structural geologists fall into five broad categories:

- ? Fault-bend folding
- ? Fault-propagation folding
- ? Fault-arrest folding
- ? Break-thrust folding
- ? Décollement folding

Geometric and Kinematic Rules

As the landmark studies cited above demonstrated the importance of fault-related folding in oil fields, researchers with quite different approaches began to investigate such folds. In one group of approaches, geometric rules are introduced in order to constrain possible fold forms at depth with limited surface and subsurface information. Thus, one makes a sufficient number of assumptions about possible geometry and then one applies those assumptions in order to describe folds (e.g., Suppe and Namson, 1979; Suppe, 1983, 1985; Woodward, Boyer, and Suppe, 1985; Mitra, 1986, 1992, 1993; Groshong, 1989; Epard and Groshong, 1993; Storti and Salvini, 1996 and Zapata and Almendinger, 1996). This practice began with folds shown by straight limbs at the beginning of the century (Holland, 1914), continued with Busk's (1929) method of drawing folds with circular arcs. The practice of showing fold limbs with straight lines was reintroduced through Suppe's fault-bend fold, and continues to dominate geometric models of fault-related folds today.

Figure 1. From McConnell and others (1997, fig. 1).

For each type of folding, a kinematic model is appended to the geometric model in order to present what is supposed to be a credible mechanism for formation of that type of folding. For example, Hedlund (1997) indicates that the *fault-bend fold* is characterized by a fault that pre-exists the displacement of the block (Figure 1a). Folding is due to repetition of strata over the ramp. The kinematics and geometries of the structure range widely depending on whether bending is concentrated in tight hinges (Figure 1a), as in the Suppe model (Suppe, 1983; Jamison 1987; Johnson and Berger, 1989), or is distributed throughout the fold (Rich, 1934; Elliott, 1976; Berger and Johnson, 1980; Johnson and Berger, 1989). Regardless of form, though, the empirical and theoretical models assume that the overburden rock migrates through spatially-fixed axial surfaces (the velocity discontinuities of Johnson and Berger, 1989).

- ? One of the goals of the proposed research is to determine theoretically which physical conditions or processes are responsible for fixed-hinge as compared to migrating-hinge folding.

A second type of fault-related fold is the *fault-propagation fold*, a fold produced by compression at the tip of a propagating fault. The original model (Suppe and Medwedeff, 1984, 1990) features angular folding at axial surfaces which move with the propagating ramp fault (Figure 1b). Fault-propagation folds thus grow by migration of axial surfaces. Such folds have been inferred at the leading edges of thrusts (e.g., Zapata and Almendinger, 1996). There have been several variations on the theme of fault-propagation folding. Jamison (1987) modified the model to include a component of forelimb thinning or thickening (Figures 1b). Chester and Chester (1990) adapted the fault-propagation fold model to folds formed above a planar detachment with a small segment of a growing ramp in its toe (Figure 1c).

Figure 2. Décollement fold according to Hedlund (1997, fig. 8)

A third type of fault-related fold is the *décollement fold* (Figures 2, 3b, 4 and 6C), applied to examples in the Appalachians by Berger and Johnson (1980) and in the Jura by Laubscher (1976). Décollement or detachment folds form by differential displacement and shortening of the hanging wall above a detachment surface. Décollement folds generally are considered to be fixed-hinge folds (e.g., Dixon and Liu, 1992), although Anastasio and others (1997) and Woodward (1997) suggest two end-member models that operate sequentially.

According to Poblet and others (1997), *décollement folding* occurs where the displacement related to layer-parallel blind thrusting is accommodated by folding in the hanging wall. Three kinematic models have been proposed for décollement folds where a homogeneous stiff layer is detached over a ductile unit (Figure 3b). According to one model (Figure 4a), the anticlines form with variable dip and constant limb length, so the hinges are fixed (DeSitter, 1956). In this model, folding and shortening occur mainly by limb rotation. The axial surfaces rotate to maintain constant stratigraphic thickness and the folds decrease in width. In a model by Mitchell and Woodward (1988) (Figure 4b), the anticlines develop with constant limb dip and variable limb length, so the hinges migrate. These folds contrast sharply with those shown in Figure 4a. Here folds grow exclusively by axial surface migration. The folds nucleate as small structures with fixed geometry and the fold grows by axial migration to accommodate shortening. Fold width increases as shortening progresses. According to Dahlstrom (1990) (Figure 4c), the anticlines develop with variable limb dip and variable limb length, so the hinges migrate various amounts. According to Poblet and others (1997), these folds nucleate as narrow folds. Shortening is accommodated by outward migration of the axial surfaces and limb rotation. The axial surfaces rotate to maintain constant stratigraphic thickness.

Fig. 1. Thrust-related fold models (after Suppe (1985), Jamison (1987) and Willis (1893), respectively): (a) fault propagation fold, (b) detachment fold, (c) break-thrust fold.

Figure 3. Fault-related folds (Woodward, 1997, fig. 1).

Poblet and others (1997) show that, even though the geometry of the final folds is the same in Figure 4, the time histories of uplift, rates of limb rotation and rates of limb lengthening are quite different for the three models. In models 1 and 3 both uplift and limb dip increase rapidly in the first stages and then increase at a linear rate. In both models all axial surfaces progressively rotate with increasing shortening in order to maintain constant stratigraphic thicknesses. In model 2 the uplift of the décollement fold increases linearly and limb dip is constant as shortening progresses. Limb length increases linearly. Thus, because these three kinematic models, with different folding mechanisms, can produce identical fold shapes, the fold form is not diagnostic of the folding mechanism.

Fig. 1. Evolution of three different models for asymmetric kink detachment folds involving a homogeneous competent layer detached over a ductile unit. The models are calculated using the parameters given in Table 1 and all three kinematic models produce the same final fold shape. All the models are area balanced for both the competent and ductile units. In the case of Models 1 and 2 variations in the depth to detachment occur in order to area balance the ductile unit (Homza and Wallace, 1995).

Figure 4. Three kinematic models of the evolution of a décollement fold (Poblet and others, 1997, fig. 1).

Cooper and Trayner (1986) and Woodward (1997) have noted that the prevalence of footwall synclines in foreland provinces suggests that décollement folding (Figure 3b) occurred prior to ramp propagation (Figure 3c) and subsequent to fault-bend folding. The fault-related fold is termed the *break-thrust fold* by Woodward (1997). According to his model the initial folding has fixed hinges, due to buckling (Figure 5a). Deformation is limited to the forelimb area. The trailing synclinal axis is fixed beneath the detachment but material is displaced through it, so the axis migrates relative to the folding material. Ramping of a thrust up from the detachment (Figure 5e) becomes the shortening mechanism when the forelimb has a dip of about 30 degrees. Then a fault breaks through a limb along a propagating ramp fault. After breaking through, the fold continues to grow, but now by fault-bend folding.

Woodward's (1997) model of a break-thrust fold.

Figure 5. (From Woodward, 1997, fig. 7). Stepwise evolutionary model of an asymmetric, predominantly fixed-hinge, break-thrust fold. Line lengths and fold limb dips are measured on topmost key bed. Lowest folded surface with 5° limb dips is preserved as the base of deformation throughout the folding. Synclinal hinges behind and in front of fold remain fixed in position and geometry. Layering that passed through trailing synclinal hinge into the backlimb of the anticline is shown with the cross-hatched area. A. and B. Fixed-hinge symmetric fold. C. and D. Inward migrating hinge. E. and F. Fixed-hinge asymmetric fold. Thrust ramp begins to grow. G. Final fixed-hinge, asymmetric fold. H. More likely scenario, showing rounded corners.

Thorbjørnsen and Dunne (1997) carefully described the form and analyzed internal structures of the Barclay anticline, part of the Browns Mountain anticlinorium in southeastern West Virginia, in order to determine which of the prevailing fault-related folding mechanisms might have been active as the fold formed. The geometric data suggested that the fold could be a fault-bend fold, a fault-arrest fold or a break-thrust fold. In order to further evaluate the mechanisms they determined strains in representative rocks. The strains did not fit any of the models, indeed, the strains were relatively uniform and apparently occurred before folding. The strains were accommodated at the microscopic scale by dislocation climb and by grain-to-grain dissolution and stylolitic dissolution features. They concluded that fold growth was achieved mainly at the mesoscopic scale, via layer-parallel slip, and that mostly in the forelimb of the fold. The back limb of the fold contains few slip surfaces. Contraction faults occur only at in the forelimb. They indicate that (op cit., p. 313), "This mesostructural distribution precludes material transport through the hinge from the forelimb. Fault-bend and fault-arrest models, therefore, cannot be used to explain the origin of the Barclay anticline." They suggest that (op cit., p. 315) "bed-parallel slip, perhaps during kinking, with negligible bending deformation in the hinge may be explained by break-thrusting." In this interpretation, the folding preceded thrust propagation and was by layer-parallel slip in the

forelimb, as in a kink band (Johnson, 1977; Pfaff and Johnson, 1989; Johnson and Fletcher, 1994; Stewart and Alvarez, 1991).

Thorbjomsen's and Dunne's (1997) models of thrust-related folding.

Figure 6. Classification of fault-related folds in terms of their timing with respect to fault propagation. A. In the break-thrust of Willis (1893), the fold precedes the faulting; the fault breaks through a limb of the fold. B. Fault-propagation folding was developed to describe folds that form during propagation of imbricate faults. C. Detachment folds (Laubscher, 1976) are either folds that amplify as displacement progresses or buckle folds that develop between bounding décollements. D. The fault-bend fold (Rich, 1934; Berger and Johnson, 1980; Suppe, 1983). In the fault-bend fold, the thrust fault and the ramps form before there is significant folding. The geometry of folding is determined by the geometry of the faults. E. Another type of fault-related folding that follows formation of the fault is the fault-arrest fold (Armstrong and Bartley, 1993; Wichham, 1995; Wei, 1997).

Mechanical Modeling

Many structural geologists have noted a feature of geometric models, illustrated in Figure 4, that geometric constructions do not uniquely define the process responsible for the geometry of a fold (e.g., Elliott, 1983). More seriously, as indicated by Wickham and Moeckel (1997), "a geometric restoration or forward model may be formally admissible...but physically impossible." The geometrical approaches are unable to describe the rich variety of fault-related folds observed in seismic profiles, and are unable to predict structures accessory to the main fault-related fold. The problem of prediction is particularly serious in the petroleum industry. As expressed eloquently by Dahlstrom (1970, p. 33): "...There is a perverse but fundamental rule of exploration which decrees that enough data to provide a sound statistical basis for extrapolation by analogy is available only in areas which are so well known that there is no longer much need for that particular kind of data extension."

We can avoid the problem of physical impossibility of an idealized fold form and face the problem of predicting fold forms by mechanical analyses of various folding processes. Mechanical analysis of fault-related folding, though, has been rare. Most of the forms of fault-related folds shown in Figures 1-6 have been largely ignored in theoretical studies. The theory of ramp folding has been applied to thrust sheets by Elliott (1976), Wiltschko (1979), Berger and Johnson (1980, 1982), Johnson and Berger (1989), Kilsdonk and Fletcher (1989) and Johnson and Fletcher (1994). There is also some theoretical research on the formation of monoclines (Sanford, 1959; Reches and Johnson, 1978). Otherwise, the theory of fault-related folding has remained undeveloped.

The proposed research is designed to address a wide range of fault-related folds. Much of the proposed research will be devoted to development of mechanical analysis that appear to capture the essence of models shown geometrically in Figures 1 through 6.

A Field Example Cited in the Proposed Research:

In order to illustrate the mechanical approach, I will consider two examples of fault-related folds that appear to be either décollement folds or fault-propagation folds of the types shown in Figures 2, 3, 4, 5 and 6. First I will derive the theoretical velocity distributions in a homogeneous flowing medium containing a detachment surface of finite length.

A Theoretical Analysis of Idealized Décollement Folding

We can derive a solution for the displacements and stresses within an flowing medium in the vicinity of a terminating, décollement fault by means of some solutions available in the literature (Tada and others, 1985) for compressible, elastic materials. Without loss of generality, we have assumed that the material is incompressible and we convert the solution to flowing materials by merely replacing displacement by velocity and modulus of rigidity by coefficient of viscosity. The basic mechanical problem is a planar fault of length $2a$ in a very large mass of flowing material. The slip velocity at midlength on the fault is $v_x(\max)$, parallel to the fault. The slip velocities range from that value to zero at each end. As a result of the slip, stresses are generated in the vicinity of the fault. We solve the problem for a slipped fault in a very large body subjected to shear stress, τ_{xy} , far from the fault. There will be normal traction, τ_n , and a shear traction, τ_s , across the fault. The shear traction is the resistance of the fault to slippage. The solution for the velocity components is,

$$v_x = v_x(\max) [\text{Im}\{\sqrt{(z/a)^2 - 1}\} + (y/a) \text{Re}\{\frac{1}{\sqrt{1-(a/z)^2}}\}] \quad (1a)$$

$$v_y = -v_x(\max) [(y/a) \text{Im}\{\frac{1}{\sqrt{1-(a/z)^2}}\}] \quad (1b)$$

in which, x is parallel and y is normal to the long dimension of the fault of length $2a$, $\text{Im}\{\}$ is the imaginary part of a complex expression and $\text{Re}\{\}$ is the real part, and z is the imaginary coordinate

$$z = x + iy \quad (1c)$$

η is coefficient of viscosity τ_{xy} is the remote, or far-field, shear stress, τ_s is the shear strength across the fault, and the maximum velocity is at $x=0$:

$$v_x(\max) = (\tau_{xy}/2\eta) [1 - \frac{\tau_s}{\tau_{xy}}] a \quad (1d)$$

I generally use a fourth-order Runge-Kutta solution (e.g., Johnson and Pfaff, 1989) to integrate for the displacements of points. In the problem illustrated here, there are passive layers that start out planar and then are deformed into arcs. Using that approach, I have calculated the successive forms shown in Figure 7.

As expected (e.g., Berger and Johnson, 1983), an anticline forms over the right-hand end and a syncline beneath the left-hand end of the fault. The anticline above the fault is expected because we know that the material in that position is subjected to fault-parallel compression. Likewise, material above the left-hand end of the fault is subjected to extension. We would expect structures such a kink fractures or tension gashes to form there (e.g., Cruikshank and others, 1991b).

Figure 7. Folds formed at terminations of double-ending décollement fault of length $2a$. γ is the shear strain. A. The medium is subjected to dextral shear. The displacement is largest near midlength along the fault and is zero at each end. B. As a result, in dextral shear an anticline forms at the right-hand end and a syncline at the left-hand end of the fault. C, D, and E, Folds formed with progressively larger angles of shearing. Shearing is shown by inclination of right- and left-hand sides of diagrams.

There is, though, another structure that will completely surprise most structural geologists, yet is potentially important to petroleum exploration and development. The presence of this structure illustrates the predictive capability of the theoretical analysis of a mechanism of fault-related folding. Furthermore, the structure necessarily occurs; it is not deduced intuitively. It is generally known to structural geologists working with seismic profiles that an anticline occurs with its culmination above the right-hand termination and a syncline beneath the left-hand termination for a planar, blind fault subjected to dextral shearing. It is known that the anticline provides an excellent target for petroleum exploration. It is not generally known, though, that a smaller, accessory anticline also forms beneath the blind fault but with a culmination somewhat behind the culmination of the anticline above the right-hand termination (Figure 7).

In Figure 7, therefore, we have an example of theoretical fault-related folds associated with a décollement in a very simple material. Let us compare its form with that deduced geometrically by Laubscher (1976, 1983). The detachment fold of Laubscher shown in Figures 3b and 5C, is stylized with straight backlimb and forelimb and with no deformation beneath the décollement. Although the idealized detachment folds (Figures 2 and 4) capture what one might call the 1st order phenomenon—an anticline above the fault at one end (and by inference) a syncline below the fault at the other—it completely misses, because it is assumed not to exist, the 2nd order phenomenon of a smaller accessory anticline beneath the larger one (Figure 7). Likewise, synclines occur both above and below the left termination for a fault with dextral slip.

The accessory anticline is an excellent target for exploration that has been predicted theoretically. This target for petroleum is generally unknown, partly because it is assumed to not exist (e.g., Figure 4) and partly because it can be difficult to see structures beneath faults in seismic profiles.

Cincinnati Décollement Fold

The analysis shown in Figure 7 was motivated by a field study of small folds in Cincinnati, Ohio by Hofmann (1966). According to Hofmann, the fold shown in Figure 8 is several meters wide and about a meter in amplitude. It formed near a deeply incised stream valley west of Cincinnati, near Miamitown, Ohio. The fold is in the Kope Formation which, here, consists of a few limestone beds on the order of centimeters thick interbedded with thicker shale beds. The shale itself consists of extremely thin laminations. This fold, and many others similar to it, apparently formed by localized bedding-plane slippage that was induced by lateral unloading of the rocks above the level of the bottom of the valley. Apparently a bedding-plane fault begins somewhere to the left of the view in Figure 8 and terminates roughly midway across the view.

Figure 8. A small fault-related fold in shale containing a few limestone beds, in the Kope Formation, Cincinnati, Ohio (From Hofmann, 1966; figure has been reversed).

In order to understand the essential features of this fold, we assume that the analysis of a planar blind thrust (Figure 7) is relevant and that the natural fault did not grow during formation of the fold, it merely slipped with a fixed termination. Thus the fault is a flat fault-arrest fold. In this case our analysis shows how a few bedding planes above and below the fault should deform.

We note that there is a remarkable similarity between the gross forms of the deformed subhorizontal lines in the theoretical model and the bedding planes in the field example. In both, the slip was right-lateral and the deformations were much too large for the materials to have behaved as linear elastic materials. In the theoretical model, the layers above the fault plane, at the fault termination, clearly form an isolated anticline whose width increases upward from the termination of the fault. The anticline has a steep limb on the right and a shallow limb on the left, just as in the geometric model (Figure 4). The natural fold has these same features.

Another striking feature of the theoretical model is that the anticline, though diminished, appears below the termination of the fault as well above and that the crest of the anticline below is behind the crest of the anticline above. In the theoretical model, the low anticline below is behind the tip of the blind fault. We see the same relation between the larger anticline above and the lower anticline below the fault in the field example. We cannot easily determine in the field example, though, exactly where the termination of the bedding-plane fault is.

Although the theoretical anticline is asymmetric, with a short, right-facing limb and a long left-facing limb, and the field example has the same sense of asymmetry, we should not assume that we have thus explained the asymmetry of the fold. The relatively straight limbs of the Cincinnati fold suggest that the fold formed by bedding-plane slip and that there was some contact strength that resisted slippage (Honea and Johnson, 1976). Furthermore, the widening of the kink-like fold in the shales upward suggests a fold with a triangular shear zone, as recognized by Erslev (1991) and Erslev and Mayborn (1997).

Thus, by combining the relatively simple theory of detachment folding in a flowing material, with our understanding of the formation of kink bands, we can understand rather thoroughly the conditions under which the Cincinnati folds formed. The termination of the detachment produced the necessary compression and the dextral shear relative to layering of shale, and the strength of contacts produced the straight limbs (Honea and Johnson, 1976; Reches and Johnson, 1976; Johnson, 1977; Pfaff and Johnson, 1989).

Although we can explain many of the gross features of the Cincinnati detachment fold with the simple viscous model, a comparison of the theoretical model, the geometric model and the field example indicates that there are several features of the field example that are missing in the fold produced with the theoretical model. The goal of the research will be not only to explain the natural example at Cincinnati, but to explore the controls of various parts of the geometric model (Figure 4). In the proposed research, I intend to address these missing features:

1. THE AMPLITUDE OF FOLDING APPEARS TO INCREASE UPWARD IN THE FIELD EXAMPLE WHEREAS IT DECREASES UPWARD IN THE THEORETICAL (FIGURE 7) AND GEOMETRIC MODELS (FIGURE 4).

A serious limitation of the theoretical model is that it contains no physical layers. The layers are defined by passive markers, not by contacts that can slip or by contrasts in stiffness. Passive layers do not bend like mechanical layers. I would suggest that the increase in amplitude upwards in the field example shows that there were really two mechanisms of folding in this example, buckle folding of a mechanically layered rock and passive, décollement folding resulting from termination of a detachment fault. There is plenty of theoretical basis for analyzing such problems (Johnson and Fletcher, 1994), although it will be somewhat difficult to analyze fault-related folding in a layered medium.

- ? THE MARKERS IN THE MODEL ARE FOLDED INTO SMOOTH ARCS WHEREAS THE LAMINATIONS IN THE ROCK ARE LARGELY BENT AT SHARP HINGES AND STRAIGHT LIMBS.

The viscous material we have analyzed is a linear, homogeneous material. The layers are strictly passive, so we cannot expect the theoretical model to represent mechanical layers, or to produce sharp bends. We have the theoretical basis for analyzing this problem. For example, I have shown that plastic yielding or power-law behavior of layers will result in narrow hinges (Johnson, 1970). One can produce a kind of yielding in elastic or elastic-plastic materials as well (Johnson, 1977, 1980; Honea and Johnson, 1976).

- ? THE GEOMETRIC MODEL SHOWS NO DEFORMATION IN THE FOOTWALL OF THE DÉCOLLEMENT FOLD (FIGURE 4). THE THEORETICAL MODEL SHOWS A SMALLER ANTICLINE IN THE FOOTWALL THAN IN THE HANGING WALL (FIGURE 7), AND SO DOES THE FIELD EXAMPLE (FIGURE 8).

It is possible that the anticline in the footwall is nonintuitive and therefore was left out of the geometric model by Laubscher. It is possible, of course, that the anticline in the footwall in the natural example and in the theoretical model indicate that conditions were different than those assumed for the geometric model. I will investigate these possibilities by examining the flow about a planar fault that is at the contact between a soft and a stiff material. If the substratum is sufficiently stiff, we know that there will be folding only above the décollement (Berger and Johnson, 1982). We could use the method of analysis explained in that paper, but modified to include stiffness contrast for media above and below. Alternatively, we could use Erdogen's solutions (Erdogen, 1971a, 1971b; Erdogen and Gupta, 1971; Erdogen and Aksggan, 1974) for a fracture at contacts between layers or within multilayers.

SUMMARY OF RESULTS OF THE RESEARCH

Grant No. DE FG02 98ER14886 "MECHANICAL MODELS OF FAULT-RELATED FOLDING"

GEOSCIENCES RESEARCH PROGRAM

The objectives of the research proposed in DE FG02 98ER14886 were to provide both a unified, mechanical infrastructure for studies of fault-related folding and to present the results in computer programs that have graphical users interfaces (*GUIs*) so that structural geologists and geophysicists can model a wide variety of fault-related folds (*FaRFs*). The objective was ambitious, but we have made considerable progress toward the objective:

Mechanical Infrastructure: We have developed fault-related folding theory in four areas: We have used fault-dislocation theory in order to study fault-related folds associated with listric and antilistric faults and with gang faults. We have developed boundary-element theory so that we can study growth of fault-related folds that form adjacent to faults that propagate or that are stationary in a flowing material. We have developed the theory of fault-related folding associated with forced folding above a rigid basement. Finally, we have used folding theory to more highly develop the theory of décollement-ramp folding in isotropic and anisotropic materials.

Programming of Graphical-User Interfaces (GUIs): We have written GUIs for each of the theoretical developments so that we and other people can use the theory to study fault-related folding processes. One GUI, SeDeRa, allows one to study folds produced by décollement ramps that are active sequentially. Another GUI, SeGaLi, will model sequential or simultaneous activation of three-dimensional dislocation/fault elements and the folds, strains and displacement fields produced by such activity. Another GUI, FoFo, was designed to allow one to simulate forced folds and tri-shear deformation patterns. Finally, two GUI's have been written, so far, to study the growth of fault-related folds in a flowing medium in which the faults are represented not by dislocations but by boundary elements.

We introduced our GUI research at the annual GSA meeting in 2000:

Johnson, A.M., Durdella, M.J., and Johnson, K.M., 2000, Purdue Faux Pli: A suite of mechanical models of fault-related folding, Geological Society of America, 2000 Annual Meeting, Abstracts with Programs, vol. 32, no. 7, p. 227.

Our computer programs have been copyrighted by the Purdue Research Foundation. The documentation with the computer programs makes clear that the programs were developed with DOE funding. Karen White (kfwhite@purdue.edu), of the Purdue Research Foundation, has contacted management of many petroleum companies in order to inform them of the availability of the software. We provide trial versions of the software on a Purdue website: www.eas.purdue.edu/fauxpli, but the trial versions have several components that are nonfunctional. However, people can get the full versions of the software, free of charge, simply by contacting Karen White or myself (gotesson@purdue.edu) and completing a use agreement. All the manuals are available for downloading at the fauxpli website.

Many individuals at companies and universities have been provided with the full software packages. A list may be obtained from Karen White, who maintains the use agreements.

Field Projects: We did three field projects as part of the research supported by DOE. One of the field projects was based on mapping and analysis of structures within the San Rafael Swell, Utah:

Johnson, K.M., 2000. Methods of structural analysis of the Spotted Wolf Section of San Rafael Monocline, Utah. M.S. Thesis, Earth and Atmospheric Sciences, Purdue University, 202 p.

The research led to nine manuscripts, three of which have been published in the Journal of Structural Geology and six of which, so far, have published only on our web site:

Johnson, K.M., and Johnson, A.M. 2000. Localization of layer-parallel faults in San Rafael Swell, Utah and other monoclinial folds, Journal of Structural Geology 22, 1455-1468.

Johnson, K.M., and Johnson, A.M., 2002. Mechanical analysis of the geometry of forced-folds. Journal of Structural Geology, 24:401-410.

Johnson, K.M., and Johnson, A.M., 2001. Mechanical models of trishear-like folds, Journal of Structural Geology, 24, 277-287.

Johnson, K.M., Manuszak, J.D., and Johnson, A.M., 2000. How Kink Bands Work and How to Work with Kink Bands. 21 p.

www.eas.purdue.edu/physproc/how_kinks_form.htm

Johnson, K.M., and Manuszak, J.D., 2000. How to Analyze Kink Folds, 10 p.

www.eas.purdue.edu/physproc/easy_guide_to_kinks.htm

Johnson, K.M., 2000. Some Comments on high-amplitude kink folding, 12 p.

www.eas.purdue.edu/physproc/high-amplitude_kinks.htm

Johnson, K.M., 2000. Analysis of multiple sets of band faults in porous sandstone, San Rafael Monocline, Utah. 47 p.

www.eas.purdue.edu/physproc/stress_analysis.htm

Johnson, K.M., 2000. Types of internal deformation in the SanRafael Monocline, Utah. 12 p.

http://www.eas.purdue.edu/physproc/small_structures.htm

Johnson, K.M., 2000. Analysis of multiple sets of band faults in porous sandstones, San Rafael monocline, Utah. 34 p.

www.eas.purdue.edu/physproc/stress_states.htm

Another field project focused on some fault-related folds in the Bighorn Basin of Wyoming:

Durdella, M.J., 2001. Mechanical modeling of fault-related folds: West flank of the Bighorn Mountains, Wyoming. M.S. Thesis, Earth and Atmospheric Sciences, Purdue University, 129 p.

Joe's project led to two manuscripts, published on our web site:

Durdella, M.J., and Johnson, A.M., 2000. Geometry and folding mechanisms of the Pitchfork anticline, Bighorn Basin, Wyoming. www.eas.purdue.edu/physproc/kink-wedge.htm

Johnson, A.M., and M.J. Durdella, 2000. The kink-wedge fold. www.eas.purdue.edu/physproc/kink-wedge.htm

The third field project was carried out by Kaj Johnson, Joe Durdella and Arvid Johnson, in conjunction with faculty and students in earthquake engineering at Purdue University and with faculty and students in earthquake engineering at the Middle East Technical University in Turkey. Grant Nos. 85270 CMS and NTP 1381 of the National Science Foundation (Civil Engineering) funded most of the research but the DOE Grant No. DE FG02 98ER14886 funded the travel of Kaj Johnson and Joe Durdella.

Kaj Johnson, Joe Durdella and Arvid Johnson mapped several stretches of the November 1999, Duzce earthquake rupture and measured the deformation of a large viaduct, the Kaynasli Viaduct of the Trans-European highway between Istanbul and Ankara, Turkey. The viaduct passed over the eastern end of the earthquake rupture, so it presented a wonderful opportunity to analyze ground deformation in the vicinity of the North Anatolian fault. Some of the results of this study have been recently published in the Journal of Seismology.

Johnson et al., 2002. An emendation of elastic rebound theory: Main rupture and adjacent belt of right-lateral distortion detected by viaduct at Kaynasli, Turkey 12 November 1999 Duzce earthquake. Journal of Seismology, 6:329-346.

The results were also presented at the NSF-Tubitak, Turkey-Taiwan Grantee Workshop, March 2002 in Antalya, Turkey. Many of the participants indicated that our results were among the more exciting presented at the workshop.

Finally, during the course of the research I have written three manuscripts, which I intend to publish at some point, on propagation of kink bands, on the growth of Atilla faults (deformation bands) and a review of Atilla faults. All the manuscripts are available on my web site:

(Note the underscores beneath the spaces)

Johnson, A.M., 2001, Yielding instability and propagation of kink bands. 50 p.

www.eas.purdue.edu/physproc/propagation_of_kinks.htm

Johnson, A.M., 2001. Propagation of Atilla faults in porous sandstone. 73 p.

Johnson, A.M., 2001. Atilla faults in Entrada Sandstone. 54 p.

www.eas.purdue.edu/physproc/pdf%20Files/AtillaFaults.pdf

LISTRIC AND GANG FAULT (SeGaLiFaRF) FOLDING:

The SeGaLi Model

Some theoretical models of fault-related folding had been explored before we began our research. Geophysicists had already developed the dislocation model of faulting to a high degree and had used the theory to analyze ground displacements during earthquakes, starting with the 1971, San Fernando earthquake in Los Angeles. In one of our approaches to fault-related folding, we have adopted their dislocation model and have put together a computer package (*SeGaLiFaRF*) that allows one to model folding associated with listric and antilistric faults, with faults with variable slip, and with gangs of faults. The computer program is capable of modeling a wide range of fault related folds considered by structural geologists because it can be run in sequential mode to model *fault-propagation folding*, or it can be run in simultaneous mode to model *fault-arrest folding*. We have programmed a GUI with Microsoft Visual Basic and have put the program on our web site: www.eas.purdue.edu/fauxpli/gang_fault_fold.htm (please note the understrikes beneath spaces) for all to use. We have found the program very useful and so have several other investigators.

The figures below show some of the kinds of problems one can model with SeGaLi.

Figure A1. Examples of faults and fault-related folds modeled with the SeGaLi program and GUI. On left, arrangement of fault (dislocation) elements. On Right, serial cross sections, showing three-dimensional structure produced by slip on the fault.

Top, antilistric fault composed of four elements. Middle, listric fault. Bottom, a reverse fault and a backthrust.

Fig. 1. Fault-related fold models: (a) fault-bend fold (Suppe, 1983); (b) fault-propagation fold (Suppe and Medwedeff, 1984, 1990); (c) fault-propagation fold (Chester and Chester, 1990).

Figure A2. From McConnell and others (1997, fig. 1).

For example, one can use the SeGaLi program to model the fault-bend fold of Suppe (Figure A2a) by drawing a horizontal upper decollement, and inclined ramp and a horizontal lower detachment and specifying the slip on each of them. The model includes a stress-free upper surface, so the effect of the free surface can be observed. It is limited, though, by the fact that the medium is homogeneous and isotropic. However, the model is superior to most fault-bend fold modeling in that it is three dimensional: the faults end in the direction normal to the drawing above.

The Fault-Tip Pucker Fold

The same program can be used to study décollement folding, which occurs

Figure A3. Décollement fold according to Hedlund (1997, fig. 8)

at the tip of an arrested fault (Figure A3). According to Berger and Johnson (1982) and to Poblet and others (1997), *décollement folding* occurs where the displacement related to layer-parallel blind thrusting is accommodated by folding in the hanging wall. We can use the SeGaLi program to model a long dislocation and study the deformation of passive layering near one tip of the horizontal fault. If the fault is near the free surface, the deformation is highly asymmetric, with most of the folding in the hanging wall as implied in Figure A3.

Figure 4. Fault-Tip Pucker fold produced by a laterally terminating, horizontal detachment with the Gang-Fault Fold model. (diagram by Joe Durdella, 2001)

Whether close to or distant from the free surface, though, an anticline occurs below and its crest is in back of the tip of the terminating fault (Figure 4). This is one of those “secondary” structures that should be of interest in petroleum exploration. If the overlying unit is layered, though, one tends to observe a different form of fold, somewhat like that shown in Figure A2b. We will discuss a field example of this type of fold that we have studied in Wyoming.

Fault-tip pucker folds form where a detachment surface terminates. If the tip is deeply buried, the amplitudes of folds die out upward, away from the fault tip (Berger and Johnson, 1982). If the fault tip is near the ground surface, though, the amplitudes of folds increase upward. Such a fold is shown in Figure 4, produced mechanically with a dislocation theoretical model

(www.eas.purdue.edu/fauxpli). The dislocation is placed near the ground surface and allowed to slip. A nearly identical form can be produced with a solution for mode II slip on a fracture parallel and near a free surface. In either case, the layering shown in Figure 4 is passive—it is not mechanical layering. The host material is isotropic and homogeneous. The termination of the slip causes the pucker to form, and the free surface causes the pucker to increase in amplitude upward.

Manuals

Explanatory manuals are provided with each of the computer programs on our web site. Volumes I and II introduce and illustrate SeGaLi FaRF, which is based on the solution for a dislocation in a homogeneous, isotropic medium, with a free upper surface that is used by many geophysicists to explore shapes of subsurface earthquake ruptures responsible for deformation at the ground surface resulting from slip along the ruptures. It is particularly nice for studying how fault-related folds form in homogeneous materials close to the ground surface. The user of the programs need not be concerned about how the theoretical model has been derived, although some aspects of the model are explained in Appendix I (www.eas.purdue.edu/fauxpli/manual2.htm). The reference is provided for the solution that is the engine to this program.

Volume I explains how the program operates, answering questions such as, how does one model the folding of passive layers during the propagation of a listric fault from depth toward the ground surface? Or, what are the strains like at the ground surface as a strike-slip fault propagates to the ground surface? Or, how does a thrust fold grow? Or, what is the distribution of horizontal and vertical displacements at the ground surface, or on any horizontal plane beneath the ground surface, as a fault or group of faults slips?

The purpose of volume II (www.eas.purdue.edu/fauxpli/examples2.htm) is to examine real-world questions such as these and provide some guidance on how one can harness some of the power provided by the SeGaLi FaRF computer program. For example, could the peculiar strain belts that we discovered in Northridge, California, by resurveying streets after the 1994 earthquake reflect blind reverse faults? We suspected so, but had limited theoretical tools to demonstrate that the strain belts could be so explained. Other field researchers at that time were certain that the strain belts were due to something—wave focusing, ground shaking, or liquefaction—anything, it seemed, but what we had suggested—simple tectonic deformation due to fault slip at depth. The theoretical results provide some, well, interesting insights. In another example, we ask whether the fracture belts, 50 to 200 m wide, along the 1992 Landers earthquake rupture could have formed as the strike-slip fault(s) propagated (in mode III) toward the ground surface? Other field researchers were certain that the fracture belts were due to ground shaking, not due to tectonic deformation. We had some field evidence that the rupture zones, some places anyway, formed before the fault(s) reached the ground surface. One earthquake geophysicist went so far as to state that it is physically impossible for such fracture belts to form by earthquake rupturing, especially before the fault reached the ground surface. The theoretical results provide a possible explanation for the formation of rupture belts ahead of a strike-slip fault propagating in mode III, but there are features of the rupture belts that remain rather mysterious. Nevertheless, we have used the computer program to explain the belts of right-lateral deformation for the 1991 earthquake rupture in Turkey:

Johnson, A.M., Johnson, K.M., Durdella, J., Sozen, M., and Gur, T., 2002. An emendation of elastic rebound theory: Main rupture and adjacent belt of right-lateral distortion detected by viaduct at Kaynasli, Turkey 12 November 1999 Duzce earthquake. *Journal of Seismology*, 6:329-346.

Also, www.eas.purdue.edu/physproc/turkey_viaduct_study.htm (note: spaces are underscored).

In yet another example we ask how can one explain wildly divergent displacements of ground in the hangingwall and footwall of the 1971 San Fernando earthquake. I don't know why I was so surprised by the differences, but I was. The theoretical results provide a rational explanation and make such observations quite interesting:

Johnson, A.M., and Davidson, X., 1999 . Strain and rupture belts along Sylmar fault segment, 1971 San Fernando earthquake. www.eas.purdue.edu/physproc/1971_san_fernando_ca.htm, 49 p (note: spaces are underscored).

Thus, the second volume introduces some field problems that we have addressed over the years, generally with limited success because of the limitations of the theoretical tools we were using then, and shows how one can use SeGaLi FaRF to analyze and provide some tentative solutions to these problems.

SEQUENTIAL DÉCOLLEMENT RAMP (SeDeRaFaRF) FOLDING

The SeDeRa Model

This Sequential Décollement Ramp Fault Related Fold (SeDeRaFaRF) program has a long history. The algorithm began with a solution derived by Johnson in the late 1970's (Berger and Johnson, 1980) and programmed in FOCAL (defunct DEC language) for the folding process that John L. Rich recognized in 1934. The original solution was for an infinite viscous medium that slides over a decollement surface with ramps and drops so that it produces ramp anticlines as well as ramp synclines (termed rollover structures). Thus the program explained the formation of the structure recognized by Rich as well as a structure, the rollover, that was not recognized until later. The solution and computer programs went through several modifications over the years.

Kilsdonk and Fletcher added a nice touch, a deformable substratum. Later, Johnson added several layers and used his method of boundary constants (Johnson and Pfaff, 1989; Johnson and Fletcher, 1994) to derive various solutions including the present one. The solutions were written into various BASIC computer programs early on, including QUICKBASIC. Then Johnson wrote a GUI program in Microsoft VISUAL BASIC (R) 2.0. Chiu Liu modified that program for VISUAL BASIC 4.0.

At this point we obtained DOE funding for studying fault-related folding. Then a professional programmer, Anthony DeNicola, completely rewrote the program in VISUAL BASIC 6.0. Since then Johnson has made several changes in the program and has tried not to destroy the nice program Anthony wrote. It is available at,

www.eas.purdue.edu/fauxpli/ramp_fold1.htm

During the last few months Anthony has completely rewritten the program, using object-oriented programming:

www.eas.purdue.edu/fauxpli/new_version.htm

The algorithm is an exact solution, to first order, of a boundary-value problem. Thus the kinematics are part of the solution, and are not invented as in most of structural geology, including the highly acclaimed fault-bend-fold perversion of the innovative folding mechanism recognized by John L. Rich.

Some examples of simple structures one can model with SeDeRa are shown below:

Figure A5. Examples of décollement ramp folds. Top, a ramp syncline, also called a rollover. Ramp is a normal fault. Middle, a duplex structure produced by a sequence of reverse-fault ramps. Below, a double ramp fold. Fold produced where there is ramping to one level and then ramping to a higher level.

Needless to say, the SeDeRa FaRF program simulates the essential features of the John L. Rich model of ramp folding, including Suppe's "fault-bend-fold."

Manuals

A manual, Volume III (www.eas.purdue.edu/fauxpli/manual1.htm), is available to help people use the SeDeRa FaRF programs. Volume IV (www.eas.purdue.edu/fauxpli/theory2.htm) explains details of the derivation of the solution.

In its most elementary uses, SeDeRa FaRF synthesizes ramp anticlines and synclines (ramp folds or fault-bend folds and rollover structures). It offers, though, insights into a much wider range of

structures. This volume explains how various aspects and modes of the program operate, answering questions with "forward modeling" such as, what is the shape of stratification in a ramp anticline in a layer overlain by the ground surface and underlain by a stiff or soft substratum? Or, what does the folding look like if one décollement activates and then, later, a new décollement forms and activates, deforming the first? Such questions are easy to answer with SeDéRa FaRF. We put in the décollement surfaces and push the blocks over one another to see what happens to stratification.

Just as easily, SeDéRa FaRF can be used in "backward modeling" to deduce the most likely pattern of décollement surfaces responsible for a fold that is only partly visible. This is a very common problem in analysis of fault-related structures. In this case, we begin with the final fold, say by importing a TIFF file of the profile of a partly visible fold. One then sketches a plausible décollement surface and ramps, and moves the blocks incrementally backwards, watching how the structure unfolds. If we can unfold the structure, we may have figured out the faulting and sequence of events responsible for the structure through the backward modeling. This is not to be confused with "inversion," because "inversion" is inherently linear. What we are dealing with here is fundamentally nonlinear. We actually have to run the process backwards to uncoil the sequence of events.

SeDéRa FaRF is based on an analytical solution for up to three décollement surfaces separating four layers in homogeneous isotropic and anisotropic media. The solution is implemented for three décollement surfaces, but for slip only along the middle one. Thus, the system consists of an overlying medium (can be air) on a décollement (surface of easy slip) on a layer. This layer is in contact at its base with another layer across the décollement surface, which may contain one or several ramps, essentially of any reasonable shape. The lower layer rests on a thick lower medium (basement) across the third décollement. The thicknesses and properties are all adjustable, so the system solves a host of fault-related folding problems. A feature that makes SeDéRa FaRF more powerful than our previous analyses of ramp folding is that the layer above and below the active décollement can be isotropic, highly anisotropic, or inversely anisotropic. The analytical solutions for all three cases are derived in detail in Volume IV of the manual.

FORCED FOLDS

Typical structures in the Rocky Mountain Foreland of the western United States are fault-related folds that form over basement faults. Outcrops and seismic profiles show that the folds are typically asymmetric monoclines with long, gently-dipping back-limbs and short, steeply-dipping forelimbs, overlying straight or curved fault surfaces in basement rock (e.g., Prucha et al., 1965; Stearns, 1971; Reches, 1978; Stone, 1983a and b, 1985; Schmidt et al., 1993; Tindall, 2000; Patton, Logan and Friedman, 1998; Tindall and Davis, 1999).

Forced-folding has been proposed as the mechanism for some of these structures as well as similar structures throughout the world (e.g., Cosgrove and Ameen, 2000; Maurin and Niviere, 2000; Keller and Lynch, 2000; Mansfield and Cartwright, 2000; Watkinson and Hooper, 2000; Sattarzadeh, Cosgrove and Vita-Finzi, 2000; Strayer and Suppe, 2002; Erickson, Strayer and Suppe, 2001; Strayer, Hudleston and Lorig, 2001; Cardozo, Bhalla, Zhender and Allmendinger, 2002; Willsey, Umhoefer and Hilley, 2002).

The essential features of the forced-folding mechanism are a sedimentary cover that deforms more or less passively and rigid basement blocks that are displaced along planar or listric faults (Stearns, 1978; Reches and Johnson, 1978). Evidence for undeformed, perhaps rigid, basement blocks that moved during folding of the sedimentary cover has been cited in several folds in the Rocky Mountain foreland (Prucha et al., 1965; Stearns, 1971; Mathews, 1986; Erslev et al. 1988; Erslev and Rogers, 1993).

Efforts to explain forced-folds have followed three, largely divergent paths: theoretical analysis, experimentation, and kinematic analysis. The earliest study was a combined experimental/theoretical analysis by Sanford (1959), who experimented with clay models and performed theoretical analyses with elasticity theory. Reches and Johnson (1978) performed a similar kind of theoretical analysis, but for layers, in order to explore possible mechanisms of formation of the Palisades monocline in the Grand Canyon. Their theoretical analysis, like that of Sanford is what we call today a zeroth-order analysis. It does not account for the deformation of the ground surface or, for that matter, the deformation of the basement-sedimentary cover contact (Johnson and Fletcher, 1995). Their solution is like that of a problem in infinitesimal elasticity theory.

Withjack et al. (1990) performed clay experiments of forced-folds over normal faults. Friedman *et al.* (1980) formed small forced-folds experimentally in rock specimens subjected to high pressures.

Kinematic models and geometric constructions have been the most common method of describing forced folds. The geometric constructions of the parallel kink construction of basement-involved folding by Narr and Suppe (1994) assumes that bed length and cross-sectional area is preserved, and bed thickness and limb dips are uniform. The folds are formed over faults composed of linked, straight-line segments. The trishear geometric model, was introduced by Erslev (1991). Trishear was invented to produce "more nearly realistic" fold geometries with rounded hinges and variable forelimb dip angles (Fig. 1A).

The trishear model was made into a kinematic model by Hardy and Ford (1997), Allmendinger (1998), and Zehnder and Allmendinger (2000). Trishear folds are formed by specifying constant velocities on either side of a central, triangular region and specifying a velocity distribution within the triangular region that satisfies the continuity equation (conservation of area).

<http://pangea.stanford.edu/~kaj/kaj.html>

The approach taken by Kaj Johnson is different from all the previous ones. It is most nearly similar to that of Sanford (1959) and Reches and Johnson (1978). It is totally different from, yes, one might even say unrelated to, the others. It is different from the other theoretical models in that the analysis is performed to first order in the slope of waveforms in the basement-cover contact, and it is derived for anisotropic materials. The materials considered here flow, so the solutions are inherently nonlinear. The methods used here were developed by Johnson and Fletcher (1995).

The approach to forced folding was explained at a GSA meeting:

Johnson, K.M., and Johnson, A.M., 2000, Comparison of a mechanical model with kinematics descriptions of basement forced-folding, Geological Society of America, 2000 Annual Meeting, Abstracts with Programs, vol. 32, no. 7, p. 30.

The purpose of the theoretical analysis is to systematically investigate, with mechanical analysis, effects of various mechanical properties on fold form. The MatLab computer program was written to implement the solution so that its power can be accessed by anyone willing to run a computer model. It was written with a GUI, a rather friendly user interface that should make most aspects of the program easy to understand.

In the Manual, Volume V, we present users guide for the GUI for rigid-basement, forced-folds and tri-shear folds, termed FoFoFaRF, derived by Kaj Johnson and programmed in MatLab. These latter solutions are all for relatively near-surface fault-related folding and treat isotropic,

anisotropic and inversely anisotropic materials. Volume V includes an “autopilot” run and explains the development of the theory. It includes two manuscripts (subsequently published as papers):

Johnson, K.M., and Johnson, A.M., 2001, Mechanical models of trishear-like folds, *Journal of Structural Geology*, 24, 277-287.

Johnson, K.M., and Johnson, A.M., 2002. Mechanical analysis of the geometry of forced-folds. *Journal of Structural Geology*, 24:401-410.

Figure A6. Examples of forced folds. The basement rocks, containing the fault, are very stiff. The overlying rocks are deformable. They are homogeneous or anisotropic.

Fig. 1. Evolution of three different models for asymmetric kink detachment folds involving a homogeneous competent layer detached over a ductile unit. The models are calculated using the parameters given in Table 1 and all three kinematic models produce the same final fold shape. All the models are area balanced for both the competent and ductile units. In the case of Models 1 and 2 variations in the depth to detachment occur in order to area balance the ductile unit (Homza and Wallace, 1995).

Figure A7. Three kinematic models of the evolution of a décollement fold (Poblet and others, 1997, fig. 1).

KINK-WEDGE (KINK DETACHMENT) FOLD

While we were studying the geometry of the Pitchfork anticline near Cody Wyoming, we recognized a kink fold, which we term the *kink-wedge fold* (Durdella and Johnson, 2001). The kink-wedge fold is, apparently, the same as the kink-detachment fold of Poblet and others (1997, Figure A7). In our view, the kink-wedge fold is merely a variation on two kinds of folds that we have studied previously, the fault-tip pucker fold that occurs at the tip of a thrust blind fault (Berger and Johnson, 1982) and the kink band (Honea and Johnson, 1976; Reches and Johnson, 1976). Like the fault-tip pucker fold, the kink-wedge fold is a fault-tip folds.

The essential mechanism of fault-tip folding is the abrupt lateral termination of slip along a detachment fault, a décollement. Because of the termination, the material piles up near and around the termination and forms an anticline. The anticline has quite different forms, depending on the properties of the host rock and the shape of the fault in the termination area. Here we will consider only two forms of fault-tip fold, the *fault-tip pucker* fold that occurs if the host rock is nearly homogeneous and isotropic, and the *kink-wedge* fold that occurs if the host rock has properties that favor the formation of kink bands under layer-parallel compression (Johnson, 1977).

Poblet and others (1997) prefer to leave the origin of the kink-wedge fold open, without a mechanical explanation. As most structural geologists, they do not believe that we can understand the formation of kink-like folds. They show three ways that a kink-wedge fold might form in Figure A7. Even though the geometry of the final folds is the same in Figure A7, the time histories of uplift, rates of limb rotation and rates of limb lengthening are quite different for the three

models. In models 1 and 3 both uplift and limb dip increase rapidly in the first stages and then increase at a linear rate. In both models all axial surfaces progressively rotate with increasing shortening in order to maintain constant stratigraphic thicknesses. In model 2 the uplift of the décollement fold increases linearly and limb dip is constant as shortening progresses. Limb length increases linearly. Thus, because these three kinematic models, with different folding mechanisms, can produce identical fold shapes, the fold form is not diagnostic of the folding mechanism.

Our view is that the kink-wedge fold forms under the same conditions as the fault-tip pucker fold, but the properties of the medium are quite different. The kink-wedge fold forms in layered materials that can slip but with some difficulty. The contact strength between layers must be overcome in order for the kink-wedge fold to form (Johnson, 1977; K. Johnson et al., 2001). The mechanics of kink-wedge folding is explained in a manuscript,

Johnson, A.M., and M.J. Durdella, 2000. The kink-wedge fold. www.eas.purdue.edu/physproc/kink-wedge.htm

The pitchfork anticline in Wyoming is essentially a type example of a kink-wedge fold. Its profile is shown below.

Figure A8. The most nearly accurate reconstruction of Pitchfork anticline based on the combination of surface data, well logs and a seismic profile. Formations are projected above the ground surface to show extent of strata incorporated in the folding of Pitchfork anticline.

We have simulated the Pitchfork anticline by applying knowledge gained from the Kink-wedge and Forced-fold models to the actual fold. The result, Figure A9 is a theoretical reconstruction of Pitchfork anticline. The theoretical reconstruction allows us to suggest the fold form where data are unavailable, such as above the ground surface, on the sides of the fold and in the area of the Precambrian basement. We constructed this profile by graphically appending (not superimposing) the solution for a kink-wedge model with a friction angle of 30° in the overburden and the forced-fold solution in the deep profile. The theoretical reconstruction (Figure A9) obeys all the data presented in the deep profile and uses the solutions to the kink-wedge and forced-fold models to better approximate the fold form where data is unavailable.

Figure A9. Theoretical reconstruction of Pitchfork anticline. This construction obeys the data provided by the deep cross section (Figure A7) and is supplemented by information provided by the theoretical models.

Details of the analysis leading to the reconstruction are presented in our manuscript,

Durdella, M.J., and Johnson, A.M., 2000. Geometry and folding mechanisms of the Pitchfork anticline, Bighorn Basin, Wyoming. www.eas.purdue.edu/physproc/kink-wedge.htm

DETAILS OF RESEARCH ACCOMPLISHED UNDER DOE GRANT

FIRST OBJECTIVE: PRESENT USER-FRIENDLY COMPUTER PROGRAMS:

THE OBJECTIVE IS TO PRESENT THE RESULTS AND METHODS OF ANALYSIS WITH COMPUTER PROGRAMS THAT ARE QUITE USER FRIENDLY, SO THAT STRUCTURAL GEOLOGISTS AND GEOMECHANICISTS CAN READILY USE THEM IN PRACTICE.

FOREWORD TO HANDBOOKS

The five volumes of this manual and the five computer programs, SeDéRa FaRF, FoFo FaRF, SeGaLi FaRF, FauxPasEQ and FauxPasVISCO, have been written for the practicing structural geologist. These programs are the first in a set of useful computer tools placed before members of the structural geology community by Faux Pli? software of Purdue University. The Department of Energy provided us with a research grant for this purpose. The computer programs and the documentation that goes with them are products of that research. The user is provided with a GUI; a graphical user's interface, that requests information from the operator of the program and returns results, generally in graphical formats.

The programs and the five manuals described here may be downloaded in demonstration form at:

www.eas.purdue.edu/fauxpli

The full versions of the programs may be obtained from Karen White at the Purdue Research Foundation (kwhite@purdue.edu) or from myself (gotesson@purdue.edu). We merely ask you to submit a user's permit before we send you the software. The software is owned by Purdue Research Foundation.

Volumes I and II of the user's manual introduce and illustrate SeGaLi FaRF, which is based on the solution for a dislocation in a homogeneous, isotropic medium, with a free upper surface that is used by many geophysicists to explore shapes of subsurface earthquake ruptures responsible for deformation at the ground surface resulting from slip along the ruptures. It is particularly nice for studying how fault-related folds form in homogeneous materials close to the ground surface. The

user of the programs need not be concerned about how the theoretical model has been derived, although some aspects of the model are explained in Appendix I. The reference is provided for the solution that is the engine to this program.

Volume I explains how various aspects and modalities of the program operate, answering questions such as, how does one model the folding of passive layers during the propagation of a listric fault from depth toward the ground surface? Or, what are the strains like at the ground surface as a strike-slip fault propagates to the ground surface? Or, how does a thrust fold grow? Or, what is the distribution of horizontal and vertical displacements at the ground surface, or on any horizontal plane beneath the ground surface, as a fault or group of faults slips?

The purpose of volume II is to examine real-world questions such as these and provide some guidance on how one can harness some of the power provided by the SeGaLi FaRF computer program. For example, could the peculiar strain belts that we discovered in Northridge, California, by resurveying streets after the 1994 earthquake reflect blind reverse faults? We suspected so, but had limited theoretical tools to demonstrate that the strain belts could be so explained. Other field researchers at that time were certain that the strain belts were due to something—wave focusing, ground shaking, or liquefaction—anything, it seemed, but what we had suggested—simple tectonic deformation due to fault slip at depth. The theoretical results provide some, well, interesting insights. In another example, we ask whether the fracture belts, 50 to 200 m wide, along the 1992 Landers earthquake rupture could have formed as the strike-slip fault(s) propagated (in mode III) toward the ground surface? Other field researchers were certain that the fracture belts were due to ground shaking, not due to tectonic deformation. We had some field evidence that the rupture zones, some places anyway, formed before the fault(s) reached the ground surface. One earthquake geophysicist went so far as to state that it is physically impossible for such fracture belts to form by earthquake rupturing, especially before the fault reached the ground surface. The theoretical results provide a possible explanation for the formation of rupture belts ahead of a strike-slip fault propagating in mode III, but there are features of the rupture belts that remain rather mysterious. In yet another example we ask how can one explain wildly divergent displacements of ground in the hangingwall and footwall of the 1971 San Fernando earthquake? I don't know why I was so surprised by the differences, but I was. The theoretical results provide a rational explanation and make such observations quite interesting. Thus, the second volume introduces some field problems that I have addressed over the years, generally with limited success because of the limitations of the theoretical tools I was using then, and shows how we can use SeGaLi FaRF to analyze and provide some tentative solutions to these problems.

Volume III introduces SeDéRa FaRF, which is based on solutions freshly derived for a four-layer system for this program. In its most elementary uses, SeDéRa FaRF synthesizes ramp anticlines and synclines (ramp folds or fault-bend folds and rollover structures). It offers, though, insights into a much wider range of structures. This volume explains how various aspects and modes of the program operate, answering questions with "forward modeling" such as, what is the shape of stratification in a ramp anticline

Figures: Examples of SeGaLi FaRF in action

in a layer overlain by the ground surface and underlain by a stiff or soft substratum? Or, what does the folding look like if one detachment activates and then, later, a new detachment forms and activates, deforming the first? Such questions are easy to answer with SeDéRa FaRF. We put in the detachment surfaces and push the blocks over one another to see what happens to stratification.

Just as easily, SeDéRa FaRF can be used in “backward modeling” to deduce the most likely pattern of detachment surfaces responsible for a fold that is only partly visible. This is a very common problem in analysis of fault-related structures. In this case, we begin with the final fold, say by importing the profile of a partly visible fold, sketching a plausible detachment surface and ramps, and moving the blocks incrementally backwards, watching how the structure unfolds. If we can

unfold the structure, we may have figured out the faulting and sequence of events responsible for the structure through the backward modeling. This is not to be confused with “inversion,” because “inversion” is inherently linear. What we

Figures: Examples of SeDeRa FaRF in action.

are dealing with here is fundamentally nonlinear. We actually have to run the process backwards to uncoil the sequence of events.

SeDéRa FaRF is based on an analytical solution for up to three detachment surfaces separating four layers in homogeneous isotropic and anisotropic media. The solution is implemented for three detachment surfaces, but for slip only along the middle one. Thus, the system consists of an overlying medium (can be air) on a detachment (surface of easy slip) on a layer. This layer is in contact at its base with another layer across the décollement surface, which may contain one or several ramps, essentially of any reasonable shape. The lower layer rests on a thick lower medium (basement) across the third detachment. The thicknesses and properties are all adjustable, so the system solves a host of fault-related folding problems. A feature that makes SeDéRa FaRF more powerful than our previous analyses of ramp folding is that the layer above and below the active décollement can be isotropic, highly anisotropic, or inversely anisotropic. The analytical solutions for all three cases are derived in detail in volume IV.

Figures: Examples of FoFo FaRF in action.

In Volume V we present a GUI for rigid–basement, forced–folds and tri–shear folds, termed FoFoFaRF, derived by Kaj Johnson and programmed in MatLab. These latter solutions are all for relatively near–surface fault–related folding and treat isotropic, anisotropic and inversely anisotropic materials. Volume V includes an “autopilot” run and explains the development of the theory. It includes two manuscripts, which have been submitted for publication, which illustrate use of the simulation program.

We are in the process of freshly re-deriving Wei’s analysis, called Faux Pas, of fault-related folding (plane flow) based on surface elements. Wei wrote his program in Fortran (Wei, 1997). In the re-derivation we will present the program in Microsoft Visual Basic, with appropriate GUIs. At this point Faux Pas is available (in beta form, only) for folding associated with a series of earthquakes on the faults (FauxPasEQ) and for folding in which there is overall deformation, such as might happen in rocks that are flowing (e.g., metamorphism or pressure solution). This version is called FauxPasVISCO.

So far we have not had time to write the manuals for FauxPasEQ and FauxPasVISCO. They will be available on the website within a year, probably.

We hope that structural geologists will become more comfortable with mechanical models and learn to appreciate and utilize their incredible power and versatility. Unlike the *ad hoc* geometric and kinematic constructions, correct mechanical models provide us all with correct intuition, which we all need in order to guess solutions to mechanical problems, and the mechanical constructions can actually predict new fold forms, forms that have heretofore been unrecognized.

I would like to thank several people and organizations. The Department of Energy has provided funding for developing theoretical models and GUI computer implementations for fault-related folding through Grant No. DE FG02 98ER14886 supervised by Nick Woodward. Phil Berger impressed upon me the importance of fault-related folding in the late 1970’s. Discussions with Nick Woodward, Mark Fisher, Ken Ridgway, Kaj Johnson, Raymond Fletcher and Rick Allmendinger have helped me more nearly understand the aspirations of mainstream structural geologists. Joe Durdella and Kaj Johnson have been tolerant sounding boards for many versions of the software. Anthony DeNicola wrote the first versions of SeGaLi and SeDéRa and provided me with a sound infrastructure in Microsoft¹ Visual Basic? on which I could build the existing programs. I accept responsibility for all the remaining errors in the programs.

Please bring errors and confusing features of the programs to my attention.

Faux Pli? Software

<http://www.eas.purdue.edu/fauxpli>

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18 August 2000 (SeGaLi)

¹ Visual Basic is a wonderfully powerful computer language for writing “user-friendly” software.

12 October 2000 (SeDéRa)

November 2000 (FoFo)

October 2002 (Faux Pas)

MANUSCRIPT ON THE KINK-WEDGE (DÉCOLLEMENT) FOLD

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FOREWORD

Two of the objectives of the proposed research were furthered by the M.S. thesis research of Joe Durdella on the forms and origins of some folds on the western side of the Bighorn Basin, Wyoming. We expected the folds to be garden-variety thrust folds of the types described widely by many authors, including Don Stone (1983a, 1983b, 1985a, 1985b, 1993). Joe did a very thorough job of determining and documenting the form of one of the folds, the Pitchfork anticline. He collected and collated surface information (field mapping), borehole information and seismic information in order to determine most geometric features of the fold. Then he compared the geometric features with those produced by a series of fault-related folding theories. In the end, we had to develop a new theoretical model in order to describe the form of the fold. The theoretical model consists of a combination of detachment-termination folding of a layered medium that favors the production of kink bands and basement forced folding.

And in the process we developed a nice model for the Cincinnati fold, shown in previous pages; we now understand that it is a result of detachment-termination folding but in a layered material that favors kink folding. We suspected this at the time the research proposal was written, but, of course, we had not yet done the research to determine whether our hunch was correct.

INTRODUCTION

While we were studying the geometry of the Pitchfork anticline near Cody Wyoming, we investigated the kink-wedge *fold* (Durdella and Johnson, 2001), which is very closely related to the highly angular décollement folds described, for example, in Figures 4 and 6 in previous pages. The kink-wedge fold is merely a variation on two kinds of décollement folds that we have studied previously, the fault-tip pucker fold that occurs at the tip of a thrust blind fault (Berger and Johnson, 1982) and the kink band (Honea and Johnson, 1976; Reches and Johnson, 1976). Both kinds of folds are fault-tip folds.

DÉCOLLEMENT FOLDS

The essential mechanism of décollement folding is the abrupt lateral termination of slip along a detachment fault, a décollement. Because of the termination, the material piles up near and around the termination and forms an anticline. The anticline has quite different forms, depending on the properties of the host rock and the shape of the fault in the termination area. Here I will consider only two forms of décollement fold, the *fault-tip pucker* fold that occurs if the host rock is nearly homogeneous and isotropic, and the kink-wedge fold that occurs if the host rock has properties that favor the formation of kink bands under layer-parallel compression (Johnson, 1977).

A.

B.

Figure 1. Fault-tip folds. A. Fault-tip fold in Entrada Formation, Henry Mountains, UT. A décollement enters the view from the right, at or above ground level. It terminates in a tail fault that cuts up through part of the section. The amount of throw on the monoclinial flexure is much

greater than on the throw on the fault. B. Fault-tip fold in interbedded limestone and shale of the Kope Formation, Cincinnati, Ohio. The décollement ends near midwidth of the bump in the light-toned beds near the base of the view.

A very nice example of a fault-tip fold is shown in Figure 1A, the Bullfrog fold, exposed on both sides of Utah Highway 276 on the southwest side of Mt. Ellsworth, near Bullfrog, Utah³. The fold is in interbedded sandstone and shale of the Entrada Sandstone. There is a predominant fault and many smaller ones. All the faults share a common strike, but some dip north and others dip south. Most of the faults are low-angle, thrust faults, but a few are steep, reverse faults. The major fault dips at 40 to 50° toward the south. The slip along the major fault is about 3 m near the base of the roadcut. The throw of the monoclinial structure at the level of the ground surface is about 4 m. Thus the amount of throw along the major fault is less than the throw of the overlying monoclinial structure.

THE FAULT-TIP PUCKER FOLD

Fault-tip pucker folds form where a detachment surface terminates. If the tip is deeply buried, the amplitudes of folds die out upward, away from the fault tip (Berger and Johnson, 1982). If the fault tip is near the ground surface, though, the amplitudes of folds increase upward. Such a fold is shown in Figure 2, produced mechanically with a dislocation theoretical model (www.eas.purdue.edu/fauxpli). The dislocation is placed near the ground surface and allowed to slip. A nearly identical form can be produced with a solution for mode II slip on a fracture parallel and near a free surface. In either case, the layering shown in Figure 2 is passive—it is not mechanical layering. The host material is isotropic and homogeneous. The termination of the slip causes the pucker to form, and the free surface causes the pucker to increase in amplitude upward.

Figure 2. Fault-Tip Pucker fold produced by a laterally terminating, horizontal detachment with the Gang-Fault Fold model. (diagram by Joe Durdella, 2001)

Whether close to or distant from the free surface, though, an anticline occurs below and its crest is in back of the tip of the terminating fault (Figure 2). This is one of those “secondary” structures that should be of interest in petroleum exploration.

³ Aydin, 1973, p. 21,

THE KINK-WEDGE FOLD

The kink-wedge fold forms under the same conditions as the fault-tip pucker fold, but the properties of the medium are quite different. The kink-wedge fold forms in layered materials that can slip but with some difficulty. The contact strength between layers must be overcome in order for the kink-wedge fold to form (Johnson, 1977; K. Johnson et al., 2001).

THEORETICAL MODEL

A sketch of a kink-wedge is shown in Figure 3. The box in Figure 3 shows the area of the termination of an idealized kink fold, consisting of the termination area plus a pair of conjugate kink bands.

Figure 3. Idealized symmetric conjugate kink fold terminating in kink wedges. The ultimate locking angle, α_u , and the angle of the lower wall, ϕ of the tilted inner limb of the kink fold are governed by the friction angle, ϕ . Slip occurs within the limb bounded by the upper and lower walls of a kink band and along a detachment at the base of the medium. The box in the center of the diagram indicates the area of the kink fold that pertains to the kink-wedge fold.

In the *kink-wedge* mechanical model, the system is a multi-layer with identical layers and contact strength between layers. The kink-wedge fold is a member of the kink-fold family, so it satisfies all the conditions that have been discovered concerning the cause of kink folding. The kink-fold mechanical model was developed many years ago (Peng and Johnson, 1972; Honea and Johnson, 1976; Reches and Johnson, 1976) and has been clarified with several subsequent studies (Pfaff and Johnson, 1989; Johnson, 1995; K. Johnson and Manuszak, 2001; K Johnson et al., 2001). The new features that characterize the *kink-wedge* are:

- ? the wedge terminates at a décollement or detachment surface

- ? and the width of the kink band in a kink wedge is variable

For many years it has been known that the kink fold mechanical model⁴ has three essential features:

- ? Kink bands form in response to layer-parallel (or nearly parallel) compression.
- ? A result of high layer-parallel compression, a localized perturbation in dip of layering will localize a *shearing instability*. The localized deformation and the inclined, straight limbs of a kink fold are results of the *shearing instability* (Peng and Johnson, 1972; Reches and Johnson, 1976).
- ? Another result of the high layer-parallel compression is that a localized perturbation will result in a *buckling instability* (Honea and Johnson, 1976; K. Johnson, 2000), in which the layers deflect laterally. The width of the kink band is generally determined by the thicknesses of layers.

Over the years, it has been shown that kink folds may form in elastic or viscous (linear or nonlinear) layers. The properties of contacts between layers (or of soft interlayers) though, must be nonlinear for the yielding instability to occur (Pfaff and Johnson, 1989). In the classic kink fold mechanical model (Johnson, 1977), there are layers in contact with each other and the contact properties are described in terms of Coulomb's law. In the generalized kink fold model (Pfaff and Johnson, 1989), the resistance to slip is simply a decreasing, nonlinear function of the amount of slip.

The kink-wedge is related to a kink band or a pair of kink bands (Figure 4A). A double kink wedge, even if asymmetric, might be the termination of a conjugate pair of kink bands, as shown in Figure 4A⁵. On the other hand, a single kink wedge might be the termination of a single kink band (Figure 5A)⁶. In the latter case the layers in the wedge are essentially parallel to layers within the kink band and the width of the wedge decreases toward the termination of the kink band.

One of the features of the monoclinical kink band, noted by Johnson (1977, p. 274), is the "backlimb rotation," that is, the tilting of a longer limb facing in the opposite direction from the kinked limb. "Backlimb rotation" is particularly noticeable in a series of monoclinical kink bands. The role of "backlimb rotation" is to return the layers to their position outside the influence zone of the kink limb. The "backlimb rotation" is noticeable next to the kink wedge as well as the kink band. In the example shown in Figure 5, the tilting of layers within the kink band is clearly due to right-lateral slippage between layers as evidenced by the displaced vertical markers in the experiment. The tilting of beds within the "backlimb" is clearly due to smaller left-lateral slippage between beds within the "backlimb."

⁴ Not to be confused with the fold representations drawn by Suppe, which are termed "fault-bend folds" and "fault-propagation folds," and have no physical basis—they are merely geometric constructs. Kink folds have been analyzed mechanically and experimentally and are predictable structures.

⁵ The thinner rubber strips were 0.08 cm thick.

⁶ The thicker rubber strips were 0.25 cm thick and 25 cm long. The marker lines drawn across the layers were spaced about 1 cm apart (Johnson, 1977, p. 158).

A.

B.

Figure 4. Experimentally produced kink fold using 0.08 cm thick rubber strips. (a) Distinction between the kink bands and the kink wedge of the asymmetric kink fold. (b) Variation of locking angles within the kink bands and the kink wedge. The angle of the lower wall is indicated as well. (Experiment archive photos from A.M. Johnson).

The kink-wedge associated with the conjugate kink bands shown in Figure 4B is asymmetric. Within the kink wedge the angle between layers inside and outside the kink range from 45° to 48° on one side and from 33° to 38° on the other. The angles are generally about 53° in the wider kink limb. Assuming that the maximum compression was parallel to nonfolded layering, the ultimate locking angle should be $\theta_u = \theta/2 - \phi$. Also, assuming the layering was at the ultimate locking angle, we estimate a friction angle of about 37° . This value is reasonable for the thin rubber strips used in the experimental fold (Johnson, 1995).

The idealized, symmetric conjugate kink fold shown in Figure 3 is imagined to have been produced by the maximum compression direction, σ_1 , applied horizontally, parallel to layering outside the fold. The friction angle was assumed to be 30° .

If there is layer-parallel shear stress, as well as compression, a monoclinical kink fold forms as explained thoroughly elsewhere (Reches and Johnson, 1976; Johnson, 1995; K. Johnson et al., 2001). Figure 6 shows three examples of idealized monoclinical *kink-wedges*. Note that in each case the short limb faces left. The direction of facing of a monoclinical kink band is opposite to that of a drag fold; in any case, for the sense of symmetry shown in Figure 6, the sense of layer-parallel shearing is right-lateral. This is equivalent to saying that the layer-normal shear stress is left-lateral (see K. Johnson and Manuszak, 2001, and K. Johnson et al., 2001 for a thorough discussion of interpretation of kink bands).

A.

B.

Figure 5. Experimentally produced kink band using 0.25 cm thick rubber strips. (a) Distinction between the kink band and the kink wedge. (b) Variation of locking angles within the kink band and kink wedge. The acute angle between the lower wall of the kink band and the layers outside the kink band is also shown. (Experiment archive photos from A.M. Johnson).

Geometry of the Idealized kink-wedge

The theory of kink-band formation and the orientations of limbs in a kink band, including a kink-wedge fold, are discussed in detail by K. Johnson and Manuzek (2000) and by K. Johnson and others (2000). The following discussion is due to them.

Figure 6 shows the three types of kink folds and the maximum compression directions responsible for the formation of each type. The inclination of the direction of maximum compression, β , determines whether the kink folds form as a set of monoclinical or conjugate kink bands. When the orientation of maximum compression is parallel to layering ($\beta = 0$), *symmetric* conjugate kink bands form (Figure 6A). The locking angles of the left-facing limb, $\theta_{L\rightarrow}$, and the right-facing limb, $\theta_{L\leftarrow}$, are equal in magnitude, $|\theta_{L\leftarrow}| = |\theta_{L\rightarrow}|$. When the maximum compression is slightly inclined ($\beta \neq 0$) to the layering in a clockwise or counterclockwise sense, kink bands may form as *asymmetric* conjugate kink bands $\theta_{L\leftarrow} \neq \theta_{L\rightarrow}$ (Figure 6B). Finally, when the maximum compression is more highly inclined to the layering, kink bands may form as *monoclinical* kink folds with all left-facing or right-facing limbs (Figure 6C).

Figure 6. The three different types of kink bands and the orientation of the maximum compression direction, β , responsible for each type. The locking angle, θ_L , is negative for right-facing and positive for left-facing kink bands. (From K. Johnson and others, 2000).

Figure 7. Butterfly diagram relating locking angles, θ_L , of right- and left-facing kink bands to orientation, β , of the maximum compression and the angle of internal friction, ϕ . Inset diagrams indicate facing of kink bands. Heavy, dashed lines are for use with the Navajo Sandstone example. The heavy, solid line is for use with the McCarthy example. (From K. Johnson and Manuzek, 2000).

The locking angles of kink bands can be expressed in terms of only two variables, the friction angle, ϕ , at layer surfaces and the inclination of maximum compression, β (Reches and Johnson, 1976). For positively sloping layers in the inner limb (left-facing limb in Figure 6B)

$$\theta_L = 90^\circ - \phi - \beta; \quad \theta > 0, \quad (1a)$$

whereas for negatively sloping layers (right-facing limb in Figure 6B),

$$\theta_L = 90^\circ - \phi + \beta; \quad \theta > 0. \quad (1b)$$

We introduce the butterfly diagram (Fig. 7), which is a plot of eqs. (1), but with appropriate limits, in order to facilitate the determination of friction angle and inclination of compression in the case of measurements of a pair of conjugate kink bands or of multiple measurements of kink bands. As we will show below, if we have measurements of the orientation of the limbs of a conjugate fold, we can use the butterfly diagram to determine both the friction angle and the inclination of the compression.

The butterfly diagram introduced by K. Johnson and Manuzek (2000) (Figure 7) helps one to determine these quantities. It is a plot of the two equations presented in the preceding paragraph, but with necessary limiting conditions. (Note: the heavy lines are for use with an examples discussed below.) It plots β as a function of locking angle θ_L for $\phi = 10, 20, 30$ and 40° . Two of the boundary lines are a result of limiting angles of maximum compression, β . Given a value for friction angle, ϕ , the angle of inclination of maximum compression, β , can range between $45^\circ - \phi/2$ and $45^\circ + \phi/2$ (the range falls within the shaded area of the butterfly diagram). This is so because, if $|\beta|$ exceeds $45^\circ - \phi/2$, the layers in the outer limbs outside the pupative kink band will slip and a kink band will not develop. As a result of the range of possible angles, β , there is a range of possible locking angles for a given friction angle, ϕ .

Figure 8. Asymmetric conjugate kink bands in cross-beds of Navajo Sandstone. Locking angle is 66° for the left facing limb -34° for the right facing limb. (from K. Johnson and Manuzek, 2000).

Figure 8 shows conjugate kink bands measured by K. Johnson (K. Johnson, 2000) in San Rafael Swell, Utah. The orientation of the inner limb varies with position along the kink band, the largest angle of inclination of the inner limb is the locking angle, θ_L . The largest angle in the left-facing limb in Figure 8 is $\theta_{Lr} = +66^\circ$ (range 34 to 66°) and the corresponding angle in the right-facing limb is $\theta_{Lr} = -34^\circ$ (range -20 to -34°).

We can use the butterfly diagram in Figure 7 to determine β and ϕ . The point (ϕ, β) which is the solution for both the right and left facing limbs must lie on the heavy, vertical dashed lines through 66° and -34° . The heavy, horizontal dashed line shows the solution: $\beta = 17^\circ$ and $\phi = 40^\circ$.

One of the critical points, here, is that the orientations of the bedding in the limbs are not arbitrary according to the theory of kink-band formation. The orientations of the limbs are determined by the inclination of the maximum compressive stress to bedding and by the friction angle between laminae in the limbs.

Applying the same method to the kink-wedge at the termination of the experimental kink bands in rubber strips shown in Figure 4, we measure θ_L on the right of about 37° and θ_L on the left of about 50° . According to the butterfly diagram in Figure 7, the solution is an inclination of the compressive stress of about $\beta = 7^\circ$ CCW from bedding and a friction angle on the order of $\phi = 46^\circ$. This is a high friction angle; perhaps too high, suggesting that the kink theory developed thus far does not apply to a kink wedge, which involves a décollement.

We can check this result with the friction angle determined for the conjugate kink bands, above the kink-wedge structure shown in Figure 4. There the angles between the limbs and the beds outside the kink bands are about 53° for both the left- and right-facing kinks. Since the angles are the same, we would assert that there was insignificant shear, that is, the maximum compression was roughly parallel to bedding ($\beta = 0$). This contrasts with the maximum compression inclined about $\beta = 7^\circ$ CCW according to the analysis of the kink wedge formed in the same rubber strips.

For maximum compression parallel to bedding and a kink angle of 53° , the butterfly diagram (Figure 7) indicates that the friction angle was about $\phi = 37^\circ$. Again, this contrasts with a friction value of $\phi = 46^\circ$ determined for the kink wedge formed in the same rubber strips.

Perhaps the kink theory developed thus far does not apply to a kink wedge, which involves a décollement. Examining the kink-wedge shown in Figure 4 carefully, we note that the kink limb facing left within the kink-wedge is inclined at about the same angle in the kink-wedge as in the left-facing kink-band above. The kink limb facing right within the kink-wedge, though, is inclined about 37° . Herein lies the discrepancy. It seems that we cannot relate, at this time, the angle of inclination of the more shallowly dipping limb to the friction angle or the inclination of the maximum compression.

On this basis, we suggest in Figure 9 three examples of idealized kink-wedge forms with the steeper limb at the theoretical ultimate locking angles. The slope of the shallower limb is given no significance here. As we have indicated above, the angle between the layering outside the kink band and the layering within the steeper limb of the kink-wedge can be determined by the ultimate locking angle, $\theta_L = \phi/2 - \beta$. If the maximum compression is parallel to layering outside the kink-wedge. In Figure 9A, the friction angle is zero and the locking angle of the short, wedge limb is 90° . In Figure 9B, the friction angle is 30° and the locking angle is 60° , and in Figure 9C the friction angle is 50° and the locking angle is 40° .

Figure 9. Three examples of idealized monoclinal kink wedges. The acute angle between tilted layers within the kink wedge and out side of it is defined by the ultimate locking angle α_u . The acute angle between the lower wall of the kink wedge and the layers outside the kink is defined by the ϕ . In all cases a layer parallel compression, σ_1 , is applied as well as right-lateral shear stress. (a) Contact friction angle of 0° produces an ultimate locking angle of 90° . (b) Contact friction angle of 30° produces an ultimate locking angle of 40° . (c) Contact friction angle of 50° produces an ultimate locking angle of 40° .

Proof that the kink-wedge Can Accommodate Translation of Adjacent Block

In order to show that the *kink-wedge* model is a viable model of décollement folding, we need to be able to show that the model can accommodate a reasonable displacement distribution in the detachment block. Indeed, we can show that the idealized *kink-wedge* model perfectly accommodates lateral *translation* of the detachment block and provides a terminal zone within which the translation is reduced to zero.

$$UT = vw = \text{Displaced Area}$$

$$\tan ? = v/w = \text{constant}$$

A.

$$\tan(?_L) = v/w_L = \text{constant}$$

$$\tan(?_R) = v/w_R = \text{constant}$$

$$UT = (1/2)v(T)[w_L(T) + w_R(T)]$$

$$Uy = (1/2)v(y)[w_L(y) + w_R(y)]$$

B.

Figure 10. The geometry of the kink-wedge model. (a) A symmetric kink-wedge. A horizontal displacement U is applied above a detachment. (b) An asymmetric kink-wedge. A horizontal displacement U above a detachment and a layer parallel shear stress are applied.

Figure 10 shows the geometry we will analyze. There is a block of thickness T above a detachment fault that terminates laterally. The block has moved from right to left an amount U . We select a coordinate, y , measured upward from the detachment. The double wedge is the terminus of the block. The layering within the left-hand wedge is inclined at a counterclockwise angle $?_L$. The layering within the right-hand wedge is inclined at a clockwise angle, $?_R$. Let w_L be the width of the wedge at a level, y , of the left wedge. Let w_R be the width at the same level in the right wedge. Let v be the height of the common side of the triangle, such that

$$\tan(?_L) = v/w_L \quad (2a)$$

$$\tan(?_R) = v/w_R \quad (2b)$$

According to Jakobson's rule (Johnson and Fleming, in prep., Chapter 2; Jakobson, 1952), the volume of material pushed above the general level in a gliding block (the gray triangular area at the top of Figure 10B) is equal to the displacement times the thickness of the block. Thus, his rule applied here to the asymmetric *kink-wedge* fold shown in Figure 10B would be,

$$UT = (1/2)v(T)[w_L(T) + w_R(T)] \quad (3)$$

Perhaps the case is clearer for the special case shown in Figure 10A, the symmetric *kink-wedge*. In order for the area to remain constant, according to Jakobson's rule, the material in the shaded triangle at the upper left hand end of Figure 10A must equal the area represented by the arrows on the right, UT .

In order for the internal layering of an asymmetric *kink-wedge* fold to follow Jakobson's rule, we ask whether we can generalize eq. (3) to,

$$Uy = (1/2)v(y)[w_L(y) + w_R(y)] \quad (4)$$

if $\tan(\theta_L) = v/w_L$ a constant, and $\tan(\theta_R) = v/w_R$ another constant. Here the width and height quantities are evaluated at the height, y .

We can write eq. (4) in two ways. Eliminating v , by substituting eqs. (1) and (2) into (4), we obtain

$$[w_L(y) + w_R(y)] = \sqrt[2]{2Uy \frac{\tan(\theta_L) \tan(\theta_R)}{(\tan(\theta_L) + \tan(\theta_R))}} \quad (5)$$

Similarly, eliminating $[w_L(y) + w_R(y)]$

$$v = \sqrt[2]{2Uy \frac{\tan(\theta_L) + \tan(\theta_R)}{(\tan(\theta_L) \tan(\theta_R))}} \quad (6)$$

Thus eq. (4) is satisfied if the geometry of the fold is given by eqs. (5) and (6). We see that the width and height of the *kink-wedge* increase upward with the square root of the height, y , above the detachment surface. Clearly this structure provides an entire balanced cross section; it even balances the cross section at every horizon a distance y above the detachment.

This completes the development of the *kink-wedge* mechanical model. We would simply comment that the kink hinges may be shown as sharp corners, but in the experiments and in the field the hinges are rounded. The sharp corner is used only to simplify construction of the gross form.

Criterion of Jakobson's Rule

We demonstrated in previous pages that Jakobson's rule is an integral part of the *kink-wedge* fold model. According to Jakobson's rule, the excess area at each horizon of the fold is equal to the uniform displacement times the height of the horizon above the detachment surface. In more general terms, the relation between the excess area and the height of a horizon is a linear function of the height.

Figure 11. Relation between excess area and height above the detachment for the Fault-Tip Pucker fold (Figure 2).

One might well ask, of course, whether this property is simply a property of any gliding block and therefore clearly nondiagnostic. It is not a property of any gliding block, as we can show with the fault-tip pucker fold model, which forms under the same conditions as the kink-wedge fold except the properties of the medium are homogeneous and isotropic. An example of the fault-tip pucker fold is shown in Figure 2. The relation between the excess area and the height, y , above the detachment surface is shown in Figure 11. In contrast to the relation for the kink-wedge fold, the relation for the fault-tip pucker fold is clearly nonlinear. In fact, the best fit polynomial to order three has the form,

$$\text{Area km}^2 = 63.434 \text{ km } y^3 - 24.113 \text{ km } y^2 + 4.8536 \text{ km } y + 0.8532 \text{ km}^2.$$

We note that there is an excess area of about 0.85 km^2 at the level of the detachment surface. Presumably this is reflected in the anticline that appears beneath the detachment surface (Figure 2).

It is clear in any case that not all folds that form at terminations of detachments display a linear relation between excess area and height above the detachment.

Thus, the balanced cross section is not an essential feature of the fault-termination folding mechanism in general. The fault-tip pucker fold (Figure 2), which is a type of fault-termination fold, does not conserve area in the sense noted above. The displacement on the detachment fault, times the distance above the detachment fault does not match the excess area at that level. This is because the medium below the detachment is able to deform similar to the medium above. This

results in an unbalanced cross section. In the kink-wedge model, the layers below the detachment are unable to deform and thus provide conservation of area.

ACKNOWLEDGEMENTS

We thank the Department of Energy, grant DE-FG02-98ER14886, supervised by Nick Woodward, for financial support of this research.

GEOMETRY AND FOLDING MECHANISMS OF THE PITCHFORK ANTICLINE

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INTRODUCTION

This paper is a detailed study of the Pitchfork anticline, one of the smaller, thrust-fold structures of Stone (1983a). It occurs as an isolated fold amongst other isolated and adjacent anticlines on the west flank of the Bighorn Basin (Figure 1). It is approximately 40 km south of Cody and 20 km west of Meeteetse, Wyoming (Figure 2). Pitchfork anticline is in Park County, Wyoming, on the west flank of the Bighorn Basin.

The Bighorn Basin is one of many basins in the Western Cordillera of the United States that formed during the Laramide orogeny, during which the Rocky Mountain foreland was partitioned into isolated structural basins by incipient Precambrian uplifts. It is both a structural and topographic basin and is bounded on three sides by the Bighorn Mountains to the east, the Absaroka and Beartooth Mountains to the west and the Owl Creek Mountains to the south. The trend of the Bighorn Basin is to the northwest.

The Bighorn Basin is underlain by some very large faults that dip away from the axis of the basin and therefore lift the ground surrounding the basin. The thrust under the Rattlesnake Mountain fold is one of these on the west side of the basin. Another is the Oregon Basin thrust, which offsets the Cambrian/Precambrian contact about 6 km. It is overlain by the Oregon Basin fold. The fold and thrust have accommodated about 6 km of vertical uplift and 4.5 km of horizontal shortening (Stone, 1953, p. 181). The thrust dips into basement rocks with a dip of about 45° to the west. The thrust under the Rattlesnake Mountain apparently dips the same way.

Figure 1. Geologic map of Wyoming, showing the location of the Bighorn Basin and other Laramide age basins. Cz = Cenozoic; Mz = Mesozoic; Pz = Paleozoic; pC = Precambrian; Tev = Tertiary Volcanics; Qv = Quaternary Volcanics

The Bighorn Basin contains many small anticlines, a few km long, distributed mostly around the rim of the basin. The anticlines trend west to northwest and are generally asymmetric. The faults underlying the small anticlines generally dip to the east on the west side of the basin and to the west on the east side of the basin. As a result, the faults beneath the small anticlines are "back-thrusts" relative to the large thrusts. The small thrusts dip toward the basin whereas the large

thrusts dip away from the basin. In map view, the small anticlines produce arcuate patterns, generally concave toward the center on the basin. Thus, on the west side of the basin the arcuate fold are concave toward the east and on the east side of the basin the arcuate folds are concave to the west. These smaller folds are the "thrust-fold structures" of Stone (1985a, 1985b). Stone (1983b) indicates that the thrusting and folding were accomplished by NE-SW compression of the crust.

Figure 2. Geologic map of the Bighorn Basin, showing the location of Pitchfork anticline and other basin flanking folds.

The stratigraphic units involved in the folding of the Pitchfork anticline range in age from Precambrian to Tertiary and are about 4 km thick (Figure 4). The Paleozoic section is about 1 km thick. Much of it, from the Mississippian Madison Limestone to the Cambrian formations, consists of a thick section of carbonates and sandstones. The Mesozoic section is about 1.8 km thick and consists mainly of alternating layers of shale and sandstone; there is a higher percentage of shales than sandstones. Most of the Mesozoic section is Cretaceous, about 1.4 km thick, and half of the Cretaceous section is the Cody Shale. The Triassic Chugwater Formation comprises approximately 290 m of uniform siltstone and sandstone. The Tertiary section, below the Fort Union, is about 1 km thick. The youngest unit involved in the folding is the Fort Union, which has an angular unconformity at its top.

A.

B.

Figure 3. Setting of the Pitchfork anticline on the west flank of the Bighorn Basin. (a) Geologic map of the west flank of the Bighorn Basin showing the location of Pitchfork anticline and adjacent folds (b) Shallow surface cross section across Fourbear and Pitchfork anticlines (Pierce, 1941). The profile shows the regional dip on the west flank of the Bighorn Basin to be 5° E between Fourbear and Pitchfork anticlines.

In its simplest terms, then, the Bighorn Basin, itself, is formed on the east, west and south by very large thrust-folds, underlain by reverse faults that dip about 45° outward from the basin and penetrate deeply into Precambrian basement rocks. The large faults lift the surrounding ground relative to the rocks within the basin. The rocks within the basin are roughly flat-lying except where disturbed locally. Within the basin are smaller thrust folds that are underlain by thrust or reverse faults that dip toward the center of the basin and that penetrate into the Precambrian

basement rocks. The smaller faults lift ground in the basin relative to the ground surrounding the basin.

BigHorn Basin, Wyoming			
Tertiary	Willwood	Tw	
	Fort Union	Tfu	
Cretaceous	Lance (335m)	Kl	Soft
	Meeteetse (335m)	Km	
	Mesaverde (365m)	Kmv	Stiff
	Cody (640-765m)	Kc	Soft
	Frontier (150-170m)	Kf	
	Mowry (120m)	Kmr	
	Thermopolis (185m)	Kt	
	Cloverly & Morrison (185m)	KJcm	
Jurassic	Sundance (120m) Gypsum Spring (30m)	Jsg	
	Triassic	Tr c	
Permian	Phosphoria (25-50m)	Pp	Stiff
Pennsylvanian	Tensleep (50-65m)	IPt	
	Amsden (70-90m)	MIPa	
Mississippian	Madison (215-245m)	Mm	
Devonian	Three Forks (60m)	Dt	
Ordovician	Bighorn (120m)	Ob	
Cambrian	Gallatin (130m) Gros Ventre (225m) Flathead (35-45m)	C	
	Precambrian	pC	

Figure 4. Stratigraphic column of the Bighorn Basin, Wyoming, showing the divisions between mechanically soft and stiff layers.

Stone (1983b) suggests that smaller thrust-fold structures are initiated by thrust slip in a granitic, Precambrian basement and propagate into the overlying Paleozoic carbonates and into the Mesozoic clastics. The amount of slip decreases upwards but the amount of folding increases upwards, so the shortening accomplished by faulting alone in the basement is accomplished by folding alone in the Mesozoic clastics.

To the northeast of the Pitchfork anticline, Spring Creek anticline trends 40° northwest and seems to converge with Pitchfork anticline at its north end (Figure 3A). To the southeast, North and South Sunshine anticlines appear *en echelon* arrangement with respect to Pitchfork anticline. The

trend of North and South Sunshine is similar to Pitchfork anticline, but it steps to the southeast. A cross section constructed by Pierce (1941) shows the dip of the broad west flank of the Bighorn Basin in the vicinity of Fourbear anticline east to Pitchfork anticline to be approximately 5° east (Figure 3B).

FORM OF THE PITCHFORK ANTICLINE

Forms Visible at and near the Ground Surface

The anticlinal folds in the Bighorn Basin are known for their classic, arcuate, crescent moon shapes at the ground surface.

Figure 7. Geologic map of the west flank of the Bighorn Basin, showing the location of Pitchfork anticline, Horse Center anticline and other well-exposed structures.

One can learn much about the forms of the folds by observing and mapping surface exposures. The northern and southern ends of the fold are symmetric; most of the fold is asymmetric, with a steep west limb. Figure 8 is an oblique aerial view from north to south along the longitudinal axis of the Pitchfork anticline. In the foreground, the Frontier Sandstone (Kf) dips 50° E on the east and 65-70° W on the west. The fold is nearly symmetric. Farther south, the west limb remains steep, with dips of 65-75°, whereas the east limb becomes shallow, with dips of 30-35°, so the fold is strongly asymmetric.

The *crestline*, which is a line tangent to the point of minimum radius of curvature along the longitudinal axis of the anticline, is exposed in the high ground all along the fold. It is arcuate in

plan, concave to the east. Shown even more clearly on the geologic map (Figure 9), the outcrop pattern of the Frontier narrows and the crestline increases in curvature toward the south.

Figure 8. Oblique aerial view from north to south of Pitchfork anticline along the crestline of the fold. The trace of the crestline of the anticline is shown with a faint, dashed, white line. Kc = Cody Shale; Kf = Frontier Sandstone; Kmr = Mowry Shale.

Many oil wells have been drilled through the Frontier in the broad, smooth hollow area east of the crestline. This area is covered with Tertiary terrace deposits throughout most of the length of the fold (Figure 8 and Figure 9). We will see that the crestlines generally migrate eastward with increasing depth in the Pitchfork anticline.

Out of the view of the photograph (Figure 8) but visible on the geologic map (Figure 9 and Plate 1) are two neighboring synclines, both plunging to the SE. The syncline to the west of Pitchfork anticline has a troughline—a line tangent to the radius of minimum curvature along the longitudinal axis of the syncline? about 2 km west of the anticline and is roughly parallel to the crestline. The troughline of the syncline to the east diverges away from the anticline toward the south, ever increasing the spacing from the crestline to the troughline.

Figure 10 is an oblique aerial view from south to north along the crestline of the fold. The Mowry Shale is exposed in the center of the anticline and the resistant Frontier sandstone forms the flanks of the anticline. The Frontier Sandstone forms a topographic ridge on the west flank of the fold and dips 70-75° W, whereas on the east flank the outcrop pattern is more eroded but still ridge-like and dips 30-35° E. The Cody Shale underlies the ground with

Figure 9. Geologic map of Pitchfork anticline, Park County, Wyoming.

low relief to the west. Also visible in Figure 10 is the curvature of the fold crestline, concave to the east.

Even at ground level one can observe some changes in fold shapes in profiles as a function of stratigraphic position. Figure 11A shows a sharp crest or hinge within a well-exposed shale bed of the Mowry. In the hillside a short distance south of the view of Figure 11A, the overlying contact between Mowry Shale and Frontier Sandstone is rounded, with a much greater radius of curvature.

The contact is much like that between Mowry and Frontier at the north end of Pitchfork anticline (Figure 11B). The contact between the Mowry and Frontier a few hundred m north of the location of Figure 11B is also rounded (Figure 11c) and is structurally lower than at the site of Figure 11B. These outcrops show that the crest of the fold can be sharp at one horizon and quite rounded at another nearby.

Figure 10. Oblique aerial view to the north along the crestline of Pitchfork anticline. Kc = Cody Shale; Kf = Frontier Sandstone; Kmr = Mowry Shale.

A.

B.

C.

Figure 11. Comparison of fold hinges at different locations along the anticline and stratigraphic positions. (a) Sharp fold hinge within a well-exposed bed of the Mowry Shale at the south end of the anticline. (b) Broad fold hinge marked by a prominent sandstone bed at the Frontier/Mowry contact. This contact is about 20 m above the marked bed in (a). (c) Broad fold hinge marked by the same Frontier sandstone bed shown in (b). The Frontier/Mowry contact at this location is structurally lower than at location (b). Photograph (b) was taken from the location of (c) and vice versa.

All the information presented thus far is summarized in four shallow surface cross sections representing the entire length of the Pitchfork anticline (Figure 12). The cross sections are constructed normal to the crestline of the fold and represent only a very thin layer of geology that is known from field observations. The geometric information includes the observations presented in photographs as well as contacts and other geometric data presented on the field map (Figure 9). At this stage of data presentation, extrapolations are minimized except for the area above the ground surface. There the folds are shown schematically in rocks that were present during the time of folding.

The following conclusions are based on the results presented in the surface cross sections:

- ? The asymmetry of the anticline changes along the crestline. The anticline becomes more symmetric toward the north.
- ? The crestline of the anticline is arcuate in plan, concave to the east.
- ? The dip of the west limb of the anticline is steepest near midlength of the fold. It shallows gradually to the south and abruptly to the north.
- ? The crestline of the anticline plunges to the north and south.
- ? The troughline of the syncline to the west is parallel to the crestline of the anticline. The troughline of the syncline to the east converges toward the north along the crestline of the anticline.
- ? The hinge of the anticline changes peakedness in different stratigraphic horizons.

Figure 12. Four shallow surface cross sections, constructed normal to the crestline of Pitchfork anticline. Transect A-A' shows formations projected above the ground surface to show the actual thickness of strata present during the time of folding. The outcrop pattern of the Frontier Sandstone is shown by a gray pattern to indicate the position of transects along the length of the fold.

Deep Cross Sections Based on Surface Data and Well Logs

There are at least 100 records of wells drilled in the Pitchfork oilfield ranging in age from 1926 to 1988. The borehole data are unevenly distributed within the fold. Most of the boreholes are near midlength and in the low-lying, hollow area east of the crestline of the fold (Figure 8). The information available is different for wells drilled before and after about 1967, but there was acceptable information for 95 of the boreholes. For about half the wells, drilled before 1967, tops of formations are provided on scout cards or they can be determined from electric logs. The scout cards provide elevation and horizontal position (x - y - z) of the drilling platform, and depths to tops of formations. For some boreholes, dipmeter surveys are also available. For wells drilled after about 1967, the record provides not only x - y - z position of the drilling platform and information about the tops of formations but also directional surveys which are used to determine the magnitude and direction of well-bore drift. For 43 of these records we could determine bottom hole x - y - z locations.

The borehole data were generalized by constructing structural contour maps, which show changes in fold geometry in plan. Although we constructed structure maps on the tops of the five formations, only two are included here (for others, see Durdella, 2001): During contouring, wells with corrected x - y locations were weighted with more importance than wells without a drift correction.

Figure 15A is a structure map constructed on the top of the Cretaceous, Muddy Sandstone, which is a member of the Thermopolis Shale. It is 30 m above the base of the Thermopolis. The crestline of the fold is arcuate along its length, concave to the east, and doubly plunging. The fold form is asymmetric, and the hinge of the fold is narrow near the crestline. The Pennsylvanian Tensleep Formation is the lowest unit drilled regularly at Pitchfork oilfield and provides the lowest contoured surface. Figure 15B shows the structure map at this horizon. The doubly plunging, arcuate crestline, concave to the east, is similar to that of the overlying horizons. The anticlinal closure shifts to the north, as in the Chugwater and Phosphoria, and the hinge of the fold is broad like the overlying Phosphoria.

Figure 15A. Muddy Sandstone structure contour map. Datum is the top of the Muddy Sandstone, which is within the Thermopolis Shale and lies 30 m above its base.

Figure 15B. Tensleep Sandstone structure contour map. Datum is the top of the Tensleep Sandstone.

The following conclusions are based on the results provided by the structural contour maps:

- ? At all horizons, the crestline of the fold is arcuate, concave to the east, and doubly plunging.
- ? At all horizons the fold is asymmetric.
- ? The hinge of the anticline changes from narrow near the surface to broad at the Chugwater. The hinge at the two horizons below the Chugwater is likewise broad, but not as broad as the Chugwater.
- ? At and below the Chugwater Formation, the anticlinal closure shifts to the north.

In plan view one can see the changes along the longitudinal trend of the fold. Figure 16 shows a plan view of the crestlines of the fold at different stratigraphic levels. It was constructed by tracing the crestline of the fold along the longitudinal axis for each horizon indicated in Figure 16. Crestline depths are indicated by a gradational change in the grayscale of lines from lightest gray at the ground surface to black at the top of the Tensleep Sandstone.

The plan view shows the following:

- ? The crestline of Pitchfork anticline is consistently arcuate, concave to the east, throughout the depth shown.
- ? The crestline of the fold migrates approximately 500 m east through the depth of about 1000 m.

Figure 16A. Plan view of Pitchfork anticline showing the of fold crestlines with changing depth.

The longitudinal profile of Pitchfork anticline allows one to see changes along the longitudinal axis of the fold as viewed from east to west (Figure 16B). The crestlines of the fold are projected horizontally to a vertical plane to show their respective stratigraphic positions. In this view, the Precambrian basement is plotted approximately according its location in well number 22-X. It is shown only to give a reference to the depth of the view, since it is improper to show the ground surface in this type of view.

The longitudinal profile indicates the following:

- ? The fold crestline is convex upward and, therefore, plunges to the north and the south.

The curve of the crestline is greatest at depth.

Figure 16B. Longitudinal profile of Pitchfork anticline, showing the change of the fold crestlines with depth.

Figure 17 shows four deep cross sections through the central half of the Pitchfork anticline based on a combination of surface data presented in the shallow cross sections and the subsurface borehole data. The resulting figure is the most nearly accurate reconstruction I can make based on these data sets. I have minimized interpretation and extrapolation. In these cross sections, some wells are projected to the line of transect and are indicated as such by the projection distance shown above well numbers. Data of different certainties are indicated with symbols. For example, a borehole represented in a cross section by a solid line indicates that top and bottom well locations are known and the formation tops are accurate. A borehole represented by a solid line near the surface and a dashed line at depth indicates that well bottom location is unknown. The formation tops are accurate where the line is solid and are of unknown accuracy where the line is dashed.

Short lines near the ground surface and along boreholes represent the dips. The dips are known where the dips are shown and inferred elsewhere.

The tops of formations are shown only within the limits of borehole data and are shown with solid lines (Figure 17).

The following conclusions are based on the results presented in the deep cross sections:

- ? The radius of curvature of the hinges of formation tops is smaller above the Chugwater than below. The folding is tighter near the ground surface.
- ? There typically is a layer-parallel fault within the Sundance Formation where the transition in curvature occurs.
- ? The crestline of the formation tops migrates eastward with increasing depth.

The dip markers shown along the borehole 22-X within the Cambrian formations indicate that the beds dip steeply to the west near to top and dip shallowly near the base. It is unclear from the well log as to a cause for the discrepancy in dip magnitudes.

Figure 17. Deep cross sections based on surface data and well logs. Numbers above each well indicate the well number and the distance it was projected to the line of transect.

Deep Cross Section Based on Surface, Well Log and Seismic Data

Figure 18 shows the interpreted, time-migrated, seismic profile (Stone, 1993), the location of which is shown on the geologic map (Figure 9) as transect G-G'. The profile was interpreted as follows. Overlaying well number 22-X, the deepest well, on the seismic profile located the top of the Precambrian basement. A prominent marker representing this contact was then traced along the limb to the east. The stratigraphic section was used to locate markers that correspond to formation tops at the location of well 42-12 at the far east side of the profile. A marker representing the top of the Chugwater Formation was traced from the east limb to the west limb, where the stratigraphic

section was used to locate the top of the Precambrian basement. Solid white lines indicate where the location of formation tops are certain, dashed where inferred and left blank where uncertain.

Figure 18. Interpreted, time-migrated (1200% Vibroseis, 1982) seismic profile for Pitchfork anticline (modified after Stone, 1993). Dashed lines represent areas of uncertainty in the interpretation.

Figure 19 is the most nearly accurate and deepest reconstruction of a profile through Pitchfork anticline. Its reconstruction is the result of combining surface and subsurface observations from all three data sets presented above. Two types of dashed lines are shown on the profile in Figure 19. The first is where formation tops have been extrapolated to the sides and above the ground surface and the second is where they have been interpolated at depth near the fault.

Figure 19. The most nearly accurate reconstruction of Pitchfork anticline based on the combination of surface data, well logs and a seismic profile. Formations are projected above the ground surface to show extent of strata incorporated in the folding of Pitchfork anticline.

The following conclusions are based on the results presented in the cross section:

- ? None of the limbs of the fold is planar
- ? The radius of curvature is 1500 m on the western limb of the anticline from the top of the Frontier Sandstone (Kf) to the top of the Morrison Formation (KJcm). It gradually decreases to 100 m at the top of the Madison Limestone (Mm).
- ? The radius of curvature is 80 m at the hinge in the section of the anticline from the top of the Mowry Shale (Kmr) to the top of the Thermopolis Shale (Kt) above the fault in the Sundance Formation (Jsg). It is 430 m at the top of the Chugwater Formation (Trc) and reduces to 240 m at the top of the Madison Formation (Mm).

- ? The radius of curvature is 5100 m on the east limb from the top of the Frontier Sandstone (Kf) to the top of the Madison Limestone (Mm), where it increases to 7200 m at the top of the Precambrian (pC).
- ? The stratigraphic throw of the Pitchfork anticline nearly doubles from 750 m at the top of the Precambrian basement (pC) to 1400 m at the top of the Frontier Sandstone (Kf).

With the addition of the seismic data combined to produce the profile in Figure 19, one can describe the asymmetry of the fold. To quantitatively describe the asymmetry of the fold we will use the definition derived by Johnson (1993), in which limb lengths are defined as the distance, lambda (λ), measured parallel to the medial plane, as shown in Figure 20A, from trough to crest and from crest to the next trough. Thus, λ_L is the length of the left limb and λ_R is the length of the right limb as shown in Figure 20B. The asymmetry of a folded surface is defined in terms of relative lengths of the left and right limbs, λ_L and λ_R :

$$\dot{A} = (\lambda_L - \lambda_R) / (\lambda_L + \lambda_R) \quad (1)$$

This quantity, $\dot{A}$ (pronounced *oh* in the Swedish language), ranges from 0 for symmetric folds to $\dot{A} = 1$. If $\dot{A} = 1$ the limb length, λ_R , is zero. Note in Figure 20B, that if the vergence or tilt of the fold is to the left rather than to the right as is shown, a negative value of $\dot{A}$ is produced. Such is the case at Pitchfork anticline if viewed from south to north parallel to the fold axis.

Applying this definition of asymmetry to Pitchfork anticline, the asymmetry of the fold along the contact at the top of the Chugwater Formation is $\dot{A} = 0.67$, which is highly asymmetric.

Figure 20. Defining diagram for asymmetry of folds. (a) three-dimensional fold showing the projection of the medial plane through the fold. (b) Diagram showing the definition of the terms used to describe the asymmetry of a fold.

THE MECHANICAL FAULT-RELATED FOLDING MODELS

Now that all the field data have been used to determine the plan and cross-section shapes of the Pitchfork anticline, we will use relevant mechanical models in order to deduce the folding mechanisms responsible for formation of the Pitchfork anticline and even estimate values of some of the parameters that quantify the mechanisms. We have considered three mechanical models of fault-related folding: *Gang/listric-Fault Folding*, *kink-wedge Folding* and *Forced-Folding*. Mechanical models are characterized by their theoretical base, which includes the complete set of mechanical equations required to describe the system, including force and moment equilibrium, compatibility of strains, boundary conditions and rheology.

*Gang/listric-Fault Fold*⁷

For the case of the *Gang/listric-Fault Fold* model, the rheology is isotropic viscosity or elasticity. The slip of faults is represented by slip along rectangular dislocations representing fault segments. The distribution of dislocations and the slip on the dislocations produces deformations in the host medium.

The *Gang/listric-Fault Folding* computer program, *SeGaLi FaRF*⁸, simulates folding produced by rectangular fault planes arranged to form three-dimensional fault surfaces with variable geometry beneath a stress-free surface, representing the ground surface. The faults may be added sequentially, in which case the folding occurs during fault propagation. Or they may be added all at one time, in which case the model produces fault-arrest folds. The power of the *SeGaLi FaRF* program is its ability to position fault segments in a myriad of arrangements in order to form truly three-dimensional fault-related fold structures. For example, by placing fault planes edge-to-edge, the program simulates folding associated with listric and antilistric faults. Although the *SeGaLi FaRF* program is based on well-known theory in geophysics (Mansinha and Smylie, 1971), it is a forward-modeling program designed for simulation of fault-related folds. The geophysics programs are for inverse modeling of faults responsible for observed patterns of displacement during a slip event.

In the *SeGaLi FaRF* model, the dimensions and positions of individual fault segments are entered according to rectangular geometry and *x-y-z* translations respectively as shown in Figure 21A. The magnitude of displacement on a fault segment is entered, as well as the sense of slip, which can be any combination of strike-slip and dip-slip. The fault segments are arranged in a map area designed by the user. The grid spacing in the map area governs the accuracy of a contour map of vertical displacements. An example is shown in Figure 21B. In addition, a series of four parallel cross sections are constructed through the map area in order to show the structure at depth (Figure 21C).

However useful *SeGaLi FaRF* may be for simulating fault-related folding of various types, it provides an incomplete mechanical solution. This is because deformation is produced without accounting for the cause of the slip on the dislocations. In this respect it is like stability analysis of

⁷ Program may be downloaded from www.eas.purdue.edu/fauxpli

⁸ *ibid.* Program written by A.M. Johnson with help of Anthony DeNicola in MS VisualBasic 1997/2000.

landslides. A basal slip surface is formed, somehow. The stability analysis assumes uniform slip over an element of a slip surface and then proceeds to draw conclusions based on the consequences of the slip. There really is no satisfactory three-dimensional analysis of faulting that could be used to study fault-related folding. In short, the dislocation representation of faulting is the best three-dimensional mechanical model of faulting available at this time and we use it to study how a medium deforms, given a particular arrangement of faults and distribution of slip.

A.

B.

C.

Figure 21. An example of the Gang/litric-Fault Fold model. (a) Block diagram showing the arrangement of input fault segments. (b) Contour map produced at the free surface after dislocation along the fault segments. (c) Serial cross sections produced by the model show change in fold geometry along the crestline of the fold.

The *Gang/litric-Fault Fold* model will be used to address features of the Pitchfork anticline such as: the interaction of Pitchfork anticline with adjacent structures, the arcuate shape of the longitudinal axis in plan and profile views and the doubly-plunging fold crestline.

Forced-Fold

The *Forced-Fold* model, developed by K. Johnson (2000) with folding theory (Johnson and Fletcher, 1994), is for material with anisotropic viscous or power law rheologies. A velocity distribution is applied to the boundary between a rigidly displaced basement block and an overlying, passively deformed sedimentary cover layer. The system is allowed to slip along a detachment fault in the basement. The forms of the folds depend on the anisotropy of the medium, on the boundary conditions, and on the type of basement faulting.

Figure 22 shows a diagram of the model. Boundary conditions for the model include a velocity distribution applied to the basement/cover contact and a traction-free upper surface. The parameters of the model are thickness of the sedimentary cover, T , dip of the fault, θ , length, L , and angle, ϕ , of the fault backlimb and the viscosity ratio of the anisotropic sedimentary cover, η_n/η_s . The contact between basement and sedimentary cover may also vary from perfectly bonded to frictionless. Equally or custom spaced passive markers are drawn on the sedimentary cover layer in the model to simulate bedding. These markers, though, are passive; they do not actually represent mechanical layering.

Figure 22. Defining diagram for the Forced-Fold model. T = sedimentary cover thickness; L = length of the rotated backlimb, α = angle of backlimb rotation; θ = fault tip angle with respect to the basement/cover boundary; μ_n = normal viscosity in the sedimentary cover; μ_s = shear viscosity in the sedimentary cover. The upper boundary behaves like a free surface.

By varying the viscosity ratio, μ_n/μ_s , the sedimentary cover may be treated as an isotropic medium, $\mu_n/\mu_s = 1$, which would simulate a single deformable layer overlying displaced, rigid basement blocks. In contrast, an anisotropic medium, $\mu_n/\mu_s > 1$, enables the sedimentary cover sequence to behave like a mechanical multilayer, which simulates folding of alternating thick and thin layers. Lastly, inverse anisotropy, $\mu_n/\mu_s < 1$, causes the sedimentary cover to deform like a multilayer with inclined layering or a power-law material undergoing overall shortening. The contact between the basement and the sedimentary cover can be varied from completely bonded to perfectly frictionless by changing the ratio of viscosity of a thin film. Fault geometry and sense of displacement can be designated as either planar or listric and normal or reverse.

In the *Forced-Fold* model we are able to change the rheology of the sedimentary cover by varying the viscosity ratio, which is defined as the ratio of viscosity for layer-parallel shortening or lengthening, μ_n , to the viscosity for layer-parallel shearing, μ_s . A viscosity ratio of 1 allows the sedimentary cover to behave as an isotropic medium or a homogenous single layer. By increasing the viscosity ratio, the sedimentary cover begins to behave more like a mechanical multilayer or as an anisotropic medium.

A series of runs with the *Forced-Fold* model illustrates effects of a few key variables on the form of the fault-related folds involving a basement cover and a fault cutting basement rocks. In all cases the thickness of the sedimentary cover and the amount of slip on the basement fault are realistic. Figure 23 shows forms of folds associated with planar and listric normal or reverse faults

in cover rocks that range from inversely anisotropic, $\eta_n/\eta_s = 0.3$, isotropic, $\eta_n/\eta_s = 1.0$, to highly anisotropic, $\eta_n/\eta_s = 10$.

The fault-related folds over normal and reverse planar faults closely resemble each other but do not reproduce the essential features of the Pitchfork anticline (Figure 23). Only the pattern of layering in the listric reverse fault in highly anisotropic material closely resembles that in the Pitchfork anticline. As in the Pitchfork anticline, the thinned layers in the steep limb and the backlimb rotation resemble those in the Pitchfork anticline. The remaining two main differences between the field observations and the theoretical predictions are, in the actual fold, the hingeline is not vertical; it dips east. In the theoretical fold the hingeline is vertical. Also, in the actual fold, the amplitude of folding increases upward. In the theoretical fold the amplitude decreases upward.

Not only are the mechanical properties of the sedimentary cover important, but also the interaction of the basement-sedimentary cover contact. The *Forced-Fold* model is able to deal with a basement-sediment cover contact that is bonded or frictionless.

Stearns (1978) reports that the contact between the sedimentary cover and the underlying basement behaves like a non-welded contact. This is due to the ductile properties of the Cambrian shales, which lubricate the boundary between the two mechanical units and allows the sedimentary cover to move freely over the basement. Figure 24 shows a comparison of a friction free basement/sediment contact, (A), and a perfectly bonded basement-sediment contact, (B). The attributes for these two runs consider a sedimentary cover with a viscosity ratio of 3 displaced over a listric reverse fault. In Figure 24A, the basement-cover contact is free slip and in Figure 24B, the basement-cover contact is perfectly bonded. Interestingly, the geometry of the fold form does not differ between the two scenarios, although the displacement vectors, represented by the black arrows, indicate that the bonded condition imparts a component of horizontal compression in the layers overlying the fault tip. This is not so in the case of the free slip condition, where displacements are isolated to only a vertical component. For the case of Pitchfork anticline the contact between basement and sedimentary cover will be treated as a bonded contact, since the hinge line of the anticline is tilted to the west, indicating a component of horizontal compression.

Figure 23. Comparison of viscosity ratio, μ_n/μ_s , to various fault geometries.

A. Friction free contact between basement and sedimentary cover.

B. Bonded contact between basement and sedimentary cover.

Figure 24. Comparison of fold geometry produced by varying the basement-sedimentary cover contact property. Displacement vector fields are indicated by the black arrows and show relative movements of the sedimentary cover with respect to the basement. (a) Relative displacements of the sedimentary cover are mainly vertical for the case of a free slip basement/cover contact. (b) Relative displacements of the sedimentary cover are inclined for the case of a bonded basement/cover contact. Note: for the condition of free slip, no layer parallel compression is applied to produce the inclined displacement vectors.

Another interesting feature produced by the *Forced-fold* model has to do with the migration of the hingeline of the fold with varying degrees of anisotropy or viscosity ratio. We have already seen that sedimentary cover with a viscosity ratio of 1 results in a fold amplitude that rapidly attenuates above a fault tip (Figure 23). And conversely, high anisotropy produces the opposite result: fold amplitude is sustained upward. In addition to this result, the viscosity ratio is responsible for migrating or tilting the hingeline of a fold. At a viscosity ratio of 1 the hingeline of the fold is set back from the crest in the direction of the backlimb as in Figure 25A. As the viscosity ratio is incrementally increased from 1 to 4, the crestline rotates from its tilted position in Figure 25A to a more nearly vertical position in Figure 25D. It is also translated toward the crestline with increasing anisotropy.

This hingeline migration and tilt, although interesting in the models, does not suitably produce the westward hingeline tilt observed at Pitchfork anticline. The increase in anisotropy of the *Forced-Fold* model rotates the hingeline from the backlimb to only a vertical position and not toward the forelimb, as is the case at Pitchfork.

Figure 25. Varying degrees of anisotropy cause the crestline of the fold to rotate to vertical from a tilted position toward the backlimb and translated toward the crestline.

The Fault-Tip Pucker Fold

A type of fault-related folding that might be relevant to the formation of the Pitchfork anticline is the folding that occurs at the termination of a blind detachment fault (Berger and Johnson, 1982). The essential mechanism of folding is the abrupt lateral termination of slip along a detachment fault. Because of the termination, the material piles up near and around the termination and forms an anticline. The anticline has quite different forms, depending on the properties of the host rock and the shape of the fault in the termination area. Here we will consider only two, the *Fault-tip pucker* fold that occurs if the host rock is nearly homogeneous and isotropic, and the kink-wedge

fold that occurs if the host rock has properties that favor the formation of kink bands under layer-parallel compression (Johnson and Durdella, 2001).

Fault-Tip Pucker folds form where a detachment surface terminates near the ground surface. They can be produced mechanically with a dislocation theoretical model, as shown in Figure 26. (see www.eas.purdue.edu/fauxpli) The dislocation is placed near the ground surface and allowed to slip. A nearly identical form can be produced with a solution for mode II slip on a fracture parallel and near a free surface.

In either case, the layering shown in Figure 26 is passive—it is not mechanical layering. The host material is isotropic and homogeneous. The termination of the slip causes the pucker to form, and the free surface causes the pucker to increase in amplitude upward. If the overburden were much thicker, the amplitude would decrease upward, so the form of the fold reflects to some extent the thickness of the overburden.

Figure 26. Fault-Tip Pucker fold produced by a laterally terminating, horizontal detachment with the Gang/Istria-Fault Fold model.

Whether close to or distant from the free surface, though, an anticline occurs below and its crest is in back of the tip of the terminating fault. This is one of those “secondary” structures that should be of interest in petroleum exploration.

The kink-wedge Fold

The kink-wedge fold forms under the same conditions as the Fault-Tip Pucker fold, but the properties of the medium are quite different. The kink-wedge fold forms in layered materials that can slip but with some difficulty. The contact strength between layers must be overcome in order for the kink-wedge fold to form (Johnson, 1977; K. Johnson et al., 2001).

The theory of the kink-wedge is developed by Johnson and Durdella (2001). They show that, where there are steeply dipping and a shallowly dipping limbs, the inclination of the steeply dipping limb appears to be close to the locking angle (Reches and Johnson, 1976). The angle of inclination of the more shallowly dipping limb, however, is not readily related to the friction angle or the inclination of the maximum compression. On this basis, we suggest in Figure 27 three examples of idealized kink-wedge forms with the steeper limb at the theoretical ultimate locking angles. The slope of the shallower limb is given no significance here. The angle between the layering outside the kink band and the layering within the steeper limb of the kink-wedge can be determined by the ultimate locking angle, $\theta_L = \theta/2 - \theta_f$ if the maximum compression is parallel to layering outside the kink-wedge. In Figure 9A, the friction angle is zero and the locking angle of the short, wedge limb is 90° . In Figure 9B, the friction angle is 30° and the locking angle is 60° , and in Figure 9C the friction angle is 50° and the locking angle is 40° .

A.

B.

C.

Figure 27. Three examples of idealized monoclinal kink wedges. The acute angle between tilted layers within the kink wedge and out side of it is defined by the ultimate locking angle α_u . The acute angle between the lower wall of the kink wedge and the layers outside the kink is defined by the ϕ . In all cases a layer parallel compression, σ_1 , is applied as well as right-lateral shear stress. (a) Contact friction angle of 0° produces an ultimate locking angle of 90° . (b) Contact friction angle of 30° produces an ultimate locking angle of 60° . (c) Contact friction angle of 50° produces an ultimate locking angle of 40° .

For further details of the kink-wedge fold, see Johnson and Durdella (2001), where we prove that the kink-wedge fold can accommodate lateral translation of a block on a décollement.

SIMULATION OF THE PITCHFORK ANTICLINE

Transverse Profile

In this section, we will be to compare the conditions and fold forms deduced from the study of the Pitchfork anticline with those of the mechanical models. We have already concluded that the gross form of the Pitchfork anticline in the sedimentary rocks above Precambrian basement is similar to the form of the asymmetric, kink-wedge model. An exception is that the Pitchfork anticline does not terminate at a tip but rather at a basement block of some type. Indeed, the fold form at the level of the Precambrian basement—as far as we can determine—resembles the backlimb rotation of the *Forced-Fold* model with a listric fault joining a deeper detachment fault. Thus we already would anticipate that the Pitchfork anticline is a result of at least two folding mechanisms, one in the basement and one in the sedimentary cover.

We have discussed the kink-wedge model in detail in a companion manuscript (Johnson and Durdella, 2001). We show there that, one of the characteristics of the kink-wedge mechanical model is that, in its idealized form, the steeper, forelimb of the fold tends to slope at an angle, relative to layering on either side that closely approximates the ultimate locking angle, which depends on the friction angle resisting slippage between layers.

The forelimb of Pitchfork anticline maintains an angle of approximately 50° to 60°. If these angles correspond to locking angles of a kink band then we would deduce a friction angle of 30° to 35°. This certainly is a reasonable friction angle for most rocks (e.g., Johnson, 1995). Thus, the idealized kink-wedge fold that should most closely resemble the Pitchfork anticline is shown in Figure 28B. The upper part of the kink-wedge fold form closely resembles the Pitchfork anticline (Figure 19). The kink-wedge mechanism appears to explain the increase in amplitude and width of the fold upward, as well as the rotation of the hingeline toward the forelimb. If the Pitchfork anticline is similar to a kink-wedge, though, the excess area should be linearly related to height within the fold. The kink-wedge conserves area and the area produced by the folding at the termination of a detachment surface increases linearly from zero at the base to maximum at the top of the wedge. If this model is to accurately apply to Pitchfork anticline, then we should see a similar area relationship in the fold.

We calculated the displaced areas (Table 1), that is, the excess area of the anticline, at all the stratigraphic levels shown in the reconstructed fold shown Figure 19 and plotted the results in Figure 29 with respect to the heights above the displaced Precambrian basement. A linear regression fits the data quite adequately.

Furthermore, we can also determine the translation of the block above the décollement that terminates at the fold (theoretically). The slope of the regression line shown in Figure 29 represents the displacement according to eq. (13). The displacement, U , equals 0.42 km.

To this point, then, we would suggest that the Pitchfork anticline within the sedimentary rock sequence resembles a kink-wedge fold and that it is a result of translation of a block about 400 m on a detachment that terminates and therefore produces the compression responsible for the kink folding. We have, though, no observational evidence for a detachment surface.

Table 1. Excess areas and relative heights.

Top of the	y (km)	Area (km ²)
pCam	0.77	1.32
Cam	1.23	1.67
Dt	1.40	1.75
Mm	1.69	1.86
MIPa	1.81	1.92
IPt	1.90	1.90
Pp	1.97	2.02
Trc	2.25	2.17
Jsg	2.51	2.16
KJcm	2.74	2.26
Kt	2.90	2.30
Kmr	3.02	2.40
Kf	3.20	2.42

Indeed, although the kink-wedge mechanism appears to apply to Pitchfork anticline, it can only relate to folding observed above the Precambrian basement. The kink -wedge model explains less than half the excess area of 2.42 km² at the level of the Frontier Formation. The excess volume at the basement-cover contact is 1.32 km². This is the area unaccounted for by the kink-wedge fold model.

Let us suppose that there is a detachment within the basement and that the basement block was translated the same amount, 0.42 km, as estimated for the overburden. Then, in order to account for the excess area, using Jakobson's rule (see Johnson and Durdella, 2001), we estimate that the detachment is at a depth of $1.32 \text{ km}^2 / 0.42 \text{ km} = 3.14 \text{ km}$. This estimate is independent of the way that the excess area is accommodated beneath the fold but it does depend on the detachment mechanism.

It is possible to suggest, however, how the excess area might have been accommodated based on what is known in the accurate, deep cross section (Figure 19) of the Pitchfork anticline. The cross section shows that the basement-cover contact in the hanging-wall is rotated with respect to the footwall contact. Lateral translation of the basement and subsequent rotation along a listric reverse fault (as other workers have proposed, Stearns, 1971; Erslev, 1986; Narr and Suppe, 1990), could cause folding in the overlying sedimentary cover sequence. By matching the displacement of the detached block from the graph in Figure 29 and the geometry of the basement-cover contact of the actual fold we obtain the basement model shown in Figure 29, in which the length of the rotated backlimb is 4.5 km and the backlimb is tilted at an angle θ of 8°. The angle of the listric fault with respect to the horizontal basement, ϕ , is 77°. The 0.42 km of horizontal displacement applied to this basement fault geometry produces a rotated basement block with an excess area equal to that of the Precambrian basement of Pitchfork anticline (Table 1). The depth to the detachment is approximately 3.3 km in this case, which is close to the value of 3.14 km estimated simply by dividing the excess area by the displacement of 0.42 km of the assumed detached block.

Figure 28. Relation between excess area and height above the uplifted Precambrian basement. The slope of the regression line represents the displacement U of the detached block. The x -intercept of the regression line represents the theoretical depth to the detachment in the model.

Let us suppose that there is a detachment within the basement and that the basement block was translated the same amount, 0.42 km, as estimated for the overburden. Then, in order to account for the excess area, using Jakobson's rule again, we estimate that the detachment is at a depth of $1.32 \text{ km}^2 / 0.42 \text{ km} = 3.14 \text{ km}$. This estimate is independent of the way that the excess area is accommodated beneath the fold but it does depend on the detachment mechanism.

It is possible to suggest, however, how the excess area might have been accommodated based on what is known in the accurate, deep cross section (Figure 19) of the Pitchfork anticline. The cross section shows that the basement-cover contact in the hanging-wall is rotated with respect to the footwall contact. Assuming that the basement is displaced and subsequently rotated along a listric reverse fault as other workers have proposed (Stearns, 1971; Erslev, 1986; Narr and Suppe, 1990) forces the folding to take place in the overlying sedimentary cover sequence. By matching the displacement of the detached block from the graph in Figure 29 and the geometry of the basement-cover contact of the actual fold we obtain the basement model shown in Figure 29, in which the length of the rotated backlimb is 4.5 km at an angle θ of 8° from horizontal. The angle of the listric fault with respect to the horizontal basement, ϕ , is 77° .

Figure 29. Forced-Fold simulation of basement in Pitchfork anticline. Model parameters include a backlimb length of $L = 4.5$ km rotated clockwise an angle $\alpha = 8^\circ$ and a listric fault dipping $\theta = 77^\circ$. These parameters equal the displaced area of 1.32 km^2 of the Pitchfork basement block.

The 0.42 km of horizontal displacement applied to this basement fault geometry produces a rotated basement block with an area equal to that of the Precambrian basement of Pitchfork anticline (Table 1). The depth to the detachment is approximately 3.3 km in this case, which is close to the value of 3.14 km estimated simply by dividing the excess area by the displacement of 0.42 km of the assumed detached block.

The last step in the simulation of Pitchfork anticline will be to apply the knowledge gained from the kink-wedge and *Forced-fold* models to the actual fold. The result, Figure 28 is a theoretical reconstruction of Pitchfork anticline. The theoretical reconstruction allows us to suggest the fold form where data are unavailable, such as above the ground surface, on the sides of the fold and in the area of the Precambrian basement.

We constructed this profile by graphically appending (not superimposing) the solution for a kink-wedge model with a friction angle of 30° (Figure 29B) in the overburden and the *forced-fold* solution (Figure 29) in the deep profile (Figure 19). The theoretical reconstruction (Figure 28) obeys all the data presented in the deep profile and uses the solutions to the kink-wedge and *forced-fold* models to better approximate the fold form where data is unavailable.

Figure 31 shows the close correlation of the accurate, deep profile with the theoretical model.

Figure 30. Theoretical reconstruction of Pitchfork anticline. This construction obeys the data provided by the deep cross section (Figure 19) and is supplemented by information provided by the theoretical models.

Figure 31. Comparison of the deep cross section (Figure 19 in black) with the theoretical reconstruction (Figure 28 in gray).

One might find a detachment at a depth of 3.3 km in basement rocks beneath the Pitchfork anticline to be surprising. The Bighorn Basin area, though, seems to be characterized by shallow to deep basement detachment faulting. For example, the thrust on the southwest side of the Wind River uplift (Figure 1), about two orders of magnitude larger than the Pitchfork anticline, displaces Precambrian rocks horizontally 18 km above a detachment approximately 25 km deep. These values are based on an analysis of profiles based on surface, well log and seismic data by Stone

(1993). We used the *Forced-Fold* model to determine the depths to detachments and the Jakobson method (eq. (13)) to calculate the displacements of detached basement blocks. Similarly, the main thrust on the east side of the Bighorn uplift (Figure 1) translates the basement horizontally toward the east 15 km above a detachment approximately 17 km in depth. Furthermore, depths of detachments of 3.3 km are quite consistent with many reconstructions of structures in the Bighorn Basin made over the years by Stone (1983a, 1983b, 1985a, 1985b, 1993).

Three-Dimensional Shape

The *Gang/litric-Fault Fold* model can be used to explain the gross, three-dimensional features of the Pitchfork anticline, including the arcuate shapes of the hingelines in both plan and longitudinal cross sectional views. The modeling is necessarily qualitative, because the rheology (linear, isotropic viscous or elastic) causes deformation produced by the faults to dissipate rapidly away from the tips of the faults.

The crestline of Pitchfork anticline converges to the north with the troughline of the adjacent syncline to the east as noted in the shallow surface profiles (Figure 12). This convergence may be due to the interaction of Spring Creek anticline to the north (Figure 32). In addition, the southern end of Pitchfork anticline trends toward the southeast before curving toward the south where the North and South Sunshine anticlines become prominent.

Using *Gang/litric-Fault Fold*, (www.eas.purdue.edu/fauxpli) we model the interactions between these three folds. Figure 32 shows the topographic surface produced by a mechanical model with the following properties: 15 km long fault segment for Spring Creek, a 5 km segment for Pitchfork and an 8 km segment for both North and South Sunshine. All the faults dip 45° east and are 2 km in width. The modeled faults can produce folds that interact with each other similar to the way Pitchfork anticline and the neighboring structures interact, as shown in Figure 32.

The crestlines of Pitchfork anticline are curved in both plan and longitudinal view (Figure 16 and Figure 21). This curvature may be related to the geometry of the fault(s) responsible for producing the structure. Using the *Gang/litric-Fault Fold* model, we can show that crestline curvature in plan view may be produced by two different arrangements of fault segments. In the first case we use fault segments joined with changing strike along the length of the fold and in the second we use fault segments with the same strike and changing fault width.

Figure 32. Topographic map view of Pitchfork anticline and adjacent structures, produced by the Ganglistric-Fault Fold model. The model shows how Pitchfork anticline interacts with adjacent folds. Each of the folds shown in the model is produced from a single rectangular fault segment dipping 45° east.

Figure 33A shows a box diagram of two fault segments joined together with different strikes. This fault arrangement produces a topographic surface feature (Figure 33B) very similar to surface

features shown on the structural contour maps presented earlier. The steep forelimb in the topographic map of the model produces a relatively sharp corner where two fault segments join. This is uncharacteristic for the Pitchfork anticline, but could be eliminated by using more than two fault segments. In addition, the structure contour map produced at the surface of the model area shows that the fold axis plunges at both ends due to the termination of the fault segments.

In the second case (Figure 34), we use multiple fault segments joined together along the same strike and decreasing widths from the center segment outward. The faults dip 45° to the right in the figure. The topographic map produced by this arrangement shows a crestline that curves more gently than in Figure 33B. This gradual curvature is reminiscent of the curvature of the actual crestline.

A.

B.

Figure 33. Gang/lisric-Fault Fold used to produce the arcuate, doubly plunging fold crestline of the Pitchfork anticline. (a) A pair of angled faults is used to produce the crescent-shaped fold axis. (b) The topographic contour produced at the surface horizon by the Gang/lisric-Fault Fold model. Scale is arbitrary on both axes.

Figure 34. Gang/lisric-Fault Fold used to produce the arcuate, doubly plunging fold crestline of the Pitchfork anticline. A group of fault blades with decreasing height is used to produce the curvature. Topographic contour map produced at the surface horizon shows a gradual curvature similar to Pitchfork. Scale is arbitrary on both axes.

Figure 35. Longitudinal profile of the crestline produced by multiple, joined fault segments with parallel strike, using the Gang/lisric-Fault Fold model. The faults dip toward the viewer. The crestline shown was constructed from the structure map in Figure 32.

In the longitudinal profile, the crestline is curved, concave downward (Figure 21). This curvature can be modeled by a single fault segment, where the amplitude of the fold produced dies off along the length of the fold similar to Figure 33, or by multiple fault segments with variable width.

The latter is the case in Figure 35, where the crestline produced in Figure 34 is shown in longitudinal view. The fault segments are shown in gray and dip toward the viewer at 45°. The curvature or plunge of the crestline in this view is gradual much like that of the actual crestline in Figure 21. North and south in Figure 35 corresponds to the top and bottom respectively of the map in Figure 34.

DISCUSSION

Although we have singled out the Pitchfork anticline for special study, we do not consider this anticline to be a representative sample of thrust folds in the Bighorn Basin. Nor are we studying a "type" fold. We do not know whether this anticline is similar or different from all the other thrust folds in the Bighorn Basin. It is simply an example of a thrust fold for which we have unusually detailed information. Before one draws brash conclusions about characteristics of a "type" fold in the Bighorn Basin, one will have to study many of the folds and determine what is common and what is different among the folds. Each fold will have to be studied at the level of detail (at least) than we have found possible with the Pitchfork anticline. Eventually, common features will become obvious.

Other Evidence for Horizontal Compression

The form of Pitchfork anticline can be produced by a combination of kink-wedge folding in the sedimentary cover and a basement block displaced on a horizontal detachment. Its termination rotated on a listric reverse fault, apparently, to form the core of the Pitchfork anticline.

Layer-parallel maximum compression is essential to the formation of the kink-like geometry of the kink-wedge. One type of evidence for compression is visible in the deepest, transverse cross section shown in Figure 19 and in the four shallower serial cross sections shown in Figure 22. Layer-parallel or low-angle faulting in intercalated shales and thin limestone beds in the Sundance Formation horizontally displace about 200 m in the crest of the Pitchfork anticline. The faulting reflects horizontal shortening and vertical thickening in beds near the crest of the anticline.

A.

B.

Figure 36. Right-lateral tear fault located on the west limb of Pitchfork anticline. (a) View to the east showing the location of the fault along the west limb. White line indicates the location of the fault and the arrow shows the direction photo (a) was taken. (b) View to the south of the tear fault. Displacement is approximately 10 meters.

The geologic map shows a series of five, northwest trending, tear faults located along the length of the west limb of the Pitchfork anticline. Each of these faults accomplishes right-lateral slip and the orientations of the fault planes are near vertical. Four of the faults offset the entire Frontier Formation and upper Mowry Sandstone beds, whereas only the north-most fault offsets the lower Frontier Sandstones. Magnitude of horizontal displacements on the tear faults range from 10 to 100 meters. The general trend is for displacements to increase toward the midwidth of the structure. Figure 36A is an oblique view of the west limb of the anticline, showing the location of one of the tear faults. The apparent strike-slip offset can be seen in Figure 36B, where a lower sandstone bed of the Mowry Formation is offset 10 m.

The largest displacement occurs at the middle of the five faults (Figure 37) and offsets resistant sandstone beds of the Frontier and Mowry by 100 meters. The faults cannot be traced into the Cody shale laying to the west, but do terminate near the center of the anticline in the Mowry shale.

Figure 37. View to the north showing the largest of five right-lateral tear faults that cut across the west limb of Pitchfork anticline. Displacement on this fault is approximately 100 m. Dotted line shows matching beds across the fault.

Criterion of Jakobson's Rule

One of the pieces of evidence that the Pitchfork anticline within the sedimentary cover is a kink-wedge fold is that the excess area at each horizon of the fold follows Jakobson's rule, that the excess area is equal to the uniform displacement times the height of the horizon above the detachment surface. In more general terms, the relation between the excess area and the height of a horizon is a linear function of the height.

We demonstrated in previous pages that Jakobson's rule is an integral part of the kink-wedge fold model. The theoretical kink-wedge has this property and the Pitchfork anticline has this property. One might well ask, of course, whether this property is simply a property of any gliding block and therefore clearly nondiagnostic.

It is not, as we can show with the Fault-Tip Pucker fold model, which forms under the same conditions as the kink-wedge fold except the properties of the medium are homogeneous and isotropic. An example of the Fault-Tip Pucker fold is shown in Figure 26. The relation between the excess area and the height, y , above the detachment surface is shown in Figure 38. In contrast to the relation for the kink-wedge fold (Figure 29), the relation for the Fault-Tip Pucker fold is clearly nonlinear. In fact, the best fit polynomial to order three has the form,

$$\text{Area km}^2 = 63.434 \text{ km } y^3 - 24.113 \text{ km } y^2 + 4.8536 \text{ km } y + 0.8532 \text{ km}^2.$$

We note that there is an excess area of about 0.85 km^2 at the level of the detachment surface. Presumably this is reflected in the anticline that appears beneath the detachment surface (Figure 26).

It is clear in any case that not all folds that form at terminations of detachments display a linear relation between excess area and height above the detachment.

Figure 38. *Relation between excess area and height above the detachment for the Fault-Tip Pucker fold (Figure 26).*

Other Descriptions, Interpretations and Names Applied to the Pitchfork Anticline

Previous studies of the structure of the Pitchfork anticline were of two types. The earlier studies of the area focused on description of the form of the structure in the upper part of the stratigraphic section and included little in the way of analysis. The later studies focused on how the fold should be classified, that is, named, and on speculation about formation scenarios of the structure.

The first published descriptions of Pitchfork anticline were by Hewett and Lupton (1917), geologists with the U.S. Geological Survey. Their investigation of the anticlines of the southern part of the Bighorn Basin provided primary and unbiased observations about the near subsurface form of the Pitchfork anticline. Their study produced the first geologic map of Pitchfork anticline (Figure 39), but they did not speculate on the fold form at depth. Pierce and Andrews (1941), also in association with the U. S. Geological Survey, conducted investigations of the surface geology and mineral resources of the areas south of Cody, Wyoming, on the west flank of the Bighorn Basin. From their study, Pierce constructed a shallow, near subsurface profile (Figure 3B) across Fourbear and Pitchfork anticlines, which showed the regional dip of the west flank of the basin. Their reports on the oil and gas potential of the area spurred subsequent investigations by Coe (1967), who reported the history of petroleum production of Pitchfork oilfield. He showed the evolution of knowledge of Pitchfork anticline with structure contour maps constructed on the top of Pennsylvanian Tensleep Formation based on the available well data from the 1920's to the 1960's, but he did not publish any cross sections through the fold. His study showed that wells drilled on the flanks of Pitchfork anticline consistently migrate in the up dip direction.

Petersen's study (1983) was the first to incorporate borehole data, to classify the Pitchfork anticline and to speculate about its mode of formation. He did not have the seismic data available today. He termed the fold a "foreland detachment structure" and constructed a cross section consistent with the prevailing notions (e.g., Rodgers, 1963) about the forms and causes of such structures (Figure 40). He envisioned a fold caused, in some fashion, by a series of listric thrusts emanating from a basal detachment just above the Precambrian basement. To Petersen, then, the fold does not involve the basement rocks. A series of listric reverse faults emanating from a detachment above the basement shown in Petersen's interpretive cross section is, unfortunately, unable to produce the fold shown in our cross sections. There is simply inadequate duplication of

stratigraphic section to account for the excess volume of the fold, at any level. Wiltchko and Chapple (1972) drew the same conclusion about explanations of the folds on the Appalachian plateau in western Pennsylvania in the 1960's (championed by Rodgers, 1963, Gwinn, 1964, 1970 and others). Wiltchko and Chapple (1977) explained why the splay fault mechanism could not produce the Plateau folds and their explanation applies as well to the Pitchfork anticline. They proposed layer-parallel shortening.

Figure 39. Geologic map of Pitchfork anticline constructed by Hewett and Lupton (1917).

Figure 40. Structural interpretation of Pitchfork anticline by Petersen (1983), who envisioned the fold as a foreland detachment structure with listric faults emanating above the basement.

A study of the Pitchfork anticline by Stone (1993) used a combination of surface, well and seismic data to construct a cross section (Figure 41) of the fold⁹. He classified the Pitchfork anticline as a "thrust-fold" (Stone, 1983), apparently implying a causative relation between a reverse fault and a fold. In Stone's thrust-fold interpretive cross section (Figure 41), a long reverse fault is shown cutting up through much of the folded section. The largest problem with Stone's interpretive cross section, and his proposed mechanism of formation of the Pitchfork anticline, in our view, is that there is no evidence in surface data, borehole data or seismic profile to support the interpretation of a fault cutting up through much of the section as Stone interprets (compare Figure 19 and Figure 41).

Figure 41. Structural cross section of Pitchfork anticline by Stone (1993), based on surface, well and seismic data.

The cross sections presented by Petersen and Stone accurately show the geometry of the Pitchfork anticline within the sedimentary section and agree with the data that we have carefully sorted and presented in Figure 19. The differences among the interpretive cross sections presented by

⁹ We have found that, of all the modern investigators of structures in the Bighorn Basin, Stone stands out as an excellent source of nearly pure description of fold form.

Petersen, Stone and us are in the different mechanisms judged to have been responsible for producing the fold forms. According to Petersen and Stone, reverse faults are responsible for the fold.

According to our study, the termination of a detachment surface is essentially responsible for the layer-parallel shortening in the sedimentary cover. The detachment surface could be at the basement/sediment cover or could be in the basement rocks. In addition, an upwarp or upthrust in the basement rocks provided the volume of rock that fills the core of the fold at the level of the basement. Curiously, a relatively recently published analysis of the Pitchfork anticline has the least feasible representation of Pitchfork anticline. Chester and Chester (1990) christened the Pitchfork anticline a "fault-propagation fold," in which a fold that follows certain geometric rules is supposed to be related to the tip of a fault. The "fault-propagation fold," like the "foreland detachment structure" of Petersen, has no theoretical basis. Chester and Chester made selective use of Petersen's borehole data as well as available seismic data and extensive use a straight-line-length balancing technique in order to draw a cross section of the anticline (Figure 42). Although their cross section matches some of the surface and subsurface data of the Pitchfork anticline and the straight-line fold limbs are reminiscent of some of the rather straight fold limbs in the carefully reconstructed fold form (Figure 19), there are some obvious transgressions of fact. There is no hint in the field, in carefully constructed cross sections or in seismic data for flat layering near the crest of the Pitchfork anticline. That feature of the cross section is a necessary requirement for the model to maintain the "fault-propagation fold" geometry. Another problem with their cross section is that it superficially reminds one of kink folding, although the geometry is inconsistent with that of kink folding. The limbs of the reconstructed folds are poor renditions of kink bands because the inclinations of the limbs make no sense physically. We have shown with the kink-wedge model that a forelimb dip of 40° corresponds to a friction angle of 50° (Figure 29C), which is abnormally high for a friction angle. Such large friction angles are essentially unheard of. Thus their cross section has few redeeming qualities and is a poor rendition of the Pitchfork anticline. The fault-propagation model presented by Chester and Chester is an unsuitable model because it fails to conform to the surface, well log and seismic data that they reportedly use in the construction.

Figure 42. Structural interpretation of Pitchfork anticline by Chester and Chester (1990). They interpreted the structure as a fault-propagation fold with basement fault. Their depiction uses line-length balancing to produce the gross fold form.

ACKNOWLEDGEMENTS

We thank the Department of Energy, grant DE-FG02-98ER14886, supervised by Nick Woodward, for financial support of this research. We are grateful to Donald Stone, consulting petroleum geologist in Denver, CO, for providing general guidance and for helping us find suitable field areas in the Bighorn Basin. Don also helped us greatly in our search for subsurface data. Even Don, though, was unable to shake loose a set of 3-D seismic data for this research! Too bad.

THIRD OBJECTIVE: GROWTH OF FAULT-RELATED STRUCTURES ASSOCIATED WITH EARTHQUAKES.

In the course of our research into fault-related folding we have used some of the theory to analyze ground deformation and rupturing associated with the 1971 San Fernando earthquake and the 1999 Düzce, Turkey earthquake. Following are manuscripts that summarize our research to date.

STRAIN AND RUPTURE BELTS ALONG SYLMAR FAULT SEGMENT, 1971 SAN FERNANDO EARTHQUAKE

by
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Written Fall 1999
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ABSTRACT

Measurements of strains, using horizontal street surveys by the City of Los Angeles before and after the 1971 San Fernando earthquake, provide new information about the visible and blind ruptures of the Sylmar segment of the San Fernando fault zone. The Sylmar segment is characterized by the largest horizontal displacements and the largest domal uplift in the hanging wall of the San Fernando fault zone. We combine the strain data with published data on horizontal

¹⁰ The maps and figures presented here are from Xan's M.S. thesis. The text is by AMJ. Davidson, X.M.J., 2000. Extent of faulting produced during the 1971 San Fernando, California earthquake. M.S. thesis, Department of Earth and Atmospheric Sciences, Purdue University, 81 p.

and vertical displacements and the data from City leveling surveys, to identify blind extensions of the Sylmar segment. A distinctive bulge in the hanging wall north of the rupture of the Sylmar segment reflects reverse faulting and probably a blind segment extending 3.0 km east of the Sylmar rupture

Parallel strain belts are associated with surface faulting. Analysis of the strains along the Sylmar rupture indicates that there are parallel belts of elevated strains, one centered on the Sylmar rupture and the other to the north. The faulting was left-lateral/reverse. Reflecting the left-lateral faulting, the belt centered on the fault trace shows left-lateral shearing. The belt to the north shows right-lateral (rebound) shearing. Reflecting the reverse faulting, the belt to the north contains tension cracks and other extensional phenomena whereas the belt to the south contains thrust faults and other compressional phenomena.

Parallel strain belts are also associated with blind faulting. The same strain belts that occur along the Sylmar segment extend about 1 km to the west of the Sylmar segment, indicating that there is a blind segment at least 1.0 km long at the west end of the Sylmar rupture.

We can model the vertical and horizontal displacements and the strains at the ground surface near the Sylmar segment and its blind extensions. A theoretical fault with a surface rupture 6 km long, dipping 45° north produces the maximum vertical and horizontal displacements as well as the essential features of the patterns of displacement. The blind eastern part appears to dip more gently, perhaps 30° north. The net slip on the single dislocation surface used to model the faulting is 4.1 m, with 3.1 m of left-lateral strike-slip and 2.7 m of reverse dip-slip. These values are in general agreement with results of fault modeling by other investigators. As a check, we used the same theoretical model to calculate strains in the ground surface. The field measurements and theoretical results are essentially the same.

Contrary to early reports, there was widespread coactive faulting during the 1971 San Fernando earthquake. The correlations of multiple belts of elevated strains and fracturing with both visible and blind fault segments at San Fernando suggest that similar belts mapped at the Winnetka-Northridge and Granada Hills deformation zones following the 1994 Northridge earthquake are reflections of coactive, reverse, blind faulting.

INTRODUCTION

In a series of papers on earthquake rupturing and ground deformation we have shown that there are distinctive fracture and strain patterns produced at the ground surface by tectonic deformation during fault slip. The 1992 Landers, California, earthquake produced spectacular *belts of ground ruptures* 50 m to 500 m wide and they occurred along a fault zone (Johnson et al., 1994, 1997; Fleming et al., 1998). The 1989 Loma Prieta and 1994 Northridge earthquakes—with deep hypocenters and main faults that did not intersect the ground surface—show clear examples of *coactive ruptures*, or slippage and fracturing along faults related in some way to the fault producing the main shock. At Loma Prieta, coactive faulting occurred in the Summit Ridge and Skyland Ridge shear zones (Martosudarmo et al., 1997) as well as along a series of thrust faults in the Los Gatos area that extend several kilometers northwest to the Stanford University campus (Plafker and Galloway, 1989; Haegerud and Ellen, 1990). Distinctive deformation belts at Winnetka and Granada Hills have side-by-side belts of compression and extension and appear to have been produced by coactive, blind reverse faults during the Northridge earthquake. We could not assert that the deformation zones are certainly a result of blind faulting because we do not see fault traces at Northridge and the measurements of horizontal and vertical displacements are widely spaced. As a result, we have only one line of evidence of blind faulting at Northridge. *The evidence is generally considered to be unconvincing to others because most earthquake researchers have ignored or do not recognize evidence of deformation belts along visible earthquake ruptures and, especially, along blind earthquake ruptures.*

Figure 1. Locations epicenter and surface rupture of the 1971 San Fernando earthquake.

In order to further explore the association of deformation belts and earthquake ruptures we have decided to re-examine ground deformation produced during the 1971 San Fernando earthquake ($M_L = 6.4$). The Sylmar damage was well-documented and many measurements of displacements (Burford et al., 1971; Yerkes et al., 1974; Savage et al., 1975) and aftershocks (Allen et al., 1971; Hanks et al., 1971; Whitcomb, 1971; Whitcomb et al., 1973; Mori et al., 1995) have been made. The rupture zones and displacements along the San Fernando fault zone were moderately well documented by Sharp (1975). The Sylmar segment was mapped in greater detail by Kamb et al. (1971).

Figure 2. Locations and types of aftershocks of the 1971 earthquake. Fault-plane solutions (due to Witcomb, 1971) show left-lateral, strike-slip on the western area of the rupture (from Davidson, 2000, fig. 2.3)

Our study adds one new data set and contributes to another. The data are derived from horizontal surveys and leveling conducted by the City of Los Angeles. Much of the area of the City was resurveyed and releveled before and after the 1971 earthquake sequence. We use the City's leveling data to determine vertical displacements and supplemented it with data by Weber (1975) and Savage et al. (1975). The City's horizontal surveys provide a means for determining surface strains at the scale of city blocks (e.g., Johnson et al., 1996; Cruikshank et al., 1996). For the San Fernando area, we analyzed 563 intersections, which produced 1100 independent strain measurements. Where data are available, strains can be determined for four quadrants at each intersection. A data set includes changes of lengths of streets and a change in the angle between streets.

Figure 3. Distribution of aftershocks (Data from Mori et al., 1994; figure from Davidson, 2000, fig. 2.5). A. Cross section of aftershocks in western part of area, the area of largely left-lateral strike slip.

Our study uses all these data sets, plus a 3-D dislocation model, to investigate the probable positions and lengths of blind extensions of the Sylmar rupture segment of the San Fernando fault zone. We use the measurements of ground deformation and displacements that occurred during the earthquake to test ideas about indicators of blind faulting developed in previous studies (e.g., Cruikshank et al., 1996). The strain measurements would be more credible as indicators of blind faulting if they were observed over areas identified as blind faults based on vertical and horizontal displacements. There are areas of surface ruptures and areas that appear to be blind faults near the

Sylmar segment. Therefore, one should be able to compare the deformation patterns adjacent to surface ruptures with deformation patterns in areas of blind faults as a way to identify them.

SURFACE RUPTURE ALONG SAN FERNANDO FAULT

The San Fernando Fault Zone

The San Fernando fault zone ruptured during the 1971 earthquake over a total length of at least 12 km (Kamb et al., 1971). A fault zone about 50 m wide shifted. The fault zone consists of five continuous segments (Figure 4) that generally trend E-W. There were few scarps and other linear features to clearly define the fault trace.

A. (Modified from Oakshot, 1958; from Davidson, 2000, fig. 1.3).

B. Traces of San Fernando rupture and Santa Susuna fault zone. The San Fernando rupture broke through to the ground surface during the 1971 earthquake. So did the co-active part of the Santa Susuna fault west of Interstate 5 (west side of map). (from Davidson, 2000, fig. 1.2).

Figure 4 . Faults in the San Fernando Valley.

The continuous segments surrounding the Sylmar segment are the Reservoir, Mission Wells, and Tujunga segments. Starting in the west (Figure 4), the Reservoir segment is near the Van Norman reservoirs and is between the San Diego and Golden State freeways. Slip consisted of left-lateral slip of 0.16 m and reverse dip-slip of 0.38 m (The ratio, R , of strike-slip to reverse dip-slip was about $R=0.42$). Slip on this segment caused failure of the Golden State Freeway (Yerkes et al., 1974; Weber, 1975).

The Mission Wells is a short segment to the east of the Reservoir segment. All the segments trend E-W except for the Mission Wells segment, which trends NE-SW. This segment was mostly left-lateral with a strike-slip of 0.46 m and a reverse dip-slip of 0.16 m ($R=2.9$) (Yerkes et al., 1974).

East of the Sylmar segment is the Tujunga segment. The Tujunga segment is about 5 km long and has a clean break with a well-defined scarp (Figure 4). The fault dips 20 to 25° toward the north and is a reverse fault. The fault generally accommodated left lateral and reverse shift (Kamb et al., 1971). The largest slip on the Tujunga was about 2.1 m and was at the Sylmar end of the segment (Sharp, 1975). Its slip was largely left-lateral ($R=2.4$). The slip on this segment varies in amount and direction more than on any other segment (Sharp, 1975).

The easternmost segment, the Lakeview segment, is about 4 km long. This segment accommodated nearly equal amounts of thrusting and left-lateral slip on the order of 0.8 m (Barrows, 1975).

The Sylmar Segment

The Sylmar segment, to the east of the Mission Wells segment, extends for about 2 kilometers and is straighter and more continuous than the other segments. A peculiar feature of the Sylmar segment is that it is stepped about 1.2 km north of the general east-west trace of the San Fernando fault zone (Figure 4). According to Sharp (1975), the Sylmar segment slipped up to 2.5 meters, with a left-lateral to reverse-slip ratio of $R=1.3$. Kamb et al. (1971), determined relative

displacements across the Sylmar fault zone using offsets of streets and other markers and assumed that the ground in the hanging wall did not move; only the footwall shifted. They obtained measurements of 0.8 to 1.9 m of net slip along trends ranging from N13°W (1.3 m) to N7°E (1.8 m) to N67°E (1.9 m). The plunge angles ranged from 9° to 35°. Even though the Sylmar segment trends E-W, the displacements vary. Faulting ranged from pure reverse dip-slip (N13°W to N7°E) to nearly pure left-lateral strike slip (N67°E).

Rupturing Within the Sylmar Fault Zone Segment

The ground rupturing along the Sylmar segment of the San Fernando fault zone was complex. It appears as though there were two types of shear zones, one reflecting the left-lateral faulting and the other reflecting the reverse faulting. The descriptions of ground rupturing seem to indicate that the left-lateral faulting produced a belt of left-lateral shearing roughly centered on the trace of the Sylmar segment. An adjacent belt of right-lateral shearing is to the north (Bonilla et al., 1971 and Kamb et al., 1971). The reverse faulting apparently produced a belt of thrust faults and related compressional structures near the Sylmar segment. To the north of this belt, there is an adjacent belt of extension. Thus the left-lateral shearing and the thrusting occur together in a belt near the trace of the Sylmar segment and the right-lateral shearing and the extension fracturing occur together in a belt to the north. Bonilla et al. (1971) divided the rupture zone into a narrow, southern belt of predominantly shearing and thrusting and a wide, northern belt of predominantly extension. The southern boundary of the shear zone is sharp. Obvious fractures occur within a few tens of meters south of the main rupture.

The northern extension belt is defined by many closely spaced fractures in the eastern part of the Sylmar rupture zone. The fractures have relatively straight traces in open ground and their spacing is generally 20 m; the maximum is about 100 m. The belt is about 1.3 km N-S and 1.8 km E-W. The traces of the fractures in this belt are roughly parallel to the main rupture. Almost invariably, the fractures within this belt show N-S opening as well as right-lateral shift. Where there is differential vertical shift, the vertical shift is nearly always downward on the north side of a fracture.

Presumably, tension cracks formed parallel to the surface trace of the Sylmar fault zone because of uplift and extension in the hanging-wall. Figure 5 shows the position of the fractures on the upwarp. The tension cracks accommodated normal faulting and right-lateral shearing as rebound phenomena. The same sequence of tension cracking and lateral slip within a fault rupture zones was documented at Loma Prieta (Johnson and Fleming, 1993; Martosudarmo et al., 1997) and at Landers (Johnson et al., 1996).

HORIZONTAL AND VERTICAL DISPLACEMENTS

Figure 2 shows horizontal displacements of 12 points near the Sylmar fault segment (data from Yerkes et al., 1974). The displacements are accurate to within ± 6 cm. The largest displacements are in the hanging wall block, just north of the Sylmar surface rupture, and point due west. Within this block, the displacements range from WNW to west. The horizontal displacements decrease in magnitude to the north and west. Displacements south of the Sylmar rupture, within the footwall block, point due north and are nearly constant at 0.5 m.

A peculiar feature of the horizontal displacements is a discontinuity between the components of the northern and southern blocks. In the northern block all the displacements point WNW; whereas, all the displacements in the southern block point NNE (Figure 2). The location of the displacement discontinuity coincides with the surface ruptures of the Sylmar, Mission Wells, and Reservoir fault segments. The same discontinuity occurs along each of the gaps between the surface ruptures and extends as far westward as Granada Hills (dashed lines in Figure 4). It has been previously

suggested that the discontinuity marks the location of a blind segment of the San Fernando fault zone west of the Sylmar fault segment (Yerkes, et al., 1974; Davidson, 2000).

Figure 5. Contours of vertical displacement throughout the San Fernando Valley. Contour interval of uplift is 0.1 m. Topographic contour interval 40 feet. The areas of uplift are mostly along the Sylmar and the Tujunga segments. There is a local uplift along the Reservoir segment. Data were collected from several sources. Most of the data are new (unpublished heretofore) provided by the Survey Division, City of Los Angeles. These data were supplemented by data from Savage et al. (1975), Weber et al. (1975) and Yerkes et al. (1974). Data for the Tujunga segment are from Burford et al. (1971). (from Davidson, 2000, fig. 1.4).

Vertical displacements near the Sylmar fault zone are also shown via contours in Figure 5. The vertical displacements shown are only part of a roughly elliptical area of uplift described by Morrison (1973). The ellipse with a major axis of 20 km in the NE-SW direction and 11 km in the NE-SW direction is centered approximately at the Olive View Hospital (Figure 4).

The map of vertical uplift was constructed using three data sets. The City of Los Angeles leveled nearly the entire city after the earthquake, and made all the data available for our analysis. These data were supplemented by data from Savage et al. (1975) in the western and central part of the Sylmar segment. For the area east of the Sylmar segment, we did not have survey data. Therefore, all the data are from Burford et al. (1971).

The contour map in Figure 5 shows one area of marked vertical uplift. It extends along the northern side of the Sylmar segment and along part of the Tujunga segment of the San Fernando fault zone. The contours show a bulge north of these fault traces. Along the Sylmar segment the bulge reaches about 1.6 m in height and is markedly asymmetric. There is a sharp uplift increase on the south flank of the bulge from 500 m south of the fault trace to 500 m north of the rupture, where the bulge crests. The bulge dies out much more slowly toward the north, on the northside of the crest. It drops from about 1.6 m at the crest of the bulge to about 0.5 m at the base of the San Gabriel Mountains (at Olive View and Veterans hospitals). Thus, the bulge accumulates about 1.4 m of uplift over a horizontal distance of about 750 m on the south flank and dissipates 1.1 m of uplift over a horizontal distance of about 2200 m on the north flank.

A. Horizontal components of displacements in western half of area. Displacements north of the fault are generally westward. Displacements south of the fault are generally northward. (From Davidson, 2000, fig. 1.5).

B. Indeed, we have extrapolated a long blind segment of the San Fernando fault rupture to the west by separating the westward-trending displacement from the northward-trending

displacements by a dotted line. (Displacements from Yerkes et al., 1971; figure from Davidson, 2000, fig. 2.1).

Figure 6. Horizontal components of displacement.

The bulge along the Tujunga segment is quite different (Figure 5). It is higher, reaching a peak of perhaps 2.2 m and it is nearly symmetrical. The crest is much farther, about 1500 m, north of the ruptured fault segment. The bulge is somewhat wider, perhaps 3000 m as compared to 2000 m for the Sylmar segment.

A level line resurveyed by Burford et al. (1971) detected a change of vertical displacement in a gap in the surface rupturing between the west end of the Sylmar segment and the northeast end of the Mission Wells segment. The releveled shows that a point on the northwest side of the gap moved upward about 0.43 m relative to a point on the southeast side of the gap. As they indicate, (op. cit., p. 81), "[this] simply testifies to the existence of faulting that for reasons unexplained has not yet propagated to the ground surface." This is what we mean by a blind fault.

Another line was releveled across the Sylmar fault to the base of the San Gabriel Mountains. The leveling circuit shows that the footwall and hanging wall blocks both tilted slightly northward. The ground between the western tip of the Sylmar segment and the base of the San Gabriel Mountains tilted northward about 0.02° (Savage et al., 1975).

MODELING OF HORIZONTAL AND VERTICAL DISPLACEMENTS

The Entire Fault Zone

Several investigators have modeled the horizontal or vertical displacements that occurred during the San Fernando earthquake using a solution for one (Savage et al., 1975), several (Trifunac, 1974; Heaton, 1982) or many (Canitez and Toksoz, 1972; Alewine and Jordon, 1973) rectangular dislocations. Most of these analyses are based on analytic expressions for an inclined fault given by Mansinha and Smylie (1971). The displacements have also been modeled with finite elements (Jungels and Frazier, 1973; McCowan et al., 1977). All of these investigations have worked with the entire rupture length of the San Fernando fault zone. Most of them, except Alewine and Jordon (1973) and Trifunac (1974), largely ignored the complicated displacement patterns in the vicinity of the Sylmar segment that are of special interest here.

Savage et al. (1975) modeled the horizontal and vertical displacements of geodetic stations in the area of the San Fernando earthquake with the solution for a single, rectangular dislocation. They assumed that the slip is 2 m of reverse-slip and 2 m of left-lateral slip. They determined that the gross pattern of measured displacements is reproduced by a simple dislocation model with a length, along strike, of 15 km, a fault width (down dip) of 8 km, a depth to top of fault at 0.75 km, a strike of $N10^\circ E$ and a dip of $45^\circ N$. They point out that the down-dip width of their model fault is too small and suggest that the displacement decreases down dip rather than remain constant as in their simple model.

A multi-element dislocation model by Alewine and Jordon (1973) represented the fault ruptures with 47 appropriately oriented rectangular fault elements to derive a best-fit model of the dip-slip component of the San Fernando fault. The model they adopted consists of a main fault segment slightly less than 45° near the ground surface and steepening to slightly more than 45° at depth. Near the ground surface, the main fault becomes a fan of five bifurcated faults with shallow to steep dips. According to their analysis, the dip-slip varies widely with depth along the fault, varying from about 2.5 m near the ground surface, to 4 m at a depth of 1.2 km, to nearly zero at 4.5 km, to about 4 m at a depth of 10 km, and to 2 m at the epicentral depth of about 11 km. The three-

dimensional finite-element modeling of the vertical displacement data by Jungels and Frazier (1973) indicates a similar pattern: larger displacements near the ground surface but then displacements linearly decrease to zero at the hypocenter.

Trifunac (1974) used a dislocation model by Haskell (1969) to model ground deformation in the San Fernando area. He tried a variety of faults represented by several rectangular elements and concluded that the best-fit model consists of eleven elements up to about 12 km long (along strike) and 18 km width (along dip). The near-surface element was about 6 km long, representing the Tujunga and Lakeview segments. A wider element below represents the Sylmar segment. The net slips vary from element to element with depth. They range from 5.2 m near the ground surface to almost zero in the third element to 11.9 at the element with the hypocenter (Trifunac, 1974).

Modeling of the Sylmar segment and its Extensions

Most of the investigators cited above used a solution by Mansinha and Smylie (1971) for the displacements around a rectangular fault-dislocation. The solution includes a dipping fault plane subjected to strike- and dip-slip.

It is a solution for a fault beneath a free surface, where normal and shear tractions are zero. The problem is solved by superimposing the solution of an infinite body containing a rectangular dislocation without a free surface and the solution of an image of the dislocation reflected across the plane of the free surface. The image of the dislocation has the opposite sense of slip as the original dislocation, so the shear traction is zero but the normal traction is not. Finally, the normal tractions are removed in order to obtain the complete solution. We have implemented the solution with Microsoft Visual Basic. The program is available at our web site, www.eas.purdue.edu/fauxpli/gang_fault_fold.htm (note underscores beneath spaces).

In our implementation of the Mansinha and Smylie (1971) solution, we can include a single rectangular dislocation or many rectangular dislocations to represent arrays of faults as well as curved faults. However, in our analysis of the Sylmar segment we found that a single rectangular dislocation largely modeled the deformation. We use the solution to generate fields of horizontal and vertical displacement and strains. The solution is for displacements at points of a grid, generally at the level of the ground surface. The solutions for displacements of grid points are then used to compute strains at the center of each rectangle defined by four corner grid points.

The purpose of the theoretical analysis is to explain, as simply as possible, the essential features of the vertical and horizontal displacements of points at the ground surface near the Sylmar segment of the San Fernando fault zone (Figure 5).

Figure 7. Illustration of effect of dip of fault and reverse slip affect patterns of vertical and horizontal displacements at the ground surface near a blind fault that terminates near the ground surface. Contours shows amount of uplift for 1 m of slip. Faults dip toward top of figure.

In modeling the discontinuity between the hanging-wall and footwall, we have run a series of examples of end-member dip-slip and strike-slip faults. Figure 7 shows horizontal and vertical displacements as dip is varied. Reverse, dip-slip faults. The rectangle in each figure is the projection of the fault plane onto the ground surface. In all examples, the fault dips toward the top of the figure and the fault tip is slightly below the ground surface. The lengths of the arrows and the contour interval are scaled the same, so the fault dip is the only parameter that is changing. Other parameters that can be changed are the length and width of the fault plane and the depth of the tip of the fault. Increasing the length (along strike) of the fault plane increases the displacements at the ground surface. An increase in the width (along dip) of the fault plane also increases the displacements at the ground surface and produces larger displacements at a greater distance from the fault. An increase in depth of the upper tip of the fault beneath the ground surface decreases the displacements at the ground surface.

Figure 8. Effects of dip of fault and strike slip affect patterns of vertical and horizontal displacements at ground surface near a blind fault. Fault dips toward top in figures. Contours show amount of uplift for 1 m of slip. Slip is left lateral, to contours show depression at right and bulge at left end of fault. (from Davidson, 2000, fig. 3.2).

- ? The magnitudes of horizontal displacements parallel to the strike of the fault are very large in the hanging wall and very small in the footwall for the strike-slip faults (Figure 8).

For strike-slip movement (Figure 8), if the fault dips more or less steeply the horizontal displacement pattern changes. For a gentler dip (30°), the displacements in the footwall are smaller and the displacements in the hanging wall are larger. For a steeper dip (60°), the displacements in the hanging wall and footwall are more nearly equal. For a vertical fault, the horizontal displacements are the same in the hanging wall and footwall. Thus, the gentler the dip of the strike-slip fault, the larger the difference between displacements in the hanging wall and footwall. The horizontal displacements in the field show a decrease in the hanging wall with the largest displacement near the upper fault tip. In the footwall the displacement values are very small (Figure 5). Therefore, we can conclude that an idealized strike-slip fault must dip at a low angle to explain the observed displacements.

- ? The horizontal displacements normal to the strike of the fault are small but are larger in the footwall than in the hanging wall for the reverse faults (Figure 7).

For pure dip-slip (reverse) on a fault dipping 45° we see in Figure 7 that the horizontal displacements are larger in the footwall than the hanging wall. The horizontal displacements increase in the footwall and decrease in the hanging wall with increasing dip of the fault.

- ? The contours of vertical uplift for the Sylmar fault show significant uplift (up to 2.2 m) in the hanging wall block but essentially no displacement in the footwall block.

To the west of the Sylmar segment the uplift is 0.5 to 0.7 m and to the south the uplift is zero to slightly negative. We suggest that the general uplift of the hanging wall of about 0.5 m is a result of the San Fernando fault zone as a whole. Thus, the maximum uplift of the Sylmar segment is about 1.7 m.

A reverse fault dipping at 45° could explain the observed pattern of large uplift in the hanging wall and small down dropping of the footwall. We would note that the down dropping of the footwall is even smaller for a fault dipping 30° . In addition, the down dropping of the footwall becomes larger as the fault steepens, as shown for a fault dipping 60° (Figure 3F).

- ? The contours show an asymmetric uplift pattern adjacent to the surface trace of the Sylmar rupture (solid line in Figure 5).

The pattern consists of a steep increase in uplift near the rupture, a crest of about 2 m at a distance of 0.5 to 0.75 km north of the rupture and a gradual decrease in uplift to 0.6 m about 2.5 km north of the rupture.

For a reverse fault dipping 45° (Figure 7), the gradient in vertical displacements within the hanging wall is very steep near the tip of the fault, reaches a ridge near midwidth in the projection of the fault, and becomes shallower farther away. Thus, the asymmetry of the vertical uplift pattern north of the Sylmar fault segment (Figure 5) might be explained in terms of a fault dipping about 45° north. The effect of dip on asymmetry is shown in Figure 7. The shallower the dip, the more asymmetrical the uplift; the steeper the dip, the less asymmetrical the uplift.

We can thus use these results to suggest an explanation for the asymmetry of the uplift north of the Sylmar segment. If we assume that the uplift pattern is a result of a single fault, we would suggest that it is due solely to the reverse component of slip and that the fault dips 45° or a little shallower.

- ? The contours shown east of the Sylmar segment indicate that the uplift becomes symmetric towards the east.

Figure 9. A. Horizontal displacements computed for an idealization of the Sylmar fault segment. The model fault dips 30° and pure left-lateral, strike-slip. B. The displacement of the Sylmar fault segment. The displacements in the hanging wall rather well match those in the hanging wall of the Sylmar segment. (from Davidson, 2000, fig. 3.4).

Figure 10. Horizontal displacements in the footwall. A. Observed horizontal displacements in the footwall along the Sylmar segment. B. Model solution at 60° dip and pure reverse slip. The solution matches the observations moderately well.

Figure 11. Vertical uplift in the vicinity of the Sylmar segment. A. Uplift measured next to the Sylmar segment. Maximum uplift is 2 m at the eastern end. B. Model solution for the uplift, assuming a single rectangular fault. Dip is 45° . Pure reverse slip. Maximum uplift is 0.5 m for a slip of 1.0 m. (from Davidson, 2000, fig. 3.6).

Figure 12. Uplift north of the outcrop of the Tujunga segment. A. The maximum uplift is 1.2 km north of the surface rupture. B. Uplift produced by two fault-segment model, an upper fault dipping 15° and a deeper fault dipping 45° . Contour interval is 0.05 m. Maximum uplift is 0.65 if there is 1 m of slip (from Davidson, 2000, fig. 3.7).

Figure 12. Favorite solution for the dip of faults in the Sylmar and Tujunga fault segments, San Fernando fault. A. Suggested, near-surface dip of 45° for Sylmar segment. B. Dip of 15° near-surface for Tujunga segment. C. At depth, the 15° Tujunga segment steepens to 45° and is continuous with the Sylmar segment.

One would favor a dip between 30 and 45° to explain the obstrike displacements.

The symmetrical pattern extending to the east of the Sylmar rupture, behind the Tujunga segment, is inconsistent with a steeply dipping fault. Figure 3D shows a shallow fault dip with symmetrical uplift. This uplift cannot be from the Tujunga segment because it is stepped 1.5 km north of the surface rupture. On this basis, we suggest a blind fault extends about 3 km east from the eastern end of the Sylmar segment (dashed line, Figure 5). We term this the Indian Canyon blind segment. According to our analysis shown in Figure 7, we would suggest that the fault responsible for the uplifted area has a steeper dip along the Sylmar segment than along the Indian Canyon segment.

The theoretical analysis produces many of the features we have observed near the Sylmar segment of the San Fernando fault zone. We can conclude that the large strike displacements in the hanging wall and zero in the footwall could be a result of a fault dipping northward

The large horizontal displacements in the footwall, normal to the strike of the fault, can be explained in terms of a fault dipping 45° . The same fault dip produces an asymmetrical pattern of vertical uplift that resembles the contours along the Sylmar segment (Figure 5).

The pattern of uplift east of the Sylmar segment suggests that there is a blind fault adjacent to the trend of the Sylmar segment. It dips less steeply, approximately 30° , and has been named the Indian Canyon segment.

After exploring a wide range of models, including two fault planes with increasing or decreasing dips downward, we conclude that we can explain the patterns of displacement near the Sylmar segment in terms of a single rectangular fault plane dipping about 45° north.

The angle of 45° is a common one estimated for the San Fernando fault zone as a whole. Savage et al. (1975) estimated a dip of 45° for a single fault plane. Langston (1976) decided that the fault consists of two rectangles, the near-surface fault dipping 29° and extending about half the distance to the hypocenter and the near-hypocenter fault dips 57° the remainder of the distance. Heaton (1982) favors a model that includes one fault that intersects the ground surface at a dip of 45° and another fault, with a dip of 54° and stepped north of the first fault, that originates near the hypocenter and approaches to within 3 to 4 km of the ground surface. Tifunac's (1974) data suggest a fault consisting of several dislocation elements dipping at 40° northward. He indicated that he was unable to distinguish a curved from a planar fault. Thus our assumption of a 45° dip for the fault plane of the Sylmar segment alone, based on our own independent analysis, corresponds with the angle generally adopted by others who modeled the faulting.

We assume that the length (along strike) of the fault was 6 km, which includes the 2 km Sylmar rupture, and the 3.0 km Indian Canyon and 1.0 km County Line blind segments (Figure 5). The fault width (along dip) was assumed to be 3 km. The near-surface end of the fault was at a depth of 200 m beneath the ground surface.

Computed Displacements for Sylmar Segment and Adjacent Blind Segments

There are several reasons that we believe the displacements associated with the Sylmar segment, including its blind extensions, can be separated from those due to other parts of the San Fernando fault zone. First, the vertical displacements produced by the Sylmar segment are much larger than those elsewhere. We have no data on vertical displacements near the Tujunga segment, except near the eastern end of the Sylmar segment. The crest of a ridge of vertical displacements there does not relate well to the trace of the Tujunga segment. To the west of the Sylmar segment, we have information on vertical displacements (e.g., Savage et al., 1975, fig. 2) but they do not show a domal uplift. They are relatively uniform at about 0.4 to 0.6 m of uplift near the Mission Wells and Reservoir segments as well as in ground to the north (Davidson, 2000).

The Mission Wells segment shows no visible change in altitude. The leveling stations are too far apart to show a bulge (Davidson, 2000, pl. 2). However, the displacements across the surface rupture suggest that this segment accommodated about 0.2 m of reverse slip.

The Reservoir segment is farther to the west. It is represented on a map of vertical displacements by a dome about 0.3 m high on an area generally uplifted about 0.5 m, to the north of the segment (Davidson, 2000; Davidson, 2000). Measurements of displacements across surface ruptures suggest this segment accommodated about 0.4 m of reverse slip.

What is lacking at the Mission Wells and Reservoir segments, as well as the blind segments joining these and extending to Granada Hills (Davidson, 2000), is the appearance of a domal uplift pattern

anywhere except along the north side of the Sylmar segment and its blind extensions. This lack, we suggest, indicates that the Sylmar segment and its blind extensions to the east and west (total length of about 6 km) are responsible for the domal uplift.

In the following example, we assume a fault length of 6 km, a width of 3 km, a dip of 45° , and a depth to the fault tip of 200 m.

In order to select strike-slip and dip-slip for the fault we use the known maximum displacements of strike-parallel displacement of 1.6 m and uplift of 2.2. The uplift, though, is partly due to the whole San Fernando fault rupture. Therefore, we subtract the nearly uniform uplift of 0.5 m that occurs to the west of the Sylmar segment. Thus, we assume nearly equal maximum values of horizontal, strike-parallel displacement (1.6 m) and vertical displacement (1.7 m) in the hanging wall of the Sylmar segment.

The strike-parallel displacement, 1.6 m, is an unknown factor, h_f , times the net slip and the vertical displacement, 1.7 m, is a factor v_f , times the net slip. We assume a ratio of strike-slip to total slip on the modeled fault plane and calculate values of the ratio of horizontal factor to vertical factor, h_f/v_f . We adjust the ratio of strike-slip to total slip until h_f/v_f is equal to the known ratio of maximum strike-parallel displacement to maximum vertical displacement. Then we have a possible solution. We can compute the net slip from either of two relations, either $NS = (\text{maximum vertical displacement})/v_f$, or $NS = (\text{maximum strike-parallel displacement})/h_f$.

In the example here, the net slip is equal to $1.7 \text{ m}/v_f$. We determine in this way that the ratio of strike-slip to net slip was -0.755 (left lateral), and the net slip was 4.1 m. The result is shown in Figure 4A. The strike-slip (left-lateral) is about 3.1 m and the dip-slip (reverse) is 2.7 m, so the slip ratio of strike-slip to dip-slip is $R = 1.2$. Note that this ratio is not so different from the ratio $R=1.3$ determined by measurement of scarps and uplifts along the Sylmar segment; also, it is quite different from the ratios $R=0.38$ for the Reservoir segment, $R=2.4$ for the Tujunga segment and $R=2.9$ for the Mission Hills segment.

We have adjusted several variables and satisfied several criteria for deciding which model best fits the observed data on displacements. We have varied the dip, size and shape of the fault plane, and the depth to the top of the fault plane. We have matched, roughly, the different displacement directions in the hanging wall and footwall (Figure 5 and Figure 4). We have matched the asymmetric pattern of contours of vertical uplift. We have matched the magnitudes of vertical uplift and strike-parallel displacements.

We have only roughly matched the relative magnitudes of displacement in the footwall and hanging wall, though. The ratio of maximum horizontal displacements in the hanging wall and footwall according to the theoretical model is 2 to 2.1. The horizontal displacements are about twice as large in the hanging wall. According to Figure 5, the ratio of maximum displacements in the hanging wall and footwall is somewhat larger, 2.2 to perhaps 2.8.

Our value of 4.1 m for net slip on the fault plane, of course, is a function of the model that we selected. It is nearly twice the net slip estimated by Bonilla et al., (1971) and Yerkes et al. (1974). Savage et al. (1975) assumed a single fault and used the analysis of Mansinha and Smylie (1971), just as we did. They estimated a net slip of 2.8 m (2 m strike-slip, 2 m dip slip). They modeled the entire 15 km of surface breakage along the San Fernando fault zone whereas we modeled only the 2 km Sylmar segment and the adjacent 4 km of blind segments. They modeled a fault plane 8 km wide (dip direction) whereas we modeled a fault plane 3 km wide. Both sets of calculations suggest a dip of 45° for a single fault.

Alewine and Jordon (1973) used tens of dislocation elements to model the entire 15 km surface rupture and extended the elements to the hypocenter. They calculated a net slip of about 4 m at a

depth of 1 km, 1 m at a depth of 4.5 km and 3.5 m at a depth of about 12 km at the hypocenter. Their model fault extended from the ground surface 14 km to the vicinity of the hypocenter. They took into account the step between the trace of the Sylmar rupture zone and the Tujunga segment of the fault zone. This provide a better model of changes in elevation for the San Fernando fault zone as a whole, and a satisfactory model for the vicinity of the Tujunga segment (Alewine and Jordon, 1973, fig.3).

Trifunac (1974) assumed a model consisting of 11 dislocations each 2 km in width (slip direction) and inverted the displacement data to determine the net slip on each of them. The one near the ground surface, which represents the Tujunga segment, had a net slip of 5.2 m. The one below that, but including the trace of the Sylmar segment had a net slip of 7.5 m.

Therefore, our estimate for net slip appears to be consistent with values derived with a wide range of techniques, including static and dynamic analyses of faulting during the 1971 San Fernando earthquake.

HORIZONTAL STRAINS

There is an independent check on the solution that we have suggested (Figure 4). We can use the theoretical model to compute values of strains and patterns of strains, without any new parameters, so we can independently compare the theoretical patterns and magnitudes of strains with the measured values. Thus, we can complete our analysis of the deformation near the Sylmar segment by considering strains in the ground surface.

Several investigators have computed strains in San Fernando Valley, using the data on horizontal displacements such as those shown in Figure 5 (Yerkes et al., 1974). The data points, though, are so far apart that the measurements describe an average strain consisting of a combination of high local strains and nearly zero strains elsewhere so the pattern is not especially illuminating. The data are insufficiently detailed to see abrupt changes in strains or strains that reflect deformations near fault zones. Since our purpose is to identify large ground deformation near fault zones, we need strain data on a tight net.

In a series of papers on ground deformation during the Landers (1992) and Northridge (1994) earthquakes, detailed horizontal surveys of tight nets before and after the earthquakes have been used to determine patterns of strains associated with ground rupture (Fleming and Johnson, 1997; Johnson et al., 1994, 1997) or with what appear to be blind faults (Johnson et al., 1996; Cruikshank et al., 1996).

For our analysis of ground deformation associated with the San Fernando earthquake, we computed strains using a reference street survey (ranging from 1928 to 1970) and post-earthquake street survey (1970 to 1973). Within the city of Sylmar, we analyzed data at 563 intersections. Some of the data are for different quadrants, or corners, of the same intersection, so the total number of strain measurements in the Valley is about 1100. For simplicity, we identify streets trending NE-SW as streets *a* and streets trending NW-SE as streets *b*. The four quadrants at each intersection are generally oriented in the cardinal directions. The data are restricted to the City of Los Angeles, but the extent of the area we studied is shown in Figure 4. The western limit is near Balboa Boulevard, in Granada Hills. The northern limit is Olive View Hospital. The eastern limit is the western end of the Tujunga segment. In the south, the limit is near Rinaldi or Maclay streets. The area includes the entire length of the surface rupture of the Sylmar segment as well as the blind Granada Hills segment described by Davidson and Johnson (in prep.). It excludes the projected, blind Indian Canyon segment, because we have no data in that area.

Computation of Strains

The computation of strains uses changes in lengths of streets intersecting at a large angle (generally 60 to 120°) and the change in the angle of the intersection (Figure 13).

Figure 13. Quantities measured at a street corner in order to determine strain.

Most intersections of city streets have a permanent subsurface monument so the surveys are from monument to monument. However, measurements do not have to be from intersection to intersection. The lengths of the streets are generally between 50 and 200 m long. Therefore, the distances reflect the irregular grid we analyze for strains. We have assessed the monumentation and accuracy of the strain measurements elsewhere (Johnson et al., 1996; Cruikshank et al., 1996) and will only briefly summarize it here. We assume that strains greater than $10^{-3.5}$ are significantly greater than the errors of measurement. The City of Los Angeles regards their determination of distances to be accurate to one hundredth of a foot (0.01 ft) or 3.1 mm. For a street length of 100 m, the normalized error would be 3×10^{-5} . However, the error will be larger for shorter streets; we estimate an error of strain of 10^{-4} for a 50 m street and an error of 2.6×10^{-5} for a 200m street. Angle measurements should contain errors less than 2 seconds. Assuming 3 seconds of error, the angular error corresponds to an apparent shear strain of 1.45×10^{-5} . Thus, angle measurements have an error that is comparable or less than the distance measurements, and shear strains can be determined to about 10^{-5} . Since we are comparing two measurements which may contain these errors, the magnitude of the real error can be twice the magnitude reported a single measurement. In interpreting the strain patterns for this study, we assume that strains must be at least $10^{-3.5}$ to be significant.

The components of strain are related to the components of the deformation gradient, such that,

$$\epsilon_{ij} = \frac{1}{2} \left(\frac{\partial x_i}{\partial X_j} + \frac{\partial x_j}{\partial X_i} \right) \quad (1)$$

in which X is the original position and x is the final position of a material point. However, only three components of strain can be calculated because street surveys do not contain data on strain in the vertical direction. In two dimensions, $x = x_1$, $y = x_2$, we have three strain components:

$$\epsilon_{xx} = \frac{1}{2} \left(\frac{\partial x}{\partial X} \right)^2 + \left(\frac{\partial y}{\partial X} \right)^2 - \frac{1}{2} \left(\frac{\partial x}{\partial Y} \right)^2 - \left(\frac{\partial y}{\partial Y} \right)^2 \quad (2a)$$

$$\epsilon_{xy} = \frac{1}{2} \left(\frac{\partial x}{\partial X} \frac{\partial x}{\partial Y} + \frac{\partial y}{\partial X} \frac{\partial y}{\partial Y} \right) \quad (2b)$$

$$\epsilon_{yy} = \frac{1}{2} \left(\frac{\partial x}{\partial Y} \right)^2 + \left(\frac{\partial y}{\partial Y} \right)^2 - \frac{1}{2} \left(\frac{\partial x}{\partial X} \right)^2 - \left(\frac{\partial y}{\partial X} \right)^2 \quad (2c)$$

We use the street surveys to determine the four unknown components of the displacement gradient, such as $\frac{\partial x}{\partial X}$ and $\frac{\partial y}{\partial X}$ in eqs (2). Since we do not know absolute orientations of streets, only the angles between streets, we are unable to determine the rotation. Thus, we assume that one street (usually street *a*) remains fixed; whereas, the other is reoriented by the deformation. Therefore, we assume that the shearing is simple shearing. In this case, $\frac{\partial y}{\partial X} = 0$, so there are only three unknown components of the deformation gradient. Since the absolute orientation is not known, the rotation of line parallel to the maximum and minimum extension directions can not be determined.

The other components of the deformation gradient are computed as follows: Define *A* as the initial and *a* as the final lengths of a street segment oriented about N45°E. As indicated above, we assume simple shearing is parallel to *a*. Therefore, one of the components of the deformation gradient is approximately

$$\frac{\partial x}{\partial X} = \frac{a}{A} \quad (3a)$$

The other components of the deformation gradient are defined in terms of *B*, the initial length and *b* the final length of the intersecting street, as well as θ the initial and θ' the final angle which is positive in a counter-clockwise direction between street segments *a* and *b* (Johnson et al., 1996; Cruikshank et al., 1996). The components are:

$$\frac{\partial y}{\partial Y} = \frac{b}{B} \frac{\sin(\theta')}{\sin(\theta)} \quad (3b)$$

$$\frac{\partial x}{\partial Y} = \frac{b}{B} \frac{\cos(\theta') - \frac{a}{A} \cos(\theta)}{\sin(\theta)} \quad (3c)$$

Since we can measure the four quantities, *a/A*, *b/B*, θ and θ' , with the resurvey of the streets, we have enough information to compute the components of the deformation gradient and, through eqs. (2), compute the strains. With the strain components, we can compute the principal values and orientations in the usual way. The methods of computing strain with resurveys of streets are explained in detail elsewhere (Johnson et al., 1996; Cruikshank et al., 1996).

Strains Computed from Street Surveys

In this analysis we have used concepts of strain in order to derive principal directions of extension or compression in terms of measurements that one can make at a street intersection (Figure 15), assuming small gradients of strain over lengths of streets and assuming continuity of streets.

It is important to realize that we actually cannot determine strain in many places with the survey data. Strain is defined at a point, whereas the length changes we measure are over the length of a city block. There are two problems with calling such measurements strain. First, the actual deformations may be highly localized, say along fractures, so the strain calculated for a street is partly determined by an arbitrary, street length that is used to compute the strain. The strain computed will be higher if the measurement points are closer together. Second, the strains may vary considerably over the length of a street.

In previous papers by Johnson et al. (1996) and Cruikshank et al. (1996), they have carefully differentiated between strains and “normalized length changes” in order to avoid misusing the term strain. In this paper, we will use the term strain but ask the reader to remember that the so-called “strains” could be entirely due to fracturing and faulting. Thus, we use strain theory to compute the principal directions and magnitudes of the principal strains.

Plotting of Data

In the previous studies, strains were reported as “shmoos” and “nerds”. Here we report strains with squares and rectangles. The square and rectangle are oriented with sides parallel to the principal directions. The area of the square is related to the order of magnitude of the largest absolute value of principal strain, which we indicate by ϵ_1 or ϵ_2 . Thus, the width of the square is

$$W = w \sqrt{\frac{|\epsilon_{\text{ref}}|}{10^{-n}}} \quad (4a)$$

in which w is the width of the square in a diagram (say 1 cm), the reference strain, ϵ_{ref} is the larger of $|\epsilon_1|$ and $|\epsilon_2|$, and n is the exponent of the maximum reference strain, $(\epsilon_{\text{ref}})_{\text{max}}$, in the data set. The widths of the squares represent a range of magnitudes and are shown as a scale of squares in Figure 15. In this example, the orders of magnitude range from 1.5 to 3, that is, the strains range from $10^{-1.5}$ to 10^{-3} .

The rectangles shown in Figure 14 and 15 for every strain figure are defined as follows: The line defining the square represents zero length change. If a principal strain is positive, the rectangle is longer than the length of the square in that direction. If a principal strain is negative, the rectangle is shorter in that direction. The lengths of the sides of the rectangle are,

$$L_1 = w \left[\sqrt{\frac{|\epsilon_{\text{ref}}|}{10^{-n}}} + \text{sgn}(\epsilon_1) \sqrt{\frac{|\epsilon_1|}{10^{-n}}} \right] \quad (4b)$$

$$L_2 = w \left[\sqrt{\frac{|\epsilon_{\text{ref}}|}{10^{-n}}} + \text{sgn}(\epsilon_2) \sqrt{\frac{|\epsilon_2|}{10^{-n}}} \right] \quad (4b)$$

Figure 14. Idealized strain patterns at the ground surface for various senses of slip.

in which $\text{sgn}()$ is the sign of the argument. This is how we define the rectangular strain figures shown in Figure 5.¹¹

Strains

We have used the approach described above to calculate principal strains throughout much of the San Fernando Valley. For this paper, we will report only the strain results relevant to the Sylmar segment (Figure 5). Davidson (2000) presents all of the raw and processed data, including the strains throughout the City of Los Angeles in the San Fernando Valley.

The strain figures centered on the San Fernando fault zone do not truly represent strain (Figure 15) because the streets measured there were broken by one or more fractures or localized shear zones; the assumption of continuity is violated in these areas (Johnson, et al., 1996). We can use the strain figures, nevertheless, to indicate direction of shift across a fault or fault zone.

¹¹ For a copy of the computer program that performs the computations, see Davidson, 2000, appendix.

Figure 15. Strains determined by street surveys by Survey Division, City of Los Angeles, in the vicinity of the Sylmar and Mission Wells fault segments. (from Davidson, 2000, fig. 4.5).

The strains measured at San Fernando are similar to those measured in deformation zones at Winnetka and Granada Hills following the 1994 Northridge earthquake. Excluding the strains of $10^{-1.5}$ along the Sylmar segment where survey lines certainly cross major surface ruptures, the strains are commonly 10^{-3} to 10^{-2} .

The pervasive ground fracturing at San Fernando (Kamb et al., 1971; Sharp, 1971, 1975) was also similar to that at Granada Hills and the NE end of the Winnetka rupture zones (Johnson et al., 1996; Cruikshank et al., 1996). The correlation of strains of magnitudes 10^{-3} to 10^{-2} with ground fracturing is consistent with strains required to fracture rock, concrete and unconsolidated soil near the ground surface. A brittle rock, such as granite (e.g., Peng and Johnson, 1972), fails under atmospheric confining pressures at compressive axial strain values of -2 to -6×10^{-3} . In the Aspen Grove landslide in Utah (Baum et al., 1988, 1993; Fleming, and Johnson, 1989, 1993) the soil in a developing landslide toe cracked at strains of about -1.4 to -2×10^{-2} and in a newly forming pull-apart basin, the strains were about 6×10^{-3} before a tension crack formed and 3.5×10^{-2} after the

tension crack had formed. Because of the pervasive fracturing, we refer to the strain measurements as strain measurements.

COMPARISON OF COMPUTED AND MEASURED STRAINS

We have determined the fault dip and the dip- and strike-slip on a simple, planar fault that matches the height (1.7 m) of the bulge as well as the maximum (1.6 m) lateral displacement in the hanging wall of the Sylmar segment. Here we use those same parameters to compute the pattern and magnitudes of strains according to the theoretical model. The strain pattern, therefore, provides an independent test of the model to predict the kinematics near the Sylmar segment.

We computed the strains and plotted them at the center point, between four grid points where displacement components were computed. The strains are displayed in the theoretical example as they are in the field example in Figure 5. In both cases, strains smaller than $10^{-3.5}$ are neglected (shown as a circle with a diagonal line).

The strain figures shown along the surface rupture are large and include left-lateral shearing (extension NW-SE, compression NE-SW). The strains are large because legs of the grid points are faulted. Nevertheless, the strains measured with field data (Figure 5, 10^{-2} to $10^{-1.5}$) are nearly an order of magnitude larger than those computed with the best-fit parameters (10^{-3} and $10^{-2.5}$). The strains are computed in essentially the same way by the two methods, but the spacing of grid points is quite different. The lengths of the streets typically range from 50 to 200 m, with an average of 100 m. The grid points used to compute the strains shown in Figure 14 are separated 500 m. In order for the strains to be comparable along the main rupture, one should reduce the spacing of grid points in the theoretical example by a factor of five. Suppose we multiply the theoretical strains along the fault trace by five. Thus, the strains along the fault trace would be about $10^{-2.3}$ to $10^{-1.8}$, which are slightly smaller than those calculated with survey data.

The strains computed for elements in the hanging wall, over the fault plane, generally range between $10^{-3.5}$ and $10^{-2.8}$ and include right-lateral shearing (Figure 14). Strains computed from surveys in the wide belt at the east end of the Sylmar segment are between 10^{-3} and 10^{-2} and are also right-lateral (Figure 5). Again, assuming that, because of localized fracturing, the grid spacing must be the same to obtain comparable results, we would obtain normalized length changes of $10^{-2.8}$ and $10^{-2.1}$. Thus, the strains computed with the theoretical model are quantitatively similar in magnitude and they show the same pattern as the field measurements: left-lateral shearing at the fault, right-lateral shearing in the hanging wall and negligible shearing in the footwall. In the field, of course, there is not a single fault but rather a zone of left-lateral shearing. The dislocation model does not relate well to a zone of shearing. It is for a plane. Nevertheless, the theoretical and computed strains are remarkably similar, given the fact that the theoretical model was adjusted for maximum horizontal and vertical displacements but not for strains.

PARALLEL BELTS OF OPPOSITE SHEARING

Belts of Shearing Strain

The shear strains also appear in belts and reflect left-lateral and right-lateral shearing rather than compression and extension. The belts of left-lateral and right-lateral shearing strains are shown with dashed lines in Figure 5. Centered on the main rupture along the Sylmar segment is a belt of left-lateral shearing about 500 to 750 m wide (Figure 15). It is just north and south of the main rupture of the Sylmar and Mission Wells segments. We do not have data west of the Mission Wells segment or east of the Sylmar segment to see if this belt includes the entire San Fernando fault zone. Kamb et al. (1971, p. 45) recognized a "zone of surface deformation" within which roads

trending NE-SW were buckled and thrust, as in a left-lateral zone, and within which "major structural damage to buildings is confined."

The belt of right-lateral shearing within the hanging wall of the Sylmar segment is perhaps more surprising than the belt of left-lateral shearing. The belt of right-lateral shearing is 250 to 1200 m wide. It includes the belt of tension cracks and right-lateral offsets on open cracks. West of the tension cracks but still in the area of right-lateral shearing, there are no known faults or tension cracks, but there are elevated strains of magnitude $10^{-2.5}$ and smaller (Figure 5).

According to Bonilla et al. (1971, p. 57), "...the zone of fractures is divided into a narrow southern zone of [left-lateral] shearing and thrusting and a wider northern zone of extension..." In fact, their map (1971, fig. 2) shows a zone of left-lateral rupturing and thrusting along the surface trace of the Sylmar segment and a zone of tension fracturing, normal faulting and right-lateral faulting north of the left-lateral zone.

County Line Blind Segment

We have shown in a companion paper (Davidson, 2000) that the field of horizontal displacements has a velocity discontinuity that appears to coincide with the traces of surface ruptures of the San Fernando fault as well as with paths of blind segments without clear surface ruptures. We showed there that the San Fernando fault zone probably ruptured as far as 10 km west of the west end of the Sylmar segment, including short surface ruptures of the Reservoir and Mission Wells segment (Figure 5).

The contours of vertical displacement suggest that the Sylmar segment may be continuous with a blind segment, the County Line blind segment, that extends about 1 km west of the west end of the Sylmar rupture (Figure 15). The County Line blind segment may be unconnected to the Mission Wells segment.

The strain measurements provide another set of data relevant to the length of the County Line blind segment. The two belts of shearing—left-lateral shearing on either side of the Sylmar surface rupture and right-lateral adjacent to this belt, but to the north—extend about 2 km west of the west end of the trace of the Sylmar rupture.

We would suggest that the Sylmar segment and its adjacent blind segments represent a fault about 6 km long. The fault includes two blind segments—the 3.0 km long blind Indian Canyon segment in the east and the 1.0 km long County Line blind segment in the west—and the 2 km long Sylmar surface rupture.

It is uncertain whether there is a blind fault connecting the Sylmar and Mission Wells segments. There is no differential uplift across the trace of such a blind fault so the blind faulting would have a different character from that of the Sylmar surface rupture. Field investigations indicate that the faulting is largely left lateral along the Mission Wells segment, perhaps the same is true along the supposed blind segment connecting the Sylmar, and Mission Wells ruptures.

DISCUSSION

Belts of Shearing along Earthquake Ruptures

Surface Ruptures

This paper is another part of a series of papers on belts of shear strain along faults that shifted during a large earthquake in California. Belts of shearing were first suspected in the study of the

Summit Ridge shear zone that activated during the 1989 Loma Prieta, California, earthquake (Johnson and Fleming, 1993; Martosudarmo et al., 1997). The Summit Ridge shear zone was a coactive, tectonic structure, even though it has been discounted by Ponti and Wells (1991); Prentice and Schwartz (1991) and Holzer (1994). Although they denied that such shear zones exist, we would note that belts of shearing were identified along the main rupture of the San Andreas fault zone by G.K. Gilbert and Andrew Lawson following the 1906, San Francisco earthquake.

Belts of shearing that were tectonic in origin became obvious to most investigators and were documented along the surface rupture of the 1992, Landers, California, earthquakes (Fleming and Johnson, 1997, Johnson et al., 1994, 1997, Lazarte et al., 1994, Fleming et al., 1998). The belts of rupturing ranged in width from a few tens of meters to perhaps a kilometer. We mapped the ruptures and measured strains in their vicinity by resurveying benchmarks that happened to be available. The conclusion that the ground rupturing at Landers consisted of broad belts of deformation and fracturing was unchallenged by the colleagues who denied such structures at Loma Prieta.

Following the 1994 Northridge, California earthquake (6.7 Ms, Shen et al., 1996), we identified two areas of possible blind faulting and used surveys of street intersections, by the City of Los Angeles, to determine patterns of strains. Strains were determined by comparing pre-earthquake surveys of the same bench marks by the City of Los Angeles after the 1994 Northridge earthquake sequence. Belts of elevated strains within the Winnetka deformation zone (Cruikshank et al., 1996) have strains with magnitudes of about 3×10^{-4} and locally 10^{-3} . The Winnetka deformation zone trends diagonally from near Canoga Park in the southwest, through Winnetka, to near Northridge in the northeast, a distance of approximately 4.5 km. The deformation zone is about 1 km wide. The northwestern two-thirds of the zone is predominantly a belt of extension and the southeastern one-third is predominantly a belt of shortening. On the northwest and southeast sides of the deformation zone the magnitude of the strains is too small to measure, less than 10^{-4} (Cruikshank et al., 1998).

A deformation zone of the same type, but shorter, was identified in Granada Hills (Johnson et al., 1996).

We would suggest that blind tectonic ruptures be identified by peculiar patterns of strains or by fracturing within belts along the trace of the suspected blind fault that can be explained in terms of the strain and displacement fields of a blind fault. Since there were no surface ruptures of faults in the San Fernando Valley during the 1994 Northridge earthquake, our conclusion was that there were active, near-surface blind faults at Winnetka and Granada Hills areas during the 1994 earthquake. A fault model indicated that the deformation zones could have been produced by near-surface blind faults.

Again, the deformation zones have been generally ignored or disputed. The deformation at Granada Hills has been ascribed by Hecker et al. (1995a, 1995b) to unknown processes such as slope failure, compaction, sediment lurching, or liquefaction. The only definite conclusion was that it was not tectonic deformation. The deformation zones at Winnetka have been dismissed as results of shaking failure of a soft substratum (Holzer et al., 1996). Perhaps the study of fractures in the San Fernando Valley following the 1971 earthquake by Sharp (1971) was a premonition of stands to be taken relative to ground fracturing during earthquakes. Sharp studied fractures outside the zones of obvious tectonic rupturing along the San Fernando fault zone to determine whether the fractures could be explained by non-tectonic causes. Tectonic fractures, he argued, "are primarily recognized by their continuity and map pattern, pattern of displacement, and indifference to the position of overlying constructed works." It is not surprising that his conclusion was that the fracturing could be explained by other, generally obscure and undocumented, causes.

In spite of these disputations, the idea that the belts of deformation at San Francisco, San Fernando, Loma Prieta, Landers, and Northridge are due to tectonic causes seems inescapable.

Blind (Subsurface) Ruptures

Intensive study of the San Fernando earthquake by the geophysics community provided considerable information on displacements and strains in the ground surface. In terms of these quantities, we have been able to explain much about the patterns of displacements and strains with a fault/dislocation theoretical model. Displacements measured immediately after the San Fernando earthquake provide valuable information about the nature of the faulting, whether blindly or rupturing the ground surface. Surface fracturing demonstrates that faults did, in fact, reach the ground surface in some places.

We can also use many of the results to look back on studies of ground rupturing in other earthquakes and place some of the studies into perspective. If we consider only the vertical displacements and the horizontal displacements perpendicular to the fault trace (Figures 6A and 6B) we see that the deformation includes belts of compression and extension with the direction of maximum compression or extension at right angles to the trace of the fault. There is a belt of extension in the hanging wall block, a belt of compression along the fault, and another belt of extension in the footwall block. If the fault dips 45° , two belts of roughly equal extension form on either side of a belt of greater compression (Figure 6B). If the fault dips at 30° , there is only one belt of extension in the hanging wall above the fault and one belt of compression along the fault (Figure 6A). The extensile strains in the footwall are smaller.

We see results of this pattern of deformation along the Sylmar segment of the 1971 San Fernando earthquake, and along the Winnetka and Granada Hills shear zones of the 1994 Northridge. At Sylmar, we see compression; primarily in the form of low-angle thrusts in the streets near the Sylmar rupture zone and extension in a belt of cracking and normal faulting parallel to the Sylmar rupture in the hanging wall (Kamb et al., 1971; Bonilla et al., 1971). Along the Granada Hills shear zone we see parallel belts of extension in the north and compression in the south, separated by a narrow zone that includes some left-lateral shearing (Johnson et al., 1996). The belts of strain, though, only extend laterally for about 0.5 km.

The case is much clearer along the Winnetka deformation zone. There were two, parallel strain belts, with seemingly contradictory states of strain, much like the belts of right- and left-lateral shearing at Sylmar. The directions of principal strains are parallel and normal to the walls of the zone. The magnitudes of strain are generally about 3×10^{-4} and locally 10^{-3} . The northwestern two-thirds of the zone is a belt of predominant maximum extension normal to the walls of the zone. The southeastern one-third of the deformation zone is a belt of predominantly maximum shortening normal to the walls of the zone. On the NW and SE sides of the deformation zone the magnitude of the strains is too small to measure (less than 10^{-4}). The strain pattern is quite similar to that for a reverse fault dipping about 30° N, as shown in Figure 6B.

The conclusions made (Johnson et al., 1996; and Cruikshank et al., 1996) that the Granada Hills and Winnetka deformation zones are surface manifestations of blind faults are quite consistent with the theoretical modeling and, more importantly, they agree with the field measurements and resurveying of ground deformation after the San Fernando Valley earthquake. Here the displacements are quite well known and strains were measured adjacent to faults that broke through to the surface as well as faults that remained blind.

Coactive Faults

We termed the hypothesized faults responsible for the Granada Hills and Winnetka deformation zones coactive faults, indicating that the faulting may represent movement of smaller faults within

the epicentral area activated about the time of the earthquake shock on the main fault. These can be the faults or shear zones that produce aftershocks or that move aseismically. Often, the coactive faults will have a different kinematic signature than the main rupture (Johnson, et al., 1996). Other earthquakes where coactive faulting is evident are the 1994 Northridge, 1992 Landers, 1989 Loma Prieta and the 1968 Borrego Mountain earthquakes.

Probably the most extensive coactive faulting was during the San Fernando earthquake along the Santa Susana fault, near the junction of Interstate 5 and Foothill Boulevard (Barrows et al., 1974). Barrows et al. (1974) mapped the ground rupture as the trace of a low-angle fault extending 0.5 km from the NE side of Interstate 5 southwestward to about 1.5 km SW of Interstate 5. According to Proctor et al. (1972, p. 1603), the Santa Susana fault also extends southwestward. During the 1971 earthquake the Santa Susana fault was displaced in a left-lateral sense as much as 30 cm along a fault trace extending about 1 km southwest of the Golden State Freeway. We were unable to compute strains in the area of the Santa Susana fault because surveys were lacking.

Field observations indicate that the Veterans fault moved during the earthquake sequence. According to Kamb et al. (1971, p. 43), the surface rupture of the Veteran's fault extends only a short distance laterally and faults are parallel to bedding, dipping 60° north. "Although it is a small fault, it is especially significant because it is the only definite tectonic displacement found in the zone along the foot of the San Gabriel Mountain escarpment, where the greatest intensity of observed ground shaking occurred." We computed slightly elevated strains near the Veterans fault (Figure 5).

Besides the Veteran's fault, there are smaller, but still elevated, strains in the vicinity of several suspected faults in the San Fernando valley. The strains in two areas are about $10^{-2.5}$. One is the Fenton fault (Figure 5). There is no surface rupture along the Fenton fault so its activity is indicated only by strain data. Barrows et al. (1974) mapped the Fenton fault as questionable. The strain data suggest that the Fenton fault was active during the 1971 earthquake sequence, although the location of the fault may be slightly different from where it was mapped. We would extend the fault slightly north of its mapped location in Figure 5. Barrows et al. (1974) also marked the DeGarmo fault as questionable. The measurements of strain near this suspected fault are quite small, so the DeGarmo fault may not have been active during the 1971 earthquake.

We see, therefore, that the San Fernando earthquake produced several examples of coactive faulting. These were unrecognized at the time of the earthquake and the surface fracturing was dismissed as unrelated to slip along even known faults (Sharp, 1971; Bonilla et al., 1971).

Shape of the Sylmar Segment

Figure 7 shows a sketch of the shape of the San Fernando fault zone according to Sharp (1975). Our work has been mainly along the Sylmar segment. The diagram shows that the Sylmar segment is stepped back (north) relative to the Tujunga segment and that the Sylmar and Tujunga segments overlap one another by at least 2 km. Perhaps the most important inaccuracy of the sketch in Figure 7 is the steepness of the faults. Sharp (1975) indicates, "The array of inferred and observed fault dips suggests that the Mission Wells and Sylmar segments on the valley surface dip steeply to vertically..." All the analyses of displacements caused by faulting indicate that the Sylmar segment fault, indeed the entire San Fernando fault zone, dips about 45°, much shallower than shown in Figure 7.

Figure 16. Sketch of subsurface shape of San Fernando fault according to Sharp (1971). Approximate shape of eastern part of San Fernando fault zone according to Sharp (1975). The eastern, Tujunga segment is shown as a low-angle thrust near the ground surface but generally dipping about 45° . The Sylmar segment is shown as vertical near the ground surface but generally dipping about 45° ; our results suggest that it is dipping about 45° , even near the ground surface. The Sylmar segment is shown as a separate blade of the San Fernando fault zone, as is a blind segment to the east. The Sylmar blade overlaps the Tujunga blade by 3 to 4 km. Our results agree with this interpretation. Our results suggest that the Sylmar blade is disconnected from the Mission Wells segment to the west and that the San Fernando fault zone, including the Mission Wells and Reservoir segments, extends perhaps 10 km west of the Sylmar blade, much farther than shown in this figure.

Sharp shows the eastward extension of the Sylmar surface rupture mostly as a blind fault. Our explanation of the extension of the dome of uplift in this area as a reflection of a steep blind fault is consistent with Sharp's interpretation. He shows blind segments between the Sylmar, Mission Wells and Reservoir segments. He also shows the tip of a blind segment which dives downward at the western end (right). Again, our interpretation is consistent with his and, furthermore, we would extend the blind fault much farther west, at least to Balboa Boulevard (Yerkes et al., 1974; Davidson, 2000).

CONCLUSIONS

Following are the conclusions of the research:

?? *Parallel strain belts are associated with surface faulting.*

There were shear belts along the Landers strike-slip rupture and along the Sylmar strike-slip/reverse rupture but the shear belts were quite different. The right-lateral faulting at Landers was mostly within vertical rupture zones. The rupture zones consisted of a belt, generally ranging in width from about 50 m to at least 500 m, where the deformation was right lateral. There were no belts of rupture on either side of the central belt of right lateral shearing; belts on either side were invisible except for small left-lateral (rebound) strains.

The Sylmar rupture dips at about 45° and the strike-slip is left-lateral. The rupture consisted of a belt of left-lateral rupturing ranging in width from 500 to 700 m, roughly centered on the trace of the rupture. The belt in the hanging wall, adjacent to the left-lateral belt but situated over the fault, contained right-lateral (rebound) strains of sufficient magnitude to fracture the ground. The belt of right-lateral shearing in the hanging wall was 700 to 1200 m wide and was highly fractured. The belt in the footwall, adjacent to the left-lateral belt, had very small (rebound) strains.

Thus, there was one rupture belt produced by strike-slip faulting along a vertical fault at Landers and two rupture belts produced by strike-slip faulting along a fault dipping perhaps 45° at San Fernando.

The Sylmar rupture also accommodated thrusting, and there are two belts associated with this deformation. These belts are superimposed on those produced by left-lateral faulting. There is a belt marked by buckling and thrusting of streets, generally within the belt of left-lateral shearing. There is another belt of extension fracturing and normal faulting within the belt of right-lateral shearing.

?? *Parallel strain belts are also associated with blind faulting.*

One of the more important results of our study of strains along the Sylmar segment is the recognition that the parallel strain belts continue westward, beyond the tip of the surface rupture of the segment. The right-lateral belt continues for perhaps one km west of the Sylmar rupture and the left-lateral belt continues even farther. Thus, strain belts form over some blind segments.

?? *The drastic difference in the horizontal displacement components parallel to the strike of the fault in the hanging wall and footwall blocks of the Sylmar fault is a result of strike-slip displacement across a gently dipping ($\sim 45^\circ$) fault.*

One of the most striking features of the 1971 San Fernando earthquake is that the horizontal displacements are so discordant across the rupture zone of the Sylmar segment. In the hanging wall, the displacements are nearly parallel to the trace of the Sylmar segment. In contrast, in the footwall the displacements are nearly at right angles to the trace of the Sylmar segment (Figure 5).

The theoretical model helps us understand this displacement pattern. The model shows that the displacement pattern is the result of a combination of reverse- and strike-slip faulting that dips gently. When we investigate the displacements in a model with a gently dipping reverse fault, we see that the displacements are at right angles to the fault trace. Near the fault trace the displacements are larger in the footwall than in the hanging wall. In contrast, when we investigate displacements in a model with a gently dipping strike-slip fault, we see that the displacements are parallel to the fault trace and are much larger in the hanging wall than in the footwall. Hence, the displacement pattern that we observed along the Sylmar segment of the San Fernando fault zone is a result of reverse and strike-slip faulting on a gently dipping fault.

Furthermore, the displacement pattern produced by a low-angle fault disappears if the fault is high-angle (60 to 90°). Therefore, we know that the Sylmar segment is not a high-angle fault.

- ?? *A distinctive bulge in the hanging wall north of the rupture of the Sylmar segment probably reflects reverse faulting and change in dip of the fault along its length.*

There is a very distinctive bulge or dome of uplift extending for about 6 km E-W in the hanging wall block of the Sylmar segment. The maximum height of the dome is 2.2 m. Near the west end of the Sylmar segment the dome is highly asymmetrical. The height increases from 0.5 m to 2 m uplift over a horizontal distance of about 0.5 km centered on the trace of the Sylmar segment, and decreases from about 2 m to 0.5 m over a horizontal distance of about 2 km to the north of the trace of the Sylmar segment. Theoretical modeling of a fault dislocation suggests that the asymmetric dome is a result of reverse faulting on a fault dipping about 45° northward.

Near the east end of the Sylmar segment, and in an area extending 3.0 km to the east of the eastern end of the Sylmar segment, the dome is nearly symmetrical. Furthermore, its axis is offset considerably from the trace of the closest parallel surface rupture, the Tujunga segment. We suggest that there is a blind fault extending about 3.0 km east of the east end of the Sylmar segment, so that the Sylmar and Tujunga segments overlap about 3 km. The theoretical model can explain the symmetric form of the dome at its eastern end if the blind fault becomes shallower, perhaps 30°, to the east. This is what we suggest.

- ?? *The modeling of the vertical and horizontal displacements and the strains at the ground surface near the Sylmar segment and its blind extensions is remarkably simple.*

We have modeled the Sylmar segment (and blind segments) with a fault dislocation dipping northward about 45°, about 6 km long (E-W) and 3 km down dip, with a particular combination of left-lateral slip and reverse slip. The net slip is about 4.1 m. The strike-slip component is about 3.1 m (left-lateral) and the dip-slip component is about 2.7 m (reverse). This fault produces an asymmetric bulge of uplift in the hanging wall with a height of 2.2 m, as observed, and produces a large left-lateral displacement of the hanging wall block near the Sylmar rupture of 1.6 m, as observed. The strains are similar in magnitude and in pattern to the strains computed from resurveys of City of Los Angeles streets.

- ?? *Contrary to early reports, there was widespread coactive faulting during the 1971 San Fernando earthquake.*

Several early investigators of ground rupture during the San Fernando earthquake argued that the only tectonic rupture in the Los Angeles area at the time of the 1971 earthquake was along the San Fernando fault zone. Several early studies and several later studies showed unequivocally that some other faults moved coactively with the rupture of the San Fernando fault zone. The Veterans fault moved and produced ground rupture. Segments of the Santa Susana thrust fault ruptured over a length of several kilometers, extending at least 3 km west of the Golden State Freeway. The Fenton fault south of Olive View Hospital apparently produced elevated strains and some ground rupture.

ACKNOWLEDGEMENTS

This research would not have been possible without the generous and enthusiastic efforts of the Survey Division, City of Los Angeles¹². In particular, we wish to thank William Prange, Surveyor, Robert Jones, Surveyor, Robert Renison, Supervisor of Surveying, Robert F. Packard, ret.

¹² Bureau of Engineering, Department of Public Works, 16th floor, 600 S. Spring, Los Angeles, California 90014

Engineer of Surveys and Glen Nave, Engineer of Surveys. All these people have been extremely helpful as we worked with their data.

This paper is part of a broader study, a collaboration of the City of Los Angeles and Purdue University through grants from the National Science Foundation EAR 9416760 and the Department of Energy grant DE-FG02-98ER14886, supervised by Nick Woodward of DOE, and support from the Department of Geosciences, Purdue University. We appreciate the support of these organizations.

Faux Pli Software

Mechanical Simulation of Fault-Related Folds (FaRF)

A Manual Introducing and Illustrating Programs SeGaLi FaRF¹³ and SeDéRa FaRF

¹³ The computer programs, SeGaLi FaRF and SeDéRa FaRF, the algorithms underlying SeDéRa FaRF, and this document are copyrighted by Arvid M. Johnson and Purdue University. They may be copied only with permission of the copyright holder. Please read Appendix IV of Volume I.

EXAMPLES OF USE OF SEQUENTIAL GANG/LISTRIC FAULT RELATED FOLDS (SEGALI FARF)

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1st Draft: August 2000

OPENING COMMENTS

This second volume of the manual for fault-related folds is to explain certain, more advanced features of the SeGaLi FaRF program and show some ways that the program can be used professionally. None of the problems discussed here has been solved in the regular published literature. The first chapter is based on a research project on the 1971 San Fernando earthquake. I wrote the first few versions of SeGaLi FaRF to work on archival information we were able to obtain from the City of Los Angeles Survey Division about this earthquake. The other two chapters are based on data collected before this program existed for the 1992 Landers and the 1994 Northridge earthquakes. The original data are published in USGS Open-file reports, so they are not readily available in most libraries. Analyses of these data with SeGaLi FaRF have answered several questions about the forms of rupture in the various earthquakes and have led to new understanding of earthquake ruptures in general. The analysis of the Winnetka deformation zone

near Northridge is particularly nice, because in the open-file report I used an approximate solution that I derived on the back of an envelope to estimate the magnitudes of strains one would see. Although I had hoped that the theory would help us understand the peculiar strain patterns, I had no idea that the estimates of magnitudes of strain would be nearly identical or that the correlation between field measurements and the analytical solution would be so beautiful. SeGaLi FaRF has turned out to be a very useful tool for our research group in neotectonics and structural geology. In any case, analysis of some of data from field observations and resurveys of control points provide opportunities to illustrate to neotectonicists and structural geologists different ways one can use the SeGaLi FaRF program to solve practical problems of ground deformation associated with earthquakes.

In order to use this second volume, you probably will have to be somewhat familiar with one or two chapters in the first volume. One probably should become familiar with the program by taking a few minutes to follow the tutorial, "autopilot," in volume I before trying to follow the specialized applications here.

CHAPTER 1. ANALYSIS OF DISPLACEMENTS DURING THE 1971 SAN FERNANDO EARTHQUAKE

We will take a different approach in this second volume of the SeGaLi FaRF manual. Instead of focusing on generalized notions of fault-related folds, we will examine faulting and folding associated with particular earthquake events. In this chapter we focus on the San Fernando, California, earthquake of 1971, and use the theoretical model to analyze some of the field measurements and observations. The San Fernando earthquake is an important earthquake to the development of American earthquake seismology and geology because it was the first large earthquake with a combination of extensive surface rupture and widespread instrumental records in the continental U.S. since the 1906 San Francisco earthquake.

San Fernando Fault Zone

The San Fernando fault zone ruptured during the 1971 earthquakes over a total length of at least 12 km of fault zone (Kamb et al., 1971). The shift was accommodated across a fault zone about 50 m wide and the fault rupture was composed of five continuous traces or segments (Figure 1) that generally trend E-W. There were few scarps and other linear features between the segments, so a continuous fault break could not be traced from one end to the other. There was no information on fault attitude along the central, Sylmar segment. The Tujunga fault segment, to the south, though, was generally a clean break with a well-defined scarp. There the faults exposed dip 20 to 25° toward the north so they are low-angle, thrust faults, but the faults generally accommodated both left lateral and reverse shift of nearly equal amounts (Kamb et al., 1971).

Figure 1. Trace of ground rupture of the 1971 San Fernando, California, earthquake. (Compiled by Xan Jones Davidson, 1996-00).

Figure 2. Sketch of possible configuration of San Fernando fault that ruptured in 1971. (After Sharp, 1975).

The continuous segments in the vicinity of the Sylmar segment are the Reservoir, Mission Wells, and Tujunga segments. The distribution of the segments and the way Robert Sharp imagined that they might connect at depth are shown in Figure 2.

Starting in the west (Figure 1), the Reservoir segment is near the Van Norman reservoirs and is centered between the San Diego and Golden State freeways. Slip consisted of left-lateral slip of 0.16 m and reverse dip-slip of 0.38 m (The ratio, R , of strike-slip to reverse dip-slip was about $R=0.42$). Slip on this segment caused failure of the Golden State Freeway (Yerkes et al., 1974; Weber, 1975). The Mission Wells is a short segment to the east of the Reservoir segment. All the segments trend E-W except for the Mission Wells segment, which trends NE-SW. This segment had left-lateral slip of 0.46 m and reverse dip-slip of 0.16 m ($R=2.9$) (Yerkes et al., 1974). This oddly-oriented segment accommodated mostly strike-slip.

The Sylmar segment, to the east of the Mission Wells segment, extends for about 2 kilometers and is straighter and more continuous than the other segments. A peculiar feature of the Sylmar segment is that it is stepped about 1.2 km north of the general east-west trace of the San Fernando fault zone (Figure 1). According to Sharp (1975), the Sylmar segment slipped up to 2.5 meters, with a left-lateral to reverse dip-slip ratio of $R=1.3$. Kamb et al. (1971, p. 45), determined relative displacements across the Sylmar fault zone using offsets of streets and other markers and assuming that the ground in the hanging wall did not move; only the footwall shifted. They obtained measurements of 0.8 to 1.9 m of net slip along trends ranging from $N13^{\circ}W$ (1.3 m) to $N7^{\circ}E$ (1.8 m) to $N67^{\circ}E$ (1.9 m). The plunge angles ranged from 9° to 35° . The displacements are variable but, since the Sylmar segment generally trends E-W, the displacements indicate faulting ranging from pure reverse dip-slip ($N13^{\circ}W$ to $N7^{\circ}E$) to nearly pure left-lateral strike slip ($N67^{\circ}E$).

East of the Sylmar segment is the Tujunga segment. The Tujunga segment is about 4 km long (Figure 1). The largest slip on the Tujunga was about 2.1 m and was at the Sylmar end of the segment (Sharp, 1975). Its slip was largely left lateral ($R=2.4$). The slip on this segment varies in amount and direction more than on any other segment (Sharp, 1975). The easternmost, Lakeview

segment, about 4 km long, accommodated roughly equal thrusting and left-lateral slip on the order of 0.8 m (Barrows, 1975).

Horizontal and Vertical Displacements

A peculiar feature of the horizontal displacements in the vicinity of the Sylmar segment is that all horizontal displacement vectors in the northern block, the block north of the San Fernando fault zone, point westward, whereas all horizontal displacement vectors in the southern block point northward. Figure 3 shows horizontal displacements of 12 points in the vicinity of the Sylmar fault segment (data from Yerkes et al., 1974). The displacements are accurate to within about 6 cm. The largest displacements (1.6 m) are in the hanging wall block, just north of the Sylmar surface rupture, and point due west. Within this block the displacements range from WNW to west. The horizontal displacements reduce to much smaller values to the north, west and south. Displacements south of the Sylmar rupture, within the footwall block, point north and are nearly constant at 0.5 m.

Figure 3. Horizontal and vertical displacements near the Sylmar segment (Compiled by Xan Jones Davidson, 1996-00).

The northern and southern blocks are separated by a displacement discontinuity with components of about 0.5 m N-W and 1.6 m E-W. The location of the displacement discontinuity, between the hanging wall and the footwall, coincides with the surface ruptures of the Sylmar fault segment. The same type of discontinuity, though, can be seen at each of the gaps between surface ruptures as well as along a line that extends as far westward as Granada Hills (dashed lines in Figure 1). Presumably the discontinuity marks the location of blind segments of the San Fernando fault zone west of the Sylmar fault segment.

There is a roughly elliptical area of uplift in the hanging wall of the Sylmar and nearby fault segments. Vertical displacements in the vicinity of the Sylmar fault zone are shown via contours in

Figure 3. The vertical displacements shown are only part of a roughly elliptical area of uplift with a major axis of 20 km in the NE-SW direction and 11 km in the NE-SW direction (Morrison, 1973, p. 304) centered approximately at the Olive View Hospital (Figure 1). The vertical displacements were compiled by Xan Jones Davidson from three sources: The City of Los Angeles re-leveled nearly the entire area of the City after the earthquake, and made all the data available for our analysis. These data supplemented data by Savage *et al.* (1975) in the western and central part of the Sylmar segment and by Burford *et al.* (1971) for the area east of the Sylmar segment.

The area of marked vertical uplift extends along the northern side of the Sylmar segment and along part of the Tujunga segment of the San Fernando fault zone. The contours show a bulge north of these fault traces. Along the Sylmar segment the bulge reaches about 1.6 m in height and is markedly asymmetric with the steep limb facing south. There is a sharp increase in uplift on the south flank of the bulge, from 500 m south of the trace of the fault rupture to perhaps 500 m north of the rupture, where the bulge crests. The bulge dies out much more slowly toward the north of the crest, dropping from about 1.6 m at the crest of the bulge to about 0.5 m at the base of the San Gabriel Mountains (at Olive View and Veterans hospitals, Figure 1). Thus the bulge accumulates about 1.4 m of uplift over a horizontal distance of about 750 m on the south flank and dissipates 1.1 m of uplift over a horizontal distance of about 2200 m on the north flank.

The bulge along the Tujunga segment is quite different. It is higher, reaching a peak of perhaps 2.2 m. It is nearly symmetrical. The crest is much farther, about 1500 m, north of the ruptured fault segment. The bulge is somewhat wider, perhaps 3000 m as compared to 2000 m for the Sylmar segment.

Using SeGaLi FaRF to Explain Many of the Observations

Now we will use SeGaLi FaRF to explain many of the observations made at the San Fernando fault zone. There are several distinctive features of the displacement pattern that a satisfactory analysis should be able to explain:

Discontinuity of Horizontal Displacements

There is a marked discontinuity in horizontal displacements across the 2 km long surface rupture (solid line in Figure 2) of the Sylmar fault zone. The displacements in the hanging-wall block (north side) are largely in the strike direction but the displacements in the footwall block (south side) are largely in the dip direction.

There is an obvious dissonance between the displacements of points in the hanging wall and footwall, but we have examined and re-examined the data and have decided that they are probably alright. The theory, it turns out, provides a simple explanation.

In order to address this observation, but to also to train our intuition so that we are able to generalize, we will run a series of examples of end-member dip-slip and strike-slip faults. Our plan is as follows: We will generally use the default values of SeGaLi FaRF until we have information that would suggest better values. Observations suggest that the slip on the fault was on the order of meters. Also, field observations indicate both reverse, dip-slip and left-lateral, strike slip so we select dip and strike-slip values of 1 m. Field observations suggest dips ranging from high to low, so we select dips of 30°, 45°, 60° and 89°. In each solution we select a grid density of 51 by 51, rather than the default values of 31 by 31 in order to improve the resolution of the contours. The fault tip is slightly below (0.01 km) the ground surface.

With these values, we are ready to proceed, using the program as explained in volume 1. Here we move quickly through the various screens to the one of interest.

First let us consider strike slip. The next four screens show contour maps of vertical displacements and vectors of horizontal displacement for faults at the three different dips, plus a nearly vertical fault.

Note: The main point, here, is not to present you with a solution. The idea is for YOU to follow the analysis with your copy of SeGaLi FaRF so that you learn to use the program as a tool. Although it appears as though I jump directly to the solution, in fact, I have run the program, filling in all the intermediate steps. You need to do the same thing.

Figure 4A1. One m left-lateral, strike-slip. Fault dips 30°.

When you reach this screen in each of your runs with SeGaLi FaRF, first you should make sure we got the same result. Compare the figure here with yours. If not, you need to find out why not. Go over the variables carefully and make sure you have entered the same numbers in each screen.

Figure 4A2. One m left-lateral, strike-slip. Fault dips 45°.

Figure 4A3. One m left-lateral, strike-slip. Fault dips 60°.

Figure 4A4. One m left-lateral, strike-slip. Fault dips 89° .

When you are satisfied with your result, you should print out a hard copy. To do so, click on **File**, then **Print Chart**. This will provide you with a permanent record and you can see the results more clearly.

The next four screens show results for faults with the same orientations, but for one meter of reverse, dip-slip:

Figure 4B1. One m reverse, dip-slip. Fault dips 30° .

Figure 4B2. One *m* reverse, dip-slip. Fault dips 45°.

Figure 4B3. One *m* reverse, dip-slip. Fault dips 60°.

Figure 4B4. One *m* vertical (reverse), dip-slip. Fault dips 89° .

We are comparing the patterns of horizontal and vertical displacements in Figure 4. The rectangle in each figure is the projection of the fault plane onto the ground surface. In all cases the faults dip toward the right in the figure. The fault tips are slightly below (0.01 km) the ground surface. The lengths of the arrows and the contour interval are scaled the same, so the interest here is in the qualitative effect of dip of a fault, alone.

Large Displacement Parallel to Strike in Hanging Wall, Small in Footwall

Let us consider the fault dipping at 45° as our reference and then compare the results for more gently and more steeply dipping faults. For pure strike slip (Figure 4A2), there is very little vertical displacement; a little uplift near the upper edge and a little depression near the lower edge in the hangingwall. The horizontal displacements decrease from a maximum amount near the left edge of the fault to smaller values in the hanging to the right of the fault. In the field example, Figure , there is an analogous slight decrease in horizontal displacement in the hanging wall north of the Sylmar segment.

Also shown in Figure 4A2 is that the strike-slip component of displacement in the footwall is much smaller than that in the hanging wall. In the field example, the horizontal displacements appear to be very small also (Figure 3).

We see by comparing Figures 4A1, 2 and 3 that the horizontal displacement pattern changes if the fault dips more or less steeply. For a more gentle dip (30°) the displacements in the footwall are smaller and the displacements in the hanging wall are larger. For a steeper dip (60°) the displacements in the hanging wall and footwall are more nearly equal.

Therefore, we can conclude that an idealized strike-slip fault that can explain the observed displacements along the Sylmar fault segment in San Fernando Valley,

? The magnitudes of horizontal displacements parallel to the strike of the fault are much larger in the hanging wall (1.8 m) than in the footwall (0.5 m).

must have a low dip.

Large Displacement Normal to Fault in FootWall, Small in HangingWall

Now we consider another observation at San Fernando:

- ? The horizontal displacements normal to the strike of the fault are small (about 0.75 m) but generally larger in the footwall than in the hangingwall (about 0.25 m).

Let us examine the results in Figure 4B1, 2 and 3. For pure dip-slip (reverse) on a fault dipping 45° we see in Figure 4B2 that the horizontal displacements normal to the strike of the fault are very small everywhere, but are larger in the footwall than in the hanging wall. In Figure 4B3 we see that the horizontal displacements increase in the footwall and decrease in the hanging wall with increasing dip of the fault. (Note that these changes are opposite to that for strike-slip faulting).

Thus, on the one hand, we can better match the components of slip normal to the Sylmar segment of the San Fernando fault zone if the segment dips somewhat greater than 45° and on the other, we can better match the components of displacement parallel to the Sylmar segment if the segment dips somewhat less than 45° . Clearly, a compromise would be 45° , if we were for some reason compelled to select a single fault blade for the Sylmar segment.

Large Uplift of Hanging Wall, Small Downdrop of Footwall

Now let us turn to the measurements of vertical displacement produced by movement on the San Fernando fault zone (Figure 3). In the vicinity of the Sylmar segment,

- ? The contours of vertical uplift for a reverse, dip-slip fault show significant uplift (up to 2.2 m) in the hanging wall block but essentially no down-dropping in the footwall block.

Examining the results in Figure 4A and 4B, again, we see that the strike-slip would not explain the pattern of uplift because strike slip faulting produces only minor uplift. The reverse-slip must be responsible for the uplift. We see in Figure 4B2 that a reverse fault dipping at about 45° could explain the pattern of high up-lifting of the hanging wall and invisible down-dropping in the footwall. We would note that the down-dropping of the footwall is even smaller for a fault dipping 30° (Figure 4B1). As would be expected, the down-dropping of the footwall gets larger as the fault steepens, as shown for a fault dipping 60° (Figure 4B3). For a vertical fault, the magnitudes of down-dropping and up-lifting are equal.

Asymmetric Uplift Pattern

- ? The contours show an asymmetric uplift pattern along the surface trace of the Sylmar rupture (solid line in Figure 2). The pattern consists of a steep increase in uplift near the rupture, a crest 0.5 to 0.75 km north of the rupture and a gradual decrease in uplift to 0.5 m about 2.5 km north of the rupture.

Figure 5A. Contours of vertical displacement for reverse fault dipping 45° to right.

Figure 5B. Zoomed image.

Let us examine the contours for the problem considered in Figure 4B2, but without the overlay of arrows. Rerunning SeGaLi FaRF again, but with a grid of 101 by 101 points, we get the screen above.

For a reverse fault dipping 45° toward the top of Figure 5 (right-hand side) the gradient in vertical displacements within the hanging wall is very steep near the tip of the fault (left edge). The contours of uplift are so close together there that 13 contours on the right of the fault tip they are shown as a single contour along the fault tip (Figure 5B). The uplift pattern is distinctly asymmetric. The two contours to the left of the fault tip surround a shallow depression.

Thus the asymmetry of the uplift pattern north of the Sylmar fault segment (Figure 2) might be explained in terms of a fault dipping about 45° north.

The effect of the dip can be visualized in Figure 4 for faults dipping 30° or 60° . The asymmetry increases with dip of the fault. If the fault were vertical, the contours of uplift in the hanging wall block of the fault would be reflected exactly in contours of depression in the footwall block. Even at 60° we note that the contours of depression are prominent in the footwall.

The asymmetry decreases with decreasing dip of the fault. For a 30° fault, the gradients on either side of the ridge are more nearly equal than for a 45° fault (Figure 3).

In summary, the patterns of displacement observed at San Fernando are the patterns one would expect for a fault dipping moderately and accommodating both left-lateral strike-slip and reverse dip-slip. The patterns looked strange simply because they were unfamiliar and therefore, nonintuitive. Our intuition has grown a little with this experience.

CHAPTER 2. BROAD RUPTURE ZONES ALONG EARTHQUAKE RUPTURES AT LANDERS

Rupture Belts Rather than Faults

The most spectacular ground rupturing of any U. S. earthquake of the 20th century was produced by the June 28, 1992, M 7.5 earthquake at Landers, California, east of Los Angeles. During this earthquake right-lateral offsets¹⁴ up to about 5 m were accommodated along segments of distinct, en echelon fault zones with a total length of 80 km (Figure 6 and Figure 7).

¹⁴ About 7 m was reported based on hasty observation by geologists trying to get credit for finding the "biggest", but careful observations by Fleming *et al.* (1997) indicate the most common offset being about 4 m and the largest being about 5 m. The same hasty observations by the same people reported zero offset at one place along the Homestead fault zone, but that observation was a result of lack of recognition that faulting can be represented by a broad rupture zone, without large faults.

Figure 6. Landers rupture zone on edge of block that rotated counterclockwise during the 1992 earthquake.

Figure 7. The Landers rupture zone and nearby active faults.

The conditions for fracturing apparently were ideal at Landers, so the earthquake rupturing provided us with an opportunity to re-discover a phenomenon that had been briefly described by Andrew Lawson and G.K. Gilbert following the 1906 San Francisco earthquake, but that has been largely ignored—nay, vigorously denied—since then by the pillars of the earthquake research community. We found that the offsets were accommodated generally not by right-lateral faults—distinct slip surfaces—but rather by right-lateral *shear zones*, tabular bands of localized shearing

(Figure 8). The shear zones were within a *belt of shear zones*, which also accommodated right-lateral shearing (Johnson *et al.*, 1993, 1994; 1996a, 1996b, Lazarte *et al.*, 1994; Li *et al.*, 1994a; 1994b). Along simple stretches of fault zones at Landers the rupture is characterized by a belt within which shear zones and intensification of shearing were telescoped: A broad shear zone of mild shearing contains narrow shear zones of more intense shearing, which, in turn, contain even-narrower shear zones of very intense shearing, which may contain a fault. The ground typically ruptured across broad belts of shearing with clearly-defined, subparallel walls.

Figure 8. Idealized belt of shear zones. (From Johnson *et al.*, 1994)

Bodick Road Rupture Belt

A good example is the Bodick Road belt of shear zones, along the Homestead Valley fault zone (Figure 9 and Figure 10). The Bodick Road belt consists of a broad zone of mild shearing, 180 m wide (other belts range generally from 50 to 500 m.), extending from wall to wall, across the entire width of the belt. The orientation of the belt is northwest-southeast (Figure 15). The broad zone is portrayed by en echelon tension cracks, oriented north-south, reflecting east-west tension generated by overall right-lateral shearing of the belt. In response to right-lateral shearing, the slices of ground bounded by the tension cracks rotated in a clockwise sense, producing left lateral shearing, and the slices were forced against the walls of the shear zone, producing thrusting¹⁵. Within the narrower shear zones, trending roughly northwest-southeast there are tension cracks, thrusts, small folds and other features typical of right-lateral shearing over a width of a few meters¹⁶.

¹⁵ Just like those reported by Johnson and Fleming (1993) along 1989 Loma Prieta coactive faults.

¹⁶ Just like those reported by Fleming and Johnson (1989) in Utah landslides.

Figure 9. Homestead Valley belt of shear zones. Boundaries marked with arrows.

Figure 10. A belt of shear zones, 180 m wide, at Landers. Tension cracks, shown in light gray, formed across the entire width of the belt. There were no fracture in ground on either side of the belt. Fractures within narrow shear zones are shown in black. On right (east) side, rupture zone about 5 m wide accommodates 1 m of right-lateral. On left side, narrower rupture zone accommodates 20 to 40 cm of right-lateral. Remaining 50 cm of right-lateral distributed through the belt. (From Johnson *et al.*, 1994, fig. 5).

The ruptures at Landers had wonderful variety and wide-ranging structural forms, as described elsewhere (Johnson *et al.*, 1996; Fleming and Johnson, 1997).

Here the crucial point is the belt of right-lateral shearing. What we observed at Landers is that strike-slip deformation, large enough to crack and fault the ground, occurred over a broad belt. We did not observe slip on a fault plane or surface. This contradicts our expectation: If ground containing a fault is loaded in right-lateral shearing, and the fault suddenly slips, as in an earthquake, we would expect to see right-lateral slip on the fault, but left-lateral shearing in the country rock containing the fault. The deformation is left-lateral because the ground rebounds toward its former position. This is not at all what we have described at Landers, thus far.

Rupturing at Tortoise Hill

In fact, though, there was left-lateral shearing at a broader scale at Landers. Another important piece of the picture emerges as we examine deformation along the Emerson fault zone segment in the northern part of the Landers rupture (Figure 11). There we had measurements of differential horizontal displacements at several different scales.

In general, the trace of the rupture zone of the Emerson fault is simple and relatively straight for about 2 km to the northwest and 1 km to the southeast of the Single-tower Transmission Line (Figure 12). A belt of shear zones 60-70 m wide characterizes the fault zone. The overall trend of the belt is N45° to 50° W. The belt contains fractures such as individual tension cracks, en-echelon tension cracks, and right-lateral fault elements. The Emerson fault zone accommodated 2.9 m of right-lateral shift at the Single-tower Transmission Line and 3.1 m at Tortoise Hill ridge.

Tortoise Hill ridge is about 1 km southeast of the power-line crossing (Figure 12). Fractures surround the northwest end of the ridge, apparently as a result of splitting of the belt of shear zones into two, separate shear zones. The pattern of the split belt is like the prow of a Tortoise-Hill-ridge boat, or perhaps a canoe, steaming northwesterly. The bounding zone on the northeast side of the ridge continues to the southeast beyond the map area. The fault zone on the southwest side of the ridge extends only to the southeast end of the ridge and stops.

The fault on the northeast side of the belt of shear zones in the Emerson fault zone throughout the mapped area carries most of the differential displacement across the belt. At the power line, 2.7 of 2.9 m of displacement is on the northeast side. At Tortoise Hill, about 2.65 of 3.1 m is concentrated in a narrow shear zone on the northeast side.

Strain Measurements

We have measurements of length- and angle-changes at length scales ranging over three orders of magnitude, from 10 km to 0.1 km. These three levels of information show deformation as a function of position with respect to the surface rupture. Changes in length of long base lines using triangulation and Global Position System (GPS) surveys provide data for far-field strain analysis (Figure 11). Resurvey of a group of pre-earthquake bench marks ranging from about 7 km from the rupture zone to within the rupture zone and across it provides data for close-in deformation determination. Measurement lengths are typically in the range of 500 m to 1000 m. Analytical aerial photogrammetry provides data on change in lengths of braced quadrilaterals where initial line lengths are in the range of 100 m and where measurement points are within and across the shear zones.

Geodetic Measurements

Calculations of changes of line lengths and angles using very long baseline trilateration networks and broad-scale Global Position System (GPS) surveys at Landers (Hudnut and others, 1994; and Freymueller and others, 1994) provide regional information about displacement fields on both sides of the rupture zone. Results of these surveys can be used to compute strains a few kilometers from the belts of surface rupture.

Figure 11. Triangles east and west of the ruptured faults were surveyed by trilateration and show

normalized length changes. The direction and magnitude of maximum extension is indicated with double-headed arrows. The measurements reflect left-lateral shearing on the order of 7×10^{-5} that accompanied elastic, right-lateral unloading of the faults. Right-lateral deformations are concentrated in the vicinity of belts of shear zones along the faults.

THE DEFORMATIONS FOR A TRIANGLE WEST OF THE LANDERS RUPTURE ARE CONSISTENT WITH UNLOADING DURING THE EARTHQUAKE SEQUENCE. FOR THE TRIANGLE EAST OF THE SURFACE RUPTURE (FIGURE 11), CHANGES ARE -0.009×10^{-2}

(compression) for MAUM/CREO, -0.004×10^{-4} (compression) for LEDG/CREO, and $+0.006 \times 10^{-2}$ (extension) for MAUM/LEDG. We note that the MAUM/LEDG leg trends roughly north-south, so the deformations are consistent with left-lateral shearing in the general direction $N45^\circ W$, reflecting unloading of the right-lateral fault zones during the earthquake. If we assume that the deformations are continuous, the direction of maximum extension would be $N3^\circ W$ and the principal extensions would be $E_1 = +0.007 \times 10^{-2}$ and $E_2 = -0.007 \times 10^{-2}$, that is, the deformation is pure shear (or simple shear).

For the triangle west of the fault zone (Figure 11), normalized length changes are -0.7×10^{-4} (compression) for ROCK/BOUL, -0.003×10^{-2} (compression) for ROCK/MEANS, and $+0.006 \times 10^{-2}$ (extension) for BOUL/MEANS. The BOUL/MEANS leg is roughly north-south,

again consistent with left-lateral shearing in the general direction N45°W. If the deformations are continuous, the direction of maximum extension would be N30°E and the principal extensions are $E_1 = + 0.007 \times 10^{-2}$ and $E_2 = - 0.009 \times 10^{-2}$, so there is both left-lateral shearing and slight area decrease.

Generally, the long base-line changes support the expectation of left-lateral shearing that is generally parallel to the direction of the rupture belt. Both sides of the rupture belt deformed in a manner consistent with unloading of the right-lateral fault zones.

Resurvey and Re-level of Net of Benchmarks

Now let us examine results at a larger scale (Figure 12). Resurvey and re-level of a group of pre-earthquake benchmarks ranging from about 7 km from the rupture zone to within the rupture zone and across it provides data for close-in deformation determination. Measurement lengths are typically in the range of 500 m to 1000 m. SoCalEd¹⁷ established benchmarks, surveyed, and mapped an area of about 10 land sections during the mid-1970s as part of the control for photogrammetric surveying of a potential site for a power plant (Figure 12). We hired SoCalEd in 1995 to repeat angle and length measurements and to repeat the leveling survey that were made in 1973 and 1976 as part of our investigation of faulting and deformation in the area of the Landers rupture (Fleming *et al.*, 1996).

The resurveys provided a very clear picture of the uplift of Tortoise Hill tectonic ridge during the 1992 earthquake (Figure 12; Fleming and Johnson, 1997). The horizontal resurvey, though, mainly indicated that the deformations were too small on either side of the belt of deformation zones to be detectable by the survey methods. Through examination of normalized length changes, we infer that length changes of about 3×10^{-4} and larger should be significant but that inferred length changes of smaller magnitude are masked by survey error. Measurements south of Section 25, T6N, R3E, which includes part of Tortoise Hill ridge, reflect such small length changes (or errors) that we are unable to obtain meaningful estimates of deformations (Fleming *et al.*, 1996; Fleming and Johnson, 1997).

¹⁷ Southern California Edison Co.

Figure 12. Contours of vertical displacement, relative to a point (large shaded circle) near Bessemer Mine Road, showing concentrated uplift at Tortoise Hill ridge, within Emerson fault zone. Land survey control points, surveyed in 1973 and 1994, shown with circles. Circles with cross (arrow moving away from observer) indicate downward movement. Circles with dot (and tips of feathers) indicate upward movement (arrow moving toward observer).

According to these results, the normalized length changes are smaller than 2×10^{-4} in an area extending 6.5 km south of Tortoise Hill and about 4.5 km southwest of the Emerson fault zone. The results suggest that normalized length changes and angle changes are below that level everywhere except where there are control points that span ruptured ground. The small normalized length changes are consistent with the regional GPS and trilateration measurements reported in Figure 11, that the normalized length changes are on the order of 5×10^{-5} .

Photogrammetric Measurement of Deformations

In order to determine supplementary length changes in relatively small areas near the rupture zone at Tortoise Hill, we have used a photogrammetric method and sequential aerial photography (see Fleming *et al.*, 1997, for details of method).

Figure 13. Fractures bounding margins of Tortoise Hill ridge and differential displacements measured photogrammetrically. A ladder of quadrilaterals extends across the ridge. At southeast edge, displacements were measured by land survey of regional grid. Maximum horizontal shift across ridge about 2.65 m. Maximum vertical displacement, relative to an assumed fixed point about 6 km south of ridge, is 1.0 m at center of ridge.

We used the photogrammetric method to measure lengths of legs and braces of a ladder of four quadrilaterals extending across Tortoise Hill ridge from the southwest to the northeast sides. The quadrilaterals, in relation to the fractures that we mapped and the topography of the hill are shown in Figure 13. The position and displacement vector of the land survey control point in Tortoise Hill that moved horizontally southward are shown near the right-hand end of Figure 10. We determined normalized length changes by comparing lengths of legs and braces in 1973 and 1992. Normalized length changes smaller in magnitude than about 3×10^{-4} are assumed to be negligible, within the range of the error of the survey (see Fleming *et al.*, 1997).

Starting with the southwest edge of the quadrilateral (Quadrilateral Q2) in the southwest, the normalized length change is marginally significant, but shows a small extension in the northwest direction. The northeast-trending legs cross a rupture zone and both show compression about ten times larger, -1.4 to -2×10^{-3} . The compression certainly reflects the thrusting of Tortoise Hill relatively toward the southwest across the southwest rupture belt. The north-south diagonal brace was shortened and the east-west diagonal brace was lengthened, reflecting the right-lateral shear across the southwest rupture belt. Thus the measurements reflect small to negligible extension parallel to the southwest belt, but significant right-lateral shearing parallel and shortening normal to the southwest belt of rupture that bounds Tortoise Hill ridge.

The next quadrilateral to the northeast (Quad Q0) near the center of Tortoise Hill (Plate 3) indicates very small to negligible deformation (normalized length changes smaller than 3×10^{-4}).

The northeast end of the next quadrilateral (Quadrilateral Q1) to the northeast is within the belt of shear zones on the northeast side of Tortoise Hill ridge. Normalized length changes are generally large, on the order of 10^{-3} and the deformation is clearly inhomogeneous within the quadrilateral. Thus the leg at the southwest edge of the quadrilateral shortened barely significantly, whereas the leg at the northeast edge shortened by -1.5×10^{-3} , apparently reflecting the fact that it traverses the right side of the belt of shear zones obliquely. Because the northwest end of the leg is deeper within the belt than the southwest end, the leg is shortened significantly. The shortening of the north-south brace and the lengthening of the east-west brace again reflect the right-lateral shearing in the belt of shear zones.

Another interesting result is that the southwest and northeast sides of quadrilateral Q1 are both extended significantly. Because of the orientations of these sides relative to the orientation of the belt of shear zones, we would expect minor extension in the southwest side and minor compression in the southeast side if there were only simple shearing. We suggest that the significant extension in both sides reflects movement of the center of Tortoise Hill southward relative to the belt of shear zones on the northeast side of Tortoise Hill. We observed this sense of differential displacement for a control point nearby (Fleming and Johnson, 1997) that moved 0.8 m toward the south.

The last quadrilateral (Quad Q3) extends from northeast of the belt of shear zones into the belt of shear zones. The northeast side of the quadrilateral stretched significantly, 1×10^{-3} ; the reason is unclear. The northwest and southeast sides also stretched large amounts, 1 to 2×10^{-2} , apparently reflecting large right-lateral shearing. The minor northeastward thrusting visible in the belt of shear zones apparently was overwhelmed by the right-lateral shearing. The right-lateral shearing is also reflected in the shortening of the north-south brace and the lengthening in the east-west brace.

Belts of Right-Lateral and Left-Lateral Shearing

In summary, at the three levels of observation, (~10 km, 1 km, 0.1 km lengths), normalized length changes provide insight into the intensity and style of deformation. In the far-field, as measured by trilateration and GPS, principal extensions are in the range of 7 to 9×10^{-5} . On both sides of the rupture belt in the far field, the deformation is *left-lateral shear* that is generally parallel to the belt of surface rupture. Re-survey of pre-earthquake bench marks near Tortoise Hill indicates that the deformation is generally below the limit of survey accuracy (3×10^{-4}) but where results appear to be significant they show left-lateral shear outside the deformation belt and right-lateral shear inside the deformation belt. The braced quadrilaterals spanning Tortoise Hill provided the most detail, and showed that there was not merely right-lateral shear inside the deformation belt, but that the right-lateral shearing was obvious. The right-lateral offsets of one to three meters along the main ruptures were obvious also, of course.

Idealization

Figure 15A shows an idealization of the strain zones detected along the Landers earthquake rupture belts. Within the belts of deformation zones, tension cracks and faults, the deformations all reflect right-lateral shearing parallel to the walls of the belts; there may be dilation normal to the wall as well, but we are not considering that deformation here. Even the tension cracks that transform into left-lateral faults reflect overall right-lateral shearing, as we have described in the 1989 Loma Prieta rupture zone and explained in detail elsewhere (Johnson and Fleming, 1993).

What we observe at Landers is right-lateral slip on one or more faults, a belt of right-lateral deformation in the vicinity of the main fault, and belts of left-lateral deformation on either side, with deformation decreasing with distance from the fault.

Possible Explanation for the Formation of Rupture Belts

What we believe happened at Landers is permanent right-lateral slip on the faults, permanent right-lateral deformation and fracturing within the right-lateral belt, and elastic, left-lateral deformation on either side of the right-lateral belt. The left lateral deformation apparently reflects what we think of as elastic rebound due to stress drop on the earthquake rupture.

We can derive an explanation for the opposite, right lateral and left lateral strain belts side-by-side if we make certain assumptions or accept certain observations as relevant:

1. The strike-slip faults that we have observed thus far in the 1983-86 large landslides in Utah (Fleming and Johnson, 1989), at the 1989 Loma Prieta ruptures (Aydin *et al.*, 1991), at the 1992 Landers (e.g., Johnson *et al.*, 1994; 1996) and the 1999 Duzce Turkey ruptures (in prep.) all reflect mode III propagation of faults (Figure 14), that is, propagation of faults from depth toward the ground surface. At no place have we observed evidence for mode II propagation, that is, propagation along the ground surface. Thus, we assume that the strain belts are a result of mode III loading.

Figure 14. Modes of propagation of fractures.

Figure 15. Idealizations of known strain belts along earthquake ruptures. A. Cartoon of strain belts recognized along Landers 1992 strike-slip fault zones. B. Strain belts along 1971 San Fernando fault zone. C. Strain belts of Winnetka deformation zone (blind fault), 1994 Northridge earthquake. D. Strain belts within Granada Hills deformation zone (blind fault), Northridge.

With this assumption we should mechanically analyze a strike-slip fault whose tip has not reached the ground surface, that is, a blind strike-slip fault.

2. We assume that the faulting is an unloading phenomenon at depth, so that deformation we observe is a result of stress drop on the fault surface and propagation of the fault to the ground surface.
3. We assume that results of a static analysis of strains around a slipping fault are similar to the results of a dynamic analysis of a propagating fault.
4. We assume that a dislocation model adequately represents the kinematics of the deformation around such a fault. That is, we assume that a fault with uniform slip on its surface is an adequate representation of the actual slip on the fault.
5. We assume that the belt of permanent, right-lateral shearing occurs before the fault reaches the ground surface.

With these assumptions, we can turn to SeGaLi FaRF to obtain some possible insights into the formation of the belts.

The result for a strike-slip fault 6 km long and 6 km wide, dipping 89° , and with its tip at depths ranging from 2 km to 0.01 km (10 m) are shown in Figures 12 through 15. The slip is one meter of right-lateral in all cases.

Note: As usual, you need to run these models yourself, using the parameters selected here, and printing out the results. You will learn more if you then modify some of the variables and see what happens to the results.

Figure 16A. Right-lateral, strike-slip fault, with tip at depth of 2 km. Otherwise, the standard fault for this series of runs; its length is 6 km, width is 6 km and dip is 89° . One meter of right-lateral strike-slip. If north is upward in the figure, the maximum extension is oriented northwest-southeast, reflecting the right-lateral shearing throughout the map area. The strains over the tip of the fault are on the order of $10^{-4.5}$ (between 10^{-5} and 10^{-4}).

Figure 16B. Fault tip at depth of 2 km. Zoomed version of 12A, showing that strain figures reflect right-lateral shearing out to the right and left edges of the figure.

Figure 17A. Right-lateral strike-slip fault, with tip at 0.5 km. Other conditions same as in Figure 16A. The strains over the fault are right-lateral, and there is a belt, about 1 km wide, where strains are on the order of $10^{-3.5}$. Elsewhere the strains are much smaller. Within a belt about 2 km wide, near midlength of the fault, the strains are generally right-lateral, but farther away they become left-lateral.

Figure 17B. Closer view of Figure 17A.

Figure 18A. Right-lateral strike slip fault with tip at depth of 0.1 km. Otherwise, same conditions as in Figure 16A and Figure 17A. Strains of elements straddling the trace of the fault (which is at depth of 100 m) are about 10^{-3} . The width of the belt is on the order of 200-300 m.

Figure 18B. Zoomed view of Figure 18A.

Figure 19. Right-lateral, strike-slip fault with tip at depth of 0.01 km. Otherwise, same conditions as in Figure 16A, Figure 17A and Figure 18A. View is zoomed. Strains of elements straddling the trace of the fault (which is at depth of 10 m) are about $10^{-2.5}$. The width of the belt or right-lateral shearing cannot be resolved, but is narrow. The strains on either side of the belt are left-lateral and on the order of 10^{-4} and smaller.

The story is largely told in the captions of the figures. We imagine that the fault is propagating toward the ground surface. When the tip of the fault is at a depth of 2 km (Figure 16), the strain figures throughout the diagram reflect northwest-southeast extension (if north is taken to be up in

the figures) and roughly equal compression at right angles, so they reflect right-lateral shearing. The magnitudes of strains over the tip of the fault are about $10^{-4.5}$. When the fault is 0.5 km beneath the ground surface (Figure 17), there is a belt, about 1 km wide, where right-lateral strains are on the order of $10^{-3.5}$. Elsewhere the strains are much smaller. Throughout a belt about 2 km wide the strains remain right-lateral, but they reduce in size with distance. Farther away the strains diminish and switch to left-lateral. Therefore, when the fault is at 0.5 km depth one can see the outer regions of elastic rebound, with left-lateral shearing, and the inner region of right-lateral shearing. Within the inner region is a narrow belt centered on the fault with rather large right-lateral shearing.

Amount of Strain Required for Fracturing

The right-lateral strains in the inner belt may be large enough to cause fractures in brittle rock and soil. If so, the strain pattern described above could explain the formation of a belt of ground damaged in right-lateral shearing. In this respect, Cruikshank *et al.* (1996) discuss the strains at which fractures form in several materials. We document there that a brittle rock, such as granite, fails under atmospheric confining pressures at compressive axial strain values of between -2 to -6×10^{-3} . In the Aspen Grove landslide in Utah, the soil in a developing landslide toe cracked at strains of about -1.4 to $-2?10^{-2}$ and in a newly forming pull-apart, the strains were about $6?10^{-3}$ before a tension crack formed and $3.5?10^{-2}$ after the tension crack had formed. In the Granada Hills area of the 1994 Northridge earthquake, the magnitudes of strains in regions where sidewalks and streets were cracked and foundations were damaged were commonly $3?10^{-3}$ and ranged up to 10^{-2} . In the Winnetka area of the Northridge earthquake, the strains were about an order of magnitude smaller, commonly $3?10^{-4}$, where there was no visible damage to streets and sidewalks, and ranged up to about 10^{-3} where there was minor to moderate damage. At a strain of $10^{-3.5}$, therefore, as in the example shown in Figure 17, we would expect only minor cracking, even in brittle materials.

We should note, though, that we have considered fault slip of one meter. Suppose the slip was three meters (at Landers the shift across fault rupture zones at the ground surface was typically about 4 m). Then, repeating the run that produced Figure 17, we obtain the result shown in Figure 20, that the maximum strain is about $10^{-2.5}$, which is definitely large enough to cause damage.

Figure 20. Same conditions as in Figure 17A except amount of slip on fault was tripled to three meters; tip of fault is within 0.5 km of the ground surface. View is zoomed. Strains of elements straddling the trace of the fault (which is at depth of 10 m) are about $10^{-2.5}$. The strains on either side of the belt are left-lateral and on the order of 10^{-4} and smaller.

A Scenario for Formation of Rupture Belts at Landers

On this basis, we might suggest that the rupture belts that we saw at Landers could be cracking and faulting of very brittle, compact alluvium as a result of high shear strains developed out ahead of a fault propagating from depth (mode III propagation). The rupture belts presumably started to form when the fault tip was on the order of 0.5 km beneath the ground surface. As the fault propagated closer to the ground surface, presumably, the right-lateral strains were higher, and occurred over a narrower zone. On either side of this narrow zone the ground would have been subjected to left-lateral shearing (rebound). Presumably there would be further fracturing within the narrower zone. With further propagation the zone of right-lateral shearing narrowed and the right-lateral strains intensified. At Bodick Road (Figure 10), the narrow shear zones and small faults in the east (right) half of the belt of shear zones may have formed at this time. The fault eventually approached the ground surface, the zone of right-lateral shearing was narrow and intense, and the rupture that we termed the "main rupture or main shear zone" appeared. At Bodick Road (Figure 10), there is a zone of intense fracturing and deformation about five m wide, along the east flank of the belt of shear zones may have formed at this time. About one m of shift was accommodated across this five m wide zone. Near the edge of this zone is presumably "the fault." Over a horizontal distance of about one m, 0.6 m of shift was accommodated by "the fault."

We documented a sequence of events something like that in the scenario within the Happy Trail shear zone, which is part of the ruptured Johnson Valley fault zone (Johnson *et al.*, 1994). Through logic and via cross-cutting relations of different kinds of fractures in the Happy Trail shear zone we showed that a narrow, right-lateral fault was the last rupture to break through the Happy Trail shear zone. Now we have a plausible theoretical basis for understanding the sequence of events.

This example very nicely illustrates the value of SeGaLi FaRF as a tool for analysis of fault-related structures. Only that we had it earlier!

CHAPTER 3. FRACTURING WITHIN THE HEADQUARTERS RIGHT STEP AT LANDERS

Setting of Headquarters Duplex Structure

The Headquarters duplex, along the northeast side of the rupture zone of the Homestead Valley fault zone (Figure 21), is one of the more fascinating structures we saw along fault ruptures at Landers. The duplex is crossed by Bodick Road and Shawnee Road is immediately east of it (Figure 21). The duplex consists of a right-stepping right-lateral shear zone on the east side of the Bodick Road belt of shear zones, connected by many smaller, diagonal right-lateral shear zones and faults (Figure 22).

Figure 21. Rupture zones along Homestead fault zone. Headquarters duplex structure is at Bodick Road.

intervening steps between the segments. At the Headquarters step, there is a duplex structure; at the Pipes Wash step, farther south, there is a left-lateral rupture zone bounding a rotating block. We are going to discuss the duplex structure at the Headquarters step.

The Duplex Structure

The duplex begins near Mikiska Boulevard, north of Bodick Road, and extends about 400 m south to Mileska Ranch, on Shawnee Road (Figure 22). Bob Fleming, Ken Cruikshank and I mapped a small part of the duplex north of Bodick Road about one month after the earthquake, when it was perfectly preserved. Bob Fleming mapped the part of the duplex structure south of Bodick Road a year after the earthquake, yet its essential features were still well defined (see Figure 23). The features south of Bodick Road are nearly as clear as those north of it.

The duplex structure is about 400 m long and 90 m wide. It consists of the bounding faults plus diagonal, shear-zone segments that form the internal elements of the duplex structure. The bounding fault on the east extends southward, to near the building marked "Headquarters" in Figure 23, about half the length of the duplex. The bounding fault on the west extends northward to near Bodick Road, about two-thirds the length of the duplex.

Offsets of fences with different orientations at the North Ranch indicate that the bounding fault on the east side of the duplex accommodated 1 to 1.2 m of right lateral shift (Johnson and others, 1993). Offset of fences with different orientations at Mileska Ranch indicates a net shift of 1.6 m across the bounding fault segment on the west side of the duplex. Besides the lateral shift, the ground to the west of the west fence of the Mileska Ranch was relatively downdropped about 10 cm.

Left-Stepping Shear Zones Within Duplex

The shear zones within the duplex are right lateral, as are the bounding faults. Individual shear zones are spaced about 10 m apart, are 100 to 150 m long, and are oriented at a clockwise angle of 20 to 30° to the bounding faults. The shear zones within the duplex are, themselves, short, right-lateral, *en echelon* fault elements 5 to 10 m long. Between many of the fault elements are short thrusts (Figure 23) that shift displacement across from one fault element to another.

Thus the Headquarters step consists of a hierarchy of stepping elements (Figure 23). The Headquarters duplex itself forms a releasing step (Figure 22) because the bounding faults are right lateral and they step right. Within the duplex, though, the fault elements have restraining steps, at two different scales. The shear-zone elements within the duplex have restraining steps because they are right lateral and step left. Furthermore, many of the shear zone segments are composed of short fault elements also arranged as restraining steps.

The lateral shift in the internal shear zones in the northwest part of the duplex, between Bodick Road and Mikiska Boulevard, was evident in the deformation of the fence on the north side of the Northern Ranch (Figure 10 and Figure 23). The lateral shift there was generally 5 to 10 cm. In the southeast part of the duplex there were few markers, but we would estimate an average of 30 cm accommodated by each short shear-zone segment, about 5 to 10 m long. We could measure vertical shift, generally a few cm, with downthrowing on either side. The openings of cracks along the internal shear zones were similarly on the order of a few centimeters, indicating some dilation in the releasing duplex.

Analysis of Conditions of Formation of the Internal Shear Zones

Perhaps we can understand some of these features (certainly not all of them; this is a very complex structure) in terms of the dislocation/fault model. We suppose that the ruptures were propagating upwards, in mode III.

Figure 24A. Strains at ground surface for a 3km by 3 km, nearly vertical, right-lateral, strike-slip fault. The upper edge of the fault is at a depth of 2 km, so the strained ground is far above the fault tip.

Left-Stepping and Right-Stepping Fault Segments

Figure 24B. Three stepping faults subjected to the same conditions as in Figure 24A. The faults are propagating upward, loading the overburden material in mode III. Grid point 31x31 used to compute strains.

We will examine the strains at the ground surface as stepping faults propagate from a depth of 2 km to near the ground surface. Figure 24A shows a single fault at a depth of 2 km. There is a belt about 3 km wide of right-lateral shearing on either side of the projection of the fault to the ground surface. Figure 24B shows three faults at a depth of 2 km. One pair of faults (upper) is right-stepping (releasing) and the other (lower) pair is left-stepping (restraining) so we can compare and contrast strain patterns. All have accommodated 1 m of right-lateral. They are essentially vertical (89°). The strain pattern is quite similar to that of a single fault. The zone of right-lateral shearing is perhaps a little wider, but otherwise the three appear to behave essentially as one longer fault.

Figure 25 Depth of 0.5 km (500 m). Zoomed image. Grid points 71x71.

Figure 26. Depth of 0.01 km (10 m). Zoomed image. Grid points 71x71.

As the faults propagate to within 0.5 km of the ground surface we see the strain pattern shown in Figure 25. We plot 71 points (rather than the 31 used in Figure 24) and zoom the image in order to see more details.

Comparing Figures 24B, 25 and 26, we see that the strain patterns between the tips of the faults change markedly as the faults approach the ground surface. For the right-step (upper step in the figures), the maximum extension ranges from about 45° to the trend of the faults for a depth of 2 km, to about 90° to the trend of the faults at a depth of 0.01 km. Thus tension cracks would tend to form at about 45° to the trend of the underlying fault when the fault is deep, to nearly at right angles to the trend of the underlying fault as the fault propagates to the ground surface.

In general we do not observe tension cracks within the area of the duplex structure, except those directly associated with internal shear zones. However, there is a peculiar cluster of tension cracks, running roughly E-W, just west of the duplex structure. These tension cracks have been enigmatic. Perhaps they are a result of near-surface interaction of faults in the belt of shear zones.

The strain diagrams indicate that right-lateral faults, which are the kinds of faults we observed in the duplex structure, would tend to be oriented roughly parallel to the underlying faults at a depth of 2 km, to perhaps an acute angle of 10 or 15° to the underlying faults as the faults reach a depth of about 0.5 km, to an angle of about 45° to the underlying faults as the faults approach to within a few meters of the ground surface (0.01 km).

Re-examining the map in Figure 23, we note that the duplex faults (or narrow shear zones) within the duplex structure trend at low angles to the right-stepping fault segments that parallel the walls of the belt of shear zones at Bodick Road. The angles range from a few degrees to perhaps 30° . According to the analysis, these duplex ruptures would have formed when the right-stepping fault segments were within 2 km of the ground surface, probably mostly when the right-stepping segments were within, say, 0.5 to 0.1 km of the ground surface.

CHAPTER 4. WINNETKA DEFORMATION ZONE, 1994 NORTHRIDGE, CALIFORNIA, EARTHQUAKE

As a final example of the use of the strain patterns generated by SeGaLi FaRF, we will use the program to analyze field measurements that uncovered distinctive strain belts in Winnetka and Northridge during the 1994 Northridge, California, earthquake. Blind faulting in the area was first suspected by Kenneth Cruikshank, who visited the Northridge area immediately after the earthquake (Cruikshank *et al.*, 1996), looking for evidence of blind faulting that other field investigators overlooked or misunderstood (e.g., Hecker *et al.*, 1996; Holzer *et al.*, 1996). We contracted the surveyors of the City of Los Angeles to resurvey the streets in the Winnetka and Northridge areas, and it was study of those resurvey results that led to recognition of the strain belts.

Figure 27. Contours of vertical uplift and location of Winnetka deformation zone, San Fernando Valley, California. Location of San Fernando Valley, between Santa Susana Mountains to north and Santa Monica Mountains to south. Epicenter of main shock at Northridge, in center of valley. Contours based on leveling data show differential vertical displacement between 1980 and 1994, and, in general, fan-shaped tilting of San Fernando Valley and Santa Susana Mountains relative to an assumed fixed point in southwest part of area. The Winnetka and Granada Hills deformation zones are shown with gray outlines.

Winnetka Deformation Zone

The Winnetka deformation zone is perhaps as long as 9-km, extending from Canoga Park in the southwest to Northridge on the northeast (Figure 27). We initially suspected a deformation zone for three in the horizontal plane in the northeastern part of the area (Figure 30) and they measured lengths of streets so that we could calculate normalized length changes for a larger area (Figures 18 and 19).

Figure 28. Normalized length changes of streets in part of the Winnetka deformation zone.

Figure 29. Normalized length changes of streets in Winnetka deformation zone (between 1970's and 1995). Magnitude indicated by radius of heavy circle. Where magnitude is smaller than 10⁻⁴, measurement indicated with circle with diagonal line. Deformation zone, about 1 km wide and 4.5 km long, trends northeast-southwest. Band of extension in northwest part (+ sign) and compression (- sign) in southeast part. The deformation zone may continue to southwest into Canoga Park, into area with many "red tagged" buildings.

reasons: The pattern of uniform tilting of the valley is disturbed near Northridge. There appeared to be more damage to streets and sidewalks southwest of Northridge than in surrounding areas. The

entire zone appeared to contain more "red-tagged" buildings than surrounding areas. In order to explore the limits of the Winnetka deformation zone we contracted with the City of Los Angeles to resurvey subsurface monuments for some streets. They had previous surveys of subsurface monuments, generally made in the 1970's or 1980's, of the entire area. The City collected complete data on angles between streets and lengths of streets so that we could calculate principal strains

Normalized Length Changes

The normalized length changes of street segments, computed with street lengths measured by the City in the 1970's and then in 1995, are shown for the Winnetka area in Figures 28 and 29. Streets extended are indicated with darker circles and streets shortened with lighter circles. The magnitude of the normalized length change can be determined by comparing the size of a circle with the size of scaled circles shown in the Explanation. An arrow in the circle indicates the direction of the street segment measured. A circle with a diagonal line indicates that a normalized length change is less than the threshold of significance, 10^{-4} .

Figure 28 shows that: 1). Trending diagonally through the Canoga Park, Winnetka and Northridge area is a deformation zone within which normalized length changes are larger and outside of which they are smaller than the threshold value. The deformation zone is about one kilometer wide and at

least four kilometers long. 2). The deformation zone contains an extension belt and a compression belt. The extension belt is in the northwest part of the deformation zone and is about 0.6 km wide and 4 or 4.5 km long. The compression belt is in the southeast part and is about 0.4 km wide and 4 km long. 3). The magnitudes of maximum extension or compression are generally about $\pm 3 \times 10^{-4}$ and decrease from northeast to southwest. The magnitudes are locally 10^{-3} . 4). The larger strains in

the northeast correlate with greater damage to structures such as roads, sidewalks and houses (Figure 27).

Elevated damage to structures in the Conoga Park area suggests that the Northridge deformation zone extends beyond the survey area to the southeast. Similarly, the high strains at the Northridge end of the zone suggests that the zone extends into the area where large commercial structures and the campus of California State University, Northridge, were damaged. The total length of the Winnetka deformation zone could be 9 km, as indicated tentatively in Figure 26.

Figure 30. Strains in northeastern part of Winnetka Deformation zone. Strain figures represent changes in lengths and angles between street segments at each intersection. Magnitude of largest extension (regardless of sign) indicated by radius of heavy circle. Magnitude is determined by comparing the size of a circle with the scale of circles. The lighter line represents the directional distribution of extension (where outside the circle) and compression (where inside circle). The strain figures (shmoos and nerds) show that the belt of extension (+ sign) in the Winnetka deformation zone is dominated by principal extensions in the northwest-southeast direction, and that the belt of compression (- sign) is dominated by a principal compression in the same direction. This pattern indicates that the Winnetka deformation zone would not represent a strike-slip fault, but most likely a reverse fault. (From Cruikshank et al., 1996).

Strains

In the northwestern end of the Winnetka deformation zone (Fig. 27) the City measured both street lengths and angles in 1995, so we were able to compute principal strains at intersections (Fig. 30) and investigate the internal structure of part of the deformation zone. The strains are shown via strain figures¹⁸ that are read as follows: The size of the circle is compared to the size of a series of scaled circles to determine the strain magnitude. The rest of the strain information is contained in the multi-curved line plotted on top of the circle. The multi-curve indicates the distribution of the strains and shows the principal strains. If the multi-curve is on the circle the strain is zero. Where the multi-curve is inside the circle there is compression, and where it is outside the circle there is extension.

The strain figures show that the maximum extension in the extension belt is parallel to the maximum compression in the compression belt of the Winnetka deformation zone. Both trend northwest-southeast, normal to the walls of the deformation zone (Fig. 27). Thus the directions of principal strains are parallel and normal to the walls of the deformation zone, indicating that the zone is not a strike-slip zone, but, rather, is a reverse (or normal) zone. The compressional setting of the San Fernando Valley suggests that the Winnetka shear zone is a surface manifestation of a coactive blind reverse fault dipping northward.

In our paper (Cruikshank *et al.*, 1996) we made some rough calculations of the magnitude of the strain generated by a blind fault, using the solution for a crack in plane strain. We estimated that the slip at Winnetka was on the order of 2 m for a fault with a width of 2 km and calculated that the strain near the ground surface would be on the order of 3×10^{-4} if the tip of the blind fault were a distance equal to the half the width of the fault, or 0.5 km (500 m), and it would be an order of magnitude larger, or about 3×10^{-3} , if the tip of the fault were a distance of 0.05 km (50 m) from

¹⁸ These "Smoos" and "Nerds" were used to show strains before the squares and rectangles, now used in the SeGaLi FaRF program, were introduced (See Johnson *et al.*, 1996; Cruikshank *et al.*, 1997).

the ground surface or would be about 10^{-2} if the tip of the fault were a distance of 0.005 km (5 m) from the ground surface.

Analysis of Deformation Zone

Let us see what SeGaLi FaRF gives for these conditions. Assuming a fault length of 2 km, a width of 2 km, a reverse dip-slip of 1 m, and a dip of 45° , we obtain the results in Figure Figure 31 for a depth of 0.5 km to the fault tip. The maximum strain is on the order of $10^{-3.5}$. There is a compression belt in front (to left of tip of fault) of the fault and an extension belt over the fault. Thus the belts are as we expected them to be, based on the field measurements. The strains, also, are of comparable magnitudes.

Although the strains are larger for the fault tip at a depth of 50 m, as the rough calculations indicated, the strain belts are quite different. The belt of high compression is much narrower, perhaps 200 m. The largest strain there is about -10^{-3} . The belt of extension is still rather large, about 500 m wide, but the extensile strains remain small, about $10^{-3.5}$.

Figure 31A. Fault tip 0.5 km below ground surface. Strain figures for reverse fault dipping 45° to right. 2 km long, 2 km wide. Slip 1 meter. Belt of extension over the fault. Belt of compression in front of fault, just as in the field example. Each belt, defined by strains at least 10^{-4} , about 0.5 km wide. The correlation between the field and theoretical results is striking.

Probable Depth to Tip of Blind Winnetka Reverse Fault

Thus the results of SeGaLi FaRF confirm our rough calculations, that the strains measured with our surveys at Winnetka can be explained by a relatively small fault, perhaps 2 km by 2 km in size, slipped one meter during the 1994 earthquake. In the field example there were peculiar belts of opposite strain, situated side-by-side to form the Winnetka deformation belt. The widths of these belts and the magnitudes of the strains can be explained in terms of a blind fault with dimensions of 2 km width and length with its upper end at a depth of 0.5 km that accommodated about 1 m of reverse dip-slip.

Figure 31B. Zoomed view of map for depth of 0.5 km.

Figure 32A. Fault tip 0.05 km (50 m) below ground surface. Strain figures for reverse fault dipping 45° to right. 2 km long, 2 km wide. Slip 1 meter. The pattern is quite different from that in Figure 31A. There is still a belt of extension over the fault and a belt of compression adjacent to it, but the belt of compression is centered on the tip of the fault, and it is much narrower, perhaps 200 m wide. The maximum compressive strain is larger, on the order of 10^{-3} (10^{-3} to $10^{-2.5}$). The maximum extension, though, is smaller. The correlation between the field and theoretical results is much better for the example in Figure 31A.

Figure 32B. Zoomed view of map for depth of 0.05 km.

AN EMENDATION OF ELASTIC REBOUND THEORY: MAIN RUPTURE AND ADJACENT BELT OF RIGHT-LATERAL DISTORTION DETECTED BY VIADUCT AT KAYNASLI, TURKEY 12 NOVEMBER 1999 DÜZCE EARTHQUAKE

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ABSTRACT

The fault trace of the 12 November 1999 earthquake in the Düzce-Bolu region in Anatolia crossed the alignment of a 2.4 km viaduct in Kaynaşlı that had been carefully surveyed. The builders of the viaduct, the ASTALDI-BAYINDIR Co., resurveyed the viaduct after the earthquake. We repeated the survey for approximately one kilometre of the eastern end of the viaduct and obtained essentially identical results. Though it was unfortunate that the earthquake damaged the new structure, the piers did produce a very rare record of ground deformation of an earthquake. In effect, the viaduct was a giant strain gage that yielded reliable data about ground movement and distortion near a fault. This paper describes the survey data and their evaluation leading to convincing evidence that (a) the fault trace must be considered, not as a fault line or plane, but as a fault zone with a finite width and that (b) the structural damage within the zone was caused,

not primarily by ground acceleration, but by ground distortion. Along the right-lateral fault at Kaynaşlı, the fault zone consists of right-lateral movement at the main trace, a zone of right-lateral distortion near the trace, bounded by left-lateral distortion. The 12 November 1999 event in Turkey, like the ground deformation and fracturing at Landers, California (Johnson et al., 1994, 1996), thus affirmed a forgotten conclusion from the studies by Lawson (1908), Gilbert and Reid (1910) of the 1906 San Francisco earthquake that earthquake ruptures typically occur throughout zones or belts, rather than along linear traces or planes.

INTRODUCTION

Kaynaşlı Viaduct I (Figure 1) is a double-ribbon bridge, 2.4 km long, with piers approximately 45 m tall. It is part of the E-W Trans-European highway under construction between Istanbul and Ankara. The viaduct trends roughly east-west, lifted above the Asarsu Valley by two sets of piers, one for eastbound and the other for westbound lanes. There are 57 pairs of piers plus an extra one under the eastbound lanes. A typical pier is number 38, shown in Figure 2. The base of the pier is the *pile cap*, a 3-m thick slab of concrete 16 by 18.7 m in plan. The pile cap rests on piles that extend to bedrock beneath the gravel of the Asarsu Valley. Standing on the pile cap is the *shaft* of the pier, 45.6 m long, 8 m wide and 3.7 m broad. The *pier cap*, resting on the pier, is 5 m by 17.25 m in plan, supports segments of the roadway.

The points in Figure 2 are the survey points of ASTALDI-BAYINDIR Co. The points provide networks of points that were surveyed after construction. During the Düzce earthquake the elements of the roadway and, presumably the piers themselves, moved in various directions. Their positions were accurately resurveyed so that the job of repairing and completing the roadway could be planned. The Purdue Task Force was provided with complete sets of the survey data through the good offices of Dr. Çetin Yılmaz of Middle East Technical University. We have used these networks in order to compute strains and deformations in the ground beneath the piers. Our own resurvey was of points a couple of metres below points corresponding to 7 and 19 in Figure 2 on piers under the eastern end of the viaduct. Thus our survey provides a network of points for part of the viaduct.

The bases of the piers moved due to offset of the main break of the November rupture. Figure 3 shows a photograph of the main break beneath the viaduct, taken shortly after the earthquake. The rupture, perhaps a metre wide, is trending eastward toward pier 47L, under the westbound lane. The tension gashes, oriented perhaps 45 degrees clockwise from the trace of the main rupture indicate right-lateral shearing. According to Barka and Altunel (2000), the offset of piers was right-lateral and on the order of 1.15 m. Figure 4 shows the rupture zone approaching pier no. 45R, under the eastbound lane, as viewed eastward. The main rupture steps left about a metre in the central view, and thrusting occurred at the step, again reflecting right-lateral offset across the main rupture. The rupture threaded its way between piers 46 R and 46 L. The pier caps at the tops of both piers 45 R and 47 L were visibly misaligned relative to the roadway, as though they had rotated a few degrees in a clockwise sense (as viewed from the ground), which is consistent with right-lateral, simple shearing. Figure 5 is a map of the main rupture and the mounds (thrusts) and depressions (pull-aparts) around the footings of some of the piers in the vicinity of piers 44 through 54 of the viaduct.

The ground deformation was not restricted to the ruptures shown in Figures 3, 4 and 5. We shall show that the spacing of piers changed many metres on either side of piers 45 and 47. The earthquake rupture intersected Viaduct I at a low angle so a long stretch of the viaduct was disturbed by the main rupture and deformation in ground on either side of the main rupture.

Figure 1. View west-northwest of Kaynaşlı Viaduct I. Arrows to left mark the main rupture. Arrow to right is approximate location of main rupture at viaduct.

Figure 2. Survey stations on pier 38. Pile cap at base. Shaft of pier above. Pier cap at top.

Figure 3. View east southeast of main rupture and some of piers of Kaynaşlı Viaduct I. Main rupture is headed toward Pier 47 L at left (photo courtesy of P. Gulkan).

Figure 4. View southeast along piers beneath Kaynaşlı Viaduct I. Dirt service road, on the right, is on the south side of viaduct. Rupture zone in foreground cuts across road and runs into pier 45R, under the eastbound lane. Left step in main trace in middle distance produces thrust along rupture. (Photo Courtesy of P. Gulkan)

BACKGROUND OF THIS STUDY

Rebound Theory vs. Observation

To explain why one would be interested in such measurements, we need to point out some inconsistencies between the elementary notions of elastic deformation before and elastic rebound during an earthquake on the one hand and the field observations made along actual earthquake ruptures on the other. We go back to the ideas of the elastic rebound theory of earthquakes and the field observations along the 1906 San Francisco earthquake rupture discussed by G. K. Gilbert (1907) and Harry Reid (1910). In these early papers, based partly on study of the 1906 San Francisco earthquake, an earthquake is considered a result of sudden slip on a fault in ground that is under high enough stress for the fault surface to fail. Thus, the earthquake is considered a result of a stress drop at a fault. Reid (1910) explained that markers at the ground surface would be offset in different ways, as a result of strike-slip earthquake rupture that broke through to the ground surface, depending on when the markers were placed. The only case of interest here is the marker placed shortly before earthquake rupture. Thus, let the dashed line in Figure 6A represent the map view of the trace of a strike slip fault and line *ab* represent a passive marker placed across the fault before earthquake rupture (Figure 6A). The ground is highly stressed before the slip. If the fault slips suddenly (traction on fault suddenly drops), the ground will shake due to radiated energy, and the ground on either side of the fault will deform elastically. The passive marker will appear as two deformed line segments, *ac* and *db*, broken by the fault (Figure 6B). Note that the offset on the fault is *right-lateral*, but the deformed line indicates that the rock on either side of the fault was distorted in a *left-lateral* sense. This is, of course, due to the *elastic rebound* according to the notion of earthquake generation. The left-lateral distortion is indicated via the exaggerated deformation of squares into parallelograms on either side of the fault (Figure 6B).

The phenomenon described is the heart of the elastic rebound theory of earthquake generation on a fault of finite length (The decrease in displacement away from the fault is a result of the finite length of the fault).

A corollary of the elastic rebound theory has been that there are three causes of earthquake damage, all described in the 1906 earthquake report (Lawson, 1907):

- ? Shaking of structures in excess of their design capabilities.
- ? Collapse of ground beneath structures due to landsliding or liquefaction of soil generated by shaking.
- ? Direct offset along a main fault rupture.

During the 80 years between the 1906 San Francisco earthquake and the 1989 Loma Prieta earthquake in California earthquake damage has been explained in terms of one of these three phenomena.

The far-field, left-lateral distortion that we associate with elastic rebound along a right-lateral fault, of course, has been thoroughly documented in the literature, new and old. This is the deformation that is shown by regional GPS and geodetic data for an entire fault rupture. For example, strains were calculated by Fleming and Johnson (1997) from GPS measurements of displacements in a network of points with spacings of 20 to 70 km reported by Hudnut *et al.* (1994) for the north end of the Landers earthquake rupture. The principal strains were 7×10^{-5} and -7×10^{-5} on the east side of the Landers rupture, confirming left-lateral distortion in the NW direction that one would expect from elastic rebound along a right-lateral fault.

Figure 5. Sketch map of main rupture beneath Kaynaşlı Viaduct I. Rectangles represent pier caps. (After figure prepared by ASTALDI-BAYINDIR Co)

A. Passive marker across impending rupture

B. Elastic rebound across ruptured fault.

C. Main rupture centered in right-lateral shear belt. Left-lateral elastic-rebound on either side of belt.

D. Main rupture on side of right-lateral shear belt. Left-lateral elastic-rebound on either side of belt.

Figure 6. Idealized ground deformation along earthquake ruptures. A. Passive marker placed across fault just prior to rupture. B. Offset of passive marker along fault and deformation in ground according to classic elastic-rebound theory. C. Belt of permanent, right-lateral deformation containing main rupture. Elastic, left-lateral deformation (elastic rebound) on either side. D. Right-lateral belt of shearing on one side of main rupture. Left-lateral deformation on either side

Figure 6 (Alternate) Shows features in color.

Right-lateral Rupture Zones

Observations made at the ground surface closer to the main rupture at Landers, though, are at odds with the simple elastic rebound theory (Johnson *et al.*, 1993; 1996a; 1996b). The most obvious feature that deviates from the simple rebound theory is that there may be a complex rupture belt at the ground surface, not merely a fault trace. In the case of right lateral, strike-slip faulting¹⁹ at Landers, there was a zone or belt of right-lateral shift 50 to 500 m wide along a right-lateral earthquake rupture, not a simple fault plane of right-lateral slip (Figures 6C and 6D) (Lawson, 1908; Johnson *et al.*, 1994). Instead of a fault plane, there may be a main rupture, which may be a metre or more wide, along which much of the right-lateral shift occurs, on either side (Figure 6D) or perhaps within the belt of right-lateral rupturing (Figure 6C) (Johnson *et al.*, 1994).

It turns out that Johnson *et al.* (1994) re-discovered belts of rupturing that had been described by Lawson, Gilbert and Reid after the 1906, San Francisco earthquake, at least along the northern stretches of the rupture (Lawson, 1908; Reid, 1910). The following quote describes a broad shear zone, about 100 m wide, as well as a narrow shear zone with long, echelon fractures at an acute clockwise angle to the walls of the shear zone. The zone of most intense rupture, "the fault-trace or rupture plane" occurred on one side or the other of a shear zone:

"... The surface of the ground was torn and heaved in furrow-like ridges. Where the surface consisted of grass sward, this was usually found to be traversed by a network of rupture lines diagonal in their orientation to the general trend of the fault... The width of the zone of surface rupturing varied usually from a [metre] up to [15 m] or more. Not uncommonly there were auxiliary cracks either branching from the main fault-trace obliquely for [30 to 100 m], or lying sub parallel to it and not... directly connected to it. Where these auxiliary cracks were features of the fault-trace, the zone of surface disturbance, which included them, generally had a width of [about 100 m]. *The displacement appears thus not always to have been confined to a single line of rupture, but to have been distributed over a zone of varying width*" (Lawson, 1908, p. 53; italics ours).

Thus, we have re-established that belts of fracturing and distortion can form along an earthquake rupture, so the picture of simple elastic rebound needs to be emended, at least near the ground surface.

Composite Idealized Model

Although at Landers there was a belt of right-lateral deformation, made visible by ground fracturing, as well as far-field left-lateral deformation, shown with GPS stations spaced 10 to 20 km apart, no information could be obtained about ground deformation in the ground between these quite different types of deformation. At Tortoise Hill, Landers, a double ladder array of control points, spaced 500 m to 1000 m apart, extended from a distance of about 4 km SW of the fault to the fault and a short distance NE of the fault (Fleming and Johnson, 1997). A re-survey of the control points showed that the horizontal strains are about 10^{-4} , which is smaller than the combined surveying errors, about 3×10^{-4} . The strains in ground outside the right-lateral rupture zone, which was about 500 m wide there, could not be determined, so it remained unclear whether they were left-lateral, rebound strains or right-lateral, permanent strains. Nor was it possible to relate the left-lateral, far-field rebound, the near-field right-lateral rupture zone, and the included main ruptures or faults.

A clue is provided by an example of right-lateral, permanent deformation along a right-lateral fault—not necessarily reflected in a belt of fracturing—that is illustrated, but not commented upon, in the 1908 Lawson report (Lawson, 1908, fig. 35, p. 98). The report shows a map of a water intake structure at one of the reservoirs south of San Francisco. There are several ruptures in the general vicinity, but not passing through the intake structure. The main rupture appears to be within 3 m of the edge of the intake structure. The opposite edge is about 32 m from the main rupture. The intake structure was a perfect circle, about 27 m in diameter, when constructed but became an ellipse during the earthquake, reflecting permanent right-lateral

¹⁹ In this discussion it will be convenient to describe what we see or infer next to right-lateral, strike-slip faults, although the observations are analogous, regardless of the kind of fault.

distortion parallel to the San Andreas fault zone. Measurements of the structure indicate that the maximum extension is about 0.15 in the direction N 80W, that the maximum shortening is about -0.15 to -0.19 in the direction N10°E, and that the main rupture here trends approximately N30W. Perhaps when faulting occurs in ground that is not as brittle as at Landers, a belt of right-lateral permanent, plastic distortion replaces the right-lateral rupture belt.

On the basis of these observations, one might propose an idealized picture of deformation zones and elastic-rebound zones adjacent to earthquake ruptures shown in Figure 6C. In Figure 6C the main rupture is central and belts of right-lateral deformation occur on either side. The belt of right-lateral deformation is consistent with the faulting offset, which we associate with permanent ground deformation. The left-lateral deformation is consistent with elastic rebound, which we associate with the generation of earthquake waves.

The survey network of Kaynaşlı Viaduct I presents an unusual opportunity to examine the details of deformation within ground on either side of a strike-slip earthquake rupture. The viaduct spans the right-lateral earthquake rupture zone, extending some 1900 m west and some 500 m east of the main rupture. The viaduct is supported by piers spaced about 39 m apart longitudinally and 21 m apart transversely to the axis of the viaduct (Figure 5) so the viaduct provides a detailed survey network with which to measure ground strains. The eastern third of the viaduct intersects the earthquake rupture at an angle of about 15°, so the length of the affected viaduct will be greater than the width of the deformation belt along the fault, showing considerable detail within the belt. We have an opportunity, here, then, to examine the entire belt of permanent deformation and perhaps even detect the zones of smaller left-lateral deformation on either side. This is why the phenomena at Kaynaşlı, Turkey, are so important to our understanding of ground deformation near earthquake ruptures.

THE DÜZCE EARTHQUAKE

In 1999 there were two major earthquakes in Turkey, the August 17 Izmit M 7.4 earthquake, centered south and east of Istanbul, and the November 12 Düzce M 7.1 earthquake, south of Düzce, extending from near Eften Lake to the mountains west of Bolu. The ground rupture of the August 17 earthquake was about 100 km long. The rupture of the November 17 earthquake essentially extended the earlier rupture another 40 km toward the east with trends ranging from about N75E to N75W (Barka and Altunel, 2000; Figure 7).

Figure 7. Major earthquake ruptures in the Düzce/Bolu area, Turkey during the 1999 earthquakes. Other major earthquakes in the Düzce area during the 1900s were along the southern branch of the Anatolian fault zone: the 1944 Bolu-Gerede earthquake ($M=7.4$), the 1957 Abant earthquake ($M=7.1$), and the 1967 Mudurnu Valley earthquake ($M=7.1$) (Ketin 1969, Ambraseys 1970). These events were predominantly strike-slip earthquakes (McKenzie 1972, Jackson and McKenzie 1984).

According to Barka and Altunel (2000), the Düzce earthquake rupture is on the northern side of the North Anatolian fault zone and its trend in this area follows a suture zone that was subjected to a combination of right-lateral faulting and thrusting. The North Anatolian fault zone here consists of two groups of faults on either side of a tectonic block known as the Almacik block. The Almacik block rotates between the two groups of faults. The southern branch of the Anatolian fault zone turns about 11° southward to the west of Bolu and goes through Lake Abant and then through the Mudurnu Valley as a series of right-stepping segments. The northern branch, the Düzce fault, steps right, to the north, and forms the southern boundary of the Düzce basin. West of Düzce the Düzce fault turns about 25° counterclockwise.

Barka and Altunel (2000) report that the ground rupture of the 12 November 1999 earthquake in the vicinity of Kaynaşlı (Figure 8) is a narrow deformation zone between half a metre and 50 m wide, but mostly between 2 and 5 m wide. Its most common offset is about 3 m of right-lateral, although there are vertical and even left-lateral offsets locally. The largest offset was reported to be 5 m, near Güven (Figure 7). The broader rupture zones contain tension cracks oriented $N50-75^\circ W$.

Figure 8. Trace of rupture of Düzce earthquake at Kaynaşlı (Modified after Barka and Altunel, 2000). Vertical lines are oriented N-S. E-W distance between vertical lines, 5 km.

Figure 8 shows the trace of the ground rupture near Kaynaşlı according to Barka and Altunel (2000). The rupture cuts highway E-5, the main road between Ankara and Istanbul, east of the area shown in Figure 8. In Kaynaşlı the surface rupture makes several right steps. Near Darıyeri Hasanbey village the rupture zone trends E-W. The maximum right-lateral offset there is about 3 m. The south side of the rupture is downthrown as much as 40 cm in this area. The right-lateral offset is about 0.7 m at Dip Mahallesi. There the surface rupture crosses the Asarsu valley and damages the columns of Kaynaşlı Viaduct I (Figure 2). Near the east end of the viaduct the right-lateral offset is about 1.15 m.

THE SURVEYS

Studies of deformations near earthquake ruptures in general use various means to measure displacements of control points in order to determine the patterns of displacement and the probable causes of those patterns. That is not what we are doing here. We are, instead, going to determine the deformation in the ground beneath the piers of Kaynaşlı Viaduct I. We are going to use measurements of distances between pairs of piers in groups in order to construct the equivalent of a strain ellipse. Then we shall determine with the signs of the principal strains and the direction of maximum strains whether the deformation is consistent with right-lateral or left-lateral distortion or some combination of distortion and area change.

For these reasons we are concerned with the accuracy of determining the change in spacing of piers after the Düzce earthquake. The approach we are following here is similar to that followed at Landers (Fleming *et al.*, 1996). We use the accuracy of the instrument to determine the error in computations of spacing of piers. In addition, we independently resurveyed points at about 2 m height for piers 38 through 57 using our own total station and have compared our results with those of nearby points provided by the ASTALDI-BAYINDIR Co (The results are at our website, www.eas.purdue.edu/physproc See Earthquake Ruptures, 1999 Düzce, Turkey).

Figure 9. Vertical view of four piers and their footing. Two corners of piers, e.g., d and D or b and B, were surveyed after the earthquake. Each group of four piers define two surveyor's quadrilaterals, involving points A,B,C,D, for one and points a, b, c and d for the other. Changes in lengths of each side or diagonal provide one strain measurement, so each group of four piers provides 12 strain measurements.

According to Timur (2000), the viaduct piers were measured with Leica reflectorless total stations, which allow the surveyor to determine orientations and measure distances without a reflector or prism up to distances of about 80 m. Also, a diagonal eyepiece allowed the piers to be plumbed accurately. This system was particularly valuable for surveying the piers because the measurements could be made high up on the piers at inaccessible points. The instrument used by Timur can

measure angles to 1.5 seconds and can measure distances to 2 mm +2 ppm accuracy. The total station can survey up to 3 km with a circular prism.

We use the survey measurements to determine the changes in distances between piers. One way to estimate the error of calculated changes is to determine the apparent changes, due to survey errors, alone. The spacing of piers is typically 39.2 to 39.3 m. Therefore the error of 2 mm + 2 ppm would correspond to 2.1×10^{-3} m. Assuming this error, the ratio of the error to the spacing of piers would be 5.3×10^{-5} or 0.0053 %. We shall consider strains equal to or smaller than 5×10^{-5} to be insignificant.

In using the spacing of piers to measure permanent ground deformation, we assume that the ground, piles, pile caps, and base of the piers moved together. All the pile caps were inspected by ASTALDI-BAYINDIR Co. There was no indication that any pile group moved with respect to the pile cap they supported.

DEFORMATION OF QUADRILATERALS OF RESURVEYED PIERS

To compute strains, we imagine the corners of piers to represent points; indeed, the four points forming the apices of surveyors' quadrilaterals, as shown in Figure 9. The viaduct provides us with a very long ladder of quadrilaterals.

In this analysis, we use our own survey data and the design coordinates of corners of the piers 38 through 57 according to ASTALDI-BAYINDIR Co. to determine the pattern of normalized length change in the eastern third of the viaduct.

Where the normalized length changes are relatively homogeneous, so that the normalized length changes of sides AB and DC on the one hand and AD and BC on the other are approximately the same, they behave much as strains and we shall assume that they do so. We can therefore use the data on normalized length change to determine the principal directions of strain.

We shall examine the calculation of a few values for sets of piers before presenting the results of the entire analysis. Figure 10 shows data for quadrilaterals for piers 44 and 45 (Quad 45), piers 45 and 46 (Quad 46), and piers 46 and 47 (Quad 47).

Each solid point in Figure 10 (except open circle) represents a normalized distance, such as the ratio of quadrilateral line AB measured after the earthquake (AB_a) and the design length (AB_b). Indeed, since we are treating the normalized length changes as strains, the normal strain, $S(?)$, in the azimuth, $?$, of line AB is calculated with the relations,

$$S(?) = AB_a/AB_b$$

$$\gamma(?) = (1/2)\{S(?)^2 - 1\}$$

In which $S(?)$ is the stretch (the ratio of the final to the initial length of a line element) of line element AB.

Thus, each of the points is represented by a strain value for a side or diagonal of a quadrilateral. There are four sides and two diagonals, so there are six points for each quadrilateral. Twelve points are plotted, though, because there are six points for the right quadrilateral (plotted as triangles) and six points for the left quadrilateral (plotted as squares).

Figure 10. Relations between azimuth of side or diagonal of quadrilateral and percent normalized length change ("strain"). Solid squares and triangles indicate strain measurements for the sides and diagonals of two quadrilaterals for each group of four piers. Open circles are averages for parallel sides or diagonals of the two quadrilaterals.

The large, circular points are mean values for a set of four piers. Two of them are averages of values for four points and two are averages for two points.

Finally, the solid line in each of the figures represents the theoretical relationship between strain, $\epsilon(\theta)$, and azimuth, θ . It is given by the expression,

$$\epsilon(\theta) = (1/2) \frac{1}{[\cos(\theta - \theta_1)/S_1]^2 + [\sin(\theta - \theta_1)/S_2]^2 - 1}$$

$$\theta_1 = (1/2)(S_1^2 - 1)$$

$$\theta_2 = (1/2)(S_2^2 - 1)$$

in which θ_1 is the maximum principal strain, θ_2 is the minimum principal strain, and θ_1 is the azimuth of the maximum principal strain. By adjusting the principal strains and the azimuth of the maximum principal strain, one can fit the solid curve to the data points, if the data points in fact represent homogeneous strain. In

all our analyses we have fit by eye, not by a statistical procedure. The solid lines shown in Figure 10 closely fit the mean strain measurements.

For quadrilateral 45, involving piers 44 R and 44 L, and 45 R and 45 L, the maximum principal strain is about 2 % and is oriented at an azimuth of 54° (N 54° E) (Figure 10). For quadrilateral 46, the maximum strain is 5 % and is oriented at an azimuth of 58° . For quadrilateral 47, the maximum strain is about 4 % and is oriented at an azimuth of about 54° .

The strain measurements are summarized in Figure 11 and Table I. The principal strains are largest near where the main rupture passed beneath the viaduct, but they are significant for all the quadrilaterals. The upper curve in Figure 11 is for maximum strains and the lower for minimum. We see that the maximum principal values are quite large from quadrilaterals 45 to 48, ranging from 2×10^{-2} to 5×10^{-2} (2 to 5 %). From quadrilaterals 39 to 44, the maximum principal strains are smaller, 0.5×10^{-2} to 10^{-2} (0.5 to 1 %), but still more than two orders of magnitude larger than the estimated instrument error of 5×10^{-5} (0.005 %).

The azimuth of principal strains is generally 55° between quadrilaterals 44–48 according to Table 1. The azimuth is generally slightly larger, 60° , for quadrilaterals 39–43 and decreases markedly between quadrilaterals 48–50.

The azimuth of principal strains indicates right-lateral distortion within a shear zone parallel to the trend of the main rupture near the viaduct. We can show this as follows. The orientation of the main rupture in the vicinity of the viaduct is roughly east-west, or 90° . Indeed, because it extends from pier 45 to pier 47 (Figure 5), the azimuth is about 98° . The viaduct between piers 39 and 49 has an azimuth of 113° , so the angle between the viaduct and the fault is about 15° . The maximum principal strain is at about 55° azimuth. Thus, the angle between the maximum principal strain and the main rupture is about 43° . If the strain associated with fault rupturing had been pure shear, we would expect the angle to be 45° . Therefore, the angle of the maximum strain is consistent with a zone of nearly pure shear along the trend of the main rupture.

Another observation that is consistent with right-lateral, simple shear parallel to the main rupture is the change in orientation of two of the piers. The broad faces of the piers, between piers 40 and 55 are oriented at an azimuth of $22.75^{\circ} \pm 0.25^{\circ}$. Piers 45 R (east bound) and 47 L (westbound), though, which lie along the trace of the main rupture, have been rotated clockwise by about 3° more than the average pier, to angles of 25.75° and 25.92° , respectively. Pier 48 L was rotated clockwise about half a degree, to 23.3° .

These results are also consistent with the reported offset of about 1.15 m near pier 47L. The width of the pile cap beneath the pier is 18.7 m. If one end of the slab moved 1.15 m with respect to the other in a clockwise sense, the slab would have rotated 3.5° . Also, the pattern of thrusting and depression formation around the pile caps of Piers 45 R and 47 L (Figure 5) is consistent with clockwise rotation in right-lateral simple shear.

The results are only slightly different for the stretch of the viaduct between piers 38 and 44. There the maximum principal strain has an azimuth of 60° , so the angle between the maximum principal strain and the fault rupture is 38° . This result suggests a combination of some compression normal to the fault but mostly right-lateral shearing along the fault.

The orientations of maximum strains are quite different toward the eastern end of the viaduct, as shown by results for quadrilaterals 48–50 (Table 1). The orientation of the maximum principal strain is changing from 54° , to 52° , to 25° to, perhaps, -85° in that area. The orientation of 25° may indicate that a zone of left-lateral shearing is being approached at the east end of the bridge. The left-lateral distortion is the sense of strain that we would expect to see associated with elastic rebound during a right-lateral earthquake rupture event. The value of -85° in quadrilateral 50, however, represents E-W shortening roughly parallel to the direction of the rupture zone.

On the strength of results suggesting a left-lateral, rebound distortion associated with the Düzce earthquake rupture based on our own survey of a subset of the piers, we have analyzed the post-earthquake survey data presented by ASTALDI-BAYINDIR Co. for the lower parts of the piers. The data are difficult to work with because they contain spurious values. As a result, the data had to be sorted and analyzed very carefully for inconsistencies, before they could be accepted for computation of principal strains. However, the results for quadrilateral 36, involving piers 35 and 36 and quadrilateral 35 involving piers 34 and 35 are consistent, larger than the expected error and instructive (Table 1 and Figure 12). The similarities of the magnitudes of

principal strains of the two results to those determined with our total station measurements are shown in Figure 11. The calculated maximum principal strain values (approximately 4×10^{-3}) are two orders of magnitude larger than the expected instrument error of 5×10^{-5} .

The results suggest a zone of left-lateral distortion appearing on the west side of the zone of right-lateral distortion. The azimuth of the principal strain is -10° to -15° for each example, so the shearing parallel to the trend of the earthquake rupture is smaller than the normal strains but the shearing has definitely switched from right- to left-lateral in the vicinity of piers 35 – 37. These strains may well portend the left-lateral distortion that occurs farther from the main rupture. Our results, therefore, are likely to have identified the entire belt of right-lateral shearing at Kaynash Viaduct I.

Figure 11. Distribution of principal normalized length changes ("strains") as a function of distance between pier 35 and the eastern end of the viaduct.

Figure 11 (addendum). Shows Orientations of maximum extension along viaduct. Orientations indicate right-lateral shearing between piers 38 and 48, and left-lateral shearing at piers 35 and 36.

Table 1. Summary of Principal Normalized Length Changes along Viaduct

QUAD. NO.	STRAIN, ϵ_1 (%)	ϵ_2 (%)	STRAIN, ϵ_1
50	0.2 9	- 0.05	2.9E -03
4 9	- 0.15	- 0.45	1.5E -03
48	1.2 0	- 1.80	1.2E -02
47	3.6 0	- 3.00	3.6E -02
46	4.9 0	- 3.10	4.9E -02
45	1.9 0	- 1.90	1.9E -02
44	0.5 0	- 0.45	5.0E -03
43	0.6 8	- 0.52	6.8E -03
42	0.8 0	- 0.50	8.0E -03
41	1.0 5	- 0.45	1.0E -02
40	0.8 0	- 0.40	8.0E -03
39	0.5 4	- 0.34	5.4E -03
38			
37			
36	0.4 4	- 0.24	4.4E -03
35	0.3 4	- 0.34	3.4E -03

The shortening of the viaduct over the 750 m length from pier 38 to pier 57 is approximately 2 m. The amount of net shift across the belt of right-lateral distortion—assuming that compression normal to the zone was negligible—was 2 m cos(15), or 1.9 m. Barka and Altunel (2000) report 1.15 m of right-lateral offset across a single rupture beneath the viaduct (Figure 3 and Figure 4). Our measurements indicate approximately 0.5 m shortening between piers 46 L and 47L and 1.1 m shortening between piers 47 L and 48

L. Thus the total amount of shift was $1.6 \text{ m} \cos(15^\circ)$, or 1.5 m over a horizontal distance of 77 m between piers 46 L and 48 L. The remainder of the zone of right-lateral shearing at Kaynaşlı, then, accommodated approximately 0.5 m of shift²⁰.

Figure 12. Relations between azimuth and normal strain for adjacent quadrilaterals that show excellent (low scatter) results and small negative azimuths. Larger circles are averages. Other symbols are 24 individual measurements for a quadrilateral of piers.

Although we have estimated that the width of the right-lateral distortion belt at the Kaynaşlı viaduct is on the order of 100 m, the width of the belt of distortions large enough to damage man-made structures is narrower. Deformation and damage to roads and sidewalks at Northridge and fracturing of rock and soil (Cruikshank *et al.*, 1997) provide a basis for estimating the width of damaging right-lateral distortion near the viaduct. Various measurements indicate that strains of 10^{-3} to 10^{-2} are sufficient to break rock or soil. Brittle rock, such as granite, fails under atmospheric confining pressures at compressive axial strain values of between -2 to -6×10^{-3} . In the Aspen Grove landslide in Utah, the soil in a developing landslide toe cracked at strains of about -1.4 to -2×10^{-2} and in a newly forming graben the strains were between 6×10^{-3} and 3.5×10^{-2} when a tension crack had formed (Fleming and Johnson, 1989). In the Granada Hills deformation zone at Northridge typical strains of 3×10^{-3} (maximum strains to 10^{-2}), produced abundant fracturing such as tension cracks in concrete and soil and reverse faults in concrete (Johnson *et al.*, 1996). Finally, in the Winnetka deformation

²⁰ We would note that similar displacement observations were made at Bodick Road at Landers by Johnson and others (1994), where about 1.4 m of a total of 1.8 m of shift across the 180 m width of the belt of shear zones were accommodated on two main ruptures, 0.2 to 0.3 m on one and 1.1 to 1.2 m on the other. About 0.4 m was distributed over the rest of the width of the belt.

zone typical strains were 3×10^{-4} (maximum strains of 10^{-3}) produced insignificant fracturing in sidewalks, pavement or soil (Cruikshank *et al.*, 1996):

Now let us compare these strain values with those at the Kaynaşlı Viaduct I. According to Table 1, the right-lateral zone over which strains exceed 10^{-2} extends from pier 44 to pier 48, a total distance of 145 m, which is equivalent to a strain belt of 40 m width. If we include strains of 5×10^{-3} and larger, the zone of damaging strains extends from pier 39 to pier 48, a total distance of 315 m, which is equivalent to a strain belt 80 m wide.

At Kaynaşlı Viaduct I, then, the right-lateral shear zone was 115 m wide, but the shear zone that produced damaging strains should have been between 40 and 80 m wide. This indicates that, although the most severe damage would be expected along the main break of the earthquake rupture, there should be broader belts of severe damage 40 to 80 m wide at Kaynaşlı.

Near Kaynaşlı Viaduct I, the severe damage should be mostly on the north side of the main rupture because the belt of damaging strains is on the north side of the main rupture. We would not generalize the location of damaging strains along the Düzce fault, though, except within Kaynaşlı itself. At Landers, California, the main rupture is on one side in some places and on the other side in other places along the belt of right-lateral distortion (Johnson *et al.*, 1996), and we know of no way to predict which side it will be.

Nevertheless, the observations and measurements of the piers of the Kaynaşlı Viaduct I provide a most interesting illustration of a phenomenon that was so clearly illustrated at Landers, California, in 1992 (Johnson *et al.*, 1994; 1996; Fleming and Johnson, 1997). Along the strike-slip rupture zones at Landers, strike-slip distortion, large enough to crack and fault the ground, occurred over a broad belt tens of metres wide as well as along one or more very narrow zones or faults within the belt. At Kaynaşlı Viaduct I, we did not observe a broad belt of rupturing, rather a broad belt of relatively high strains. Presumably, the ground would have ruptured if it had been brittle, as was the ground at Landers.

EXPLANATION OF RIGHT-LATERAL PROCESS ZONE ON EITHER SIDE OF RIGHT-LATERAL FAULT

Process Zone

We believe that we can explain what happened at Kaynaşlı in terms of permanent deformation near the ground surface accompanying upward propagation of a strike-slip fault. The right-lateral ground distortion is compared to a feature known as a *process zone* in fracture mechanics (e.g., Atkinson, 1987, p. 12). The process zone at Landers is the zone of fractures that reflects right-lateral distortion. The left-lateral distortion measured at Landers apparently reflects elastic rebound due to stress drop on the earthquake rupture. We believe the same thing happened at Kaynaşlı, except there was minor fracturing within the process zone because the soil was so ductile beneath the viaduct that it merely developed permanent deformations.

Assumptions and Observations

Based on this conception of the formation of the process zone along earthquake ruptures, we can derive an explanation for the opposite, right lateral and left lateral strain belts side-by-side if we make certain assumptions or accept certain observations as relevant:

1. The strike-slip faults that we have observed everywhere²¹ in Utah, California and in Turkey reflect mode III propagation of faults (Figure 13), that is, propagation of faults from below to the ground surface. At no

²¹ In the 1983-86 large landslides in Utah (Fleming and Johnson, 1989), in the 1989 Loma Prieta (Aydin *et al.*, 1991; Martosudarmo *et al.*, 1997), 1992 Landers (e.g., Johnson *et al.*, 1994; 1996) rupture zones and in the 1999 Düzce, Turkey zone of right-lateral distortion.

place have we observed evidence for mode II propagation, that is, propagation along the ground surface. Indeed there is abundant evidence that mode II propagation has not occurred.

Thus, we assume that the strain belts are a result of mode III loading of ground between the tip of a propagating fault and the ground surface.

2. We assume that the belt of permanent, right-lateral shearing occurs before the fault tip reaches the ground surface. Because the tip has not reached the ground surface, the fault is a “blind” strike-slip fault (i.e., invisible to the eye) at the time the process zone forms.

3. We assume that the faulting is an unloading phenomenon at depth, so that deformation we observe is a result of a drop in shear stress on the fault surface as the fault propagates toward the ground surface.

Figure 13. Modes of propagation of fractures and faults.

4. We assume that results of a static analysis of strains around a slipping fault are similar to the results of a dynamic analysis of a fault propagating upward during an earthquake.

5. We assume that a dislocation model of a fault, with uniform slip on the fault surface, adequately represents the deformation around such a fault.

6. We assume dimensional parameters for the fault of the Düzce earthquake determined by Yagi and Kikuchi (Online): Fault length 40 km; fault extends to depth of 20 km; maximum slip on fault 5.6 m.

Strains Above Strike-slip Fault Propagating in Mode III

With these assumptions, we can investigate the strains in the plane of the ground surface, as a right-lateral fault propagates in mode III toward the ground surface. We shall not present details of the analyses because they are included in several diagrams. The diagrams are available on the Internet in Volume II of a manual for the *faux pli?* computer programs. The computer program used for the analysis described in Volume II can be downloaded at one of our Purdue websites²².

²² See www.eas.purdue.edu/fauxpli . Download the gang fault program. It can be run on PC computers. Volume I of the manual provides instructions. Volume II of the manual, *Examples of Use of Sequential Gang/Listric Fault-related Folds*, can also be downloaded at the web site. See

We assume that the upper fault tip at different depths below the ground surface, and shall examine strains at the ground surface near midlength of the fault.

5 km Depth

For a fault with its upper edge at a depth of 5 km, the strains all show *right-lateral distortion*, everywhere within at least 8 km on either side of the fault trace (Figure 14C). Thus the width of the shear belt shown in Figure 6C is 16 km. The strains, however, are very small. The strains at the ground surface, even over the fault tip, are less than 10^{-4} (Figure 14A). The width of the belt of significant strains—those greater than $10^{-3.5}$ —is negligible (Figure 14B).

2 km Depth

For a fault with its upper edge at a depth of 2 km, the width of the belt of significant strains is 0.4 km (Figure 14B), and the entire width of the belt of right-lateral strains is about 5 km (Figure 14C) on either side of the fault trace. The zone of permanent, right-lateral deformation is thus at most 400 m wide on each side of the fault; most of the zone of right-lateral deformation would still be elastic when the fault had propagated to this depth.

1 km Depth

For a fault with its upper edge at a depth of 1 km, the pattern of strains clearly shows three zones of distortion. Within a belt about 3.7 km wide, on each side of the projection of the fault to the ground surface, the distortions are right-lateral. This belt corresponds to the shear belt shown in Figure 6C. The distortions die off to zero at the edges of this belt. Farther away, more than 3.7 km to either side of the fault trace, the distortions are left-lateral and are smaller than the distortions close to the fault trace.

In a zone approximately 1.2 km wide on each side of the fault trace the right-lateral distortion is at least $10^{-3.5}$. Thus, if the ground begins to fracture at a strain of $10^{-3.5}$, there would be a 2.4 km belt of minor fracturing centered on the fault trace. The maximum strain, over the fault tip, is still only 10^{-3} .

Chapter 2, *Broad Rupture Zones Along Earthquake Ruptures at Landers*, for the analysis that explains the belts of right-lateral distortion.

A.

B.

C.

Figure 14. Relation between depth to fault tip and A. maximum, right-lateral shearing, B. width of belt of significant right lateral shearing, and C. width of entire belt of right lateral shearing. The fault is assumed to be 40 km long and 20 km deep, and to have slipped 5.6 m.

0.5 km Depth

When the fault has propagated upward so that its upper edge is at a depth of 0.5 km, the width of the belt of significant right-lateral distortion (at least $10^{-3.5}$) has narrowed markedly, from approximately 1.2 to 0.4 km on either side of the fault trace (Figure 14C). The maximum strain is $10^{-2.8}$ near the fault trace (Figure 14A). If the ground were brittle, we would expect fracturing across a belt approximately 800 m wide.

0.1 km Depth

As the fault propagates to the ground surface, the belt of right-lateral shearing narrows and the magnitude of the right-lateral shearing increases. For a fault-tip depth of 0.1 km, the maximum strain at the ground surface is $10^{-2.5}$ and the belt of significant right-lateral shearing is only 50 m wide on either side of the fault trace. If the ground were brittle, the fracturing would extend across a belt approximately 100 m wide.

The analysis thus suggests that much of the permanent, fracturing or plastic flow that occurs in ground on either side of the fault occurs when the fault tip is propagating from a depth of about 0.5 km to the ground surface.

Results of Simulations

- ? The closer the fault tip to the ground surface, the narrower the belt of right-lateral distortion
- ? The maximum principal strain at the ground surface increases as the fault tip propagates upward and increases as the amount of strike slip increases.
- ? There is a certain depth of the fault tip at which the width of the belt of significant right-lateral shearing is maximized.

Figure 14 largely generalizes the results of the simulations. There is a very important difference, however, between the simulation and the real situation. In the simulation, of course, the deformations are linearly elastic, so the zone of right-lateral distortion narrows as the fault propagates toward the ground surface. In process zones of actual faults, though, a broad zone of right-lateral distortion at the ground surface remains as permanent strain as the fault propagates upward. The width of the zone of permanent strain depends on the properties (brittleness, for example) of the ground, the size of the fault, the depth of the fault tip below the ground surface, and the amount of slip on the fault. According to the theoretical analysis, the permanent right-lateral strain will be largest near that fault and will decrease with distance from the fault.

At Kaynaşlı, we observed a right-lateral belt about 115 m wide. According to our calculations, such a belt should have begun to form over a fault tip at a depth of approximately 100 to 200 m. According to Figure 14, the maximum principal strain over the fault tip should have been between 10^{-3} and $10^{-2.5}$ for fault slip of 5.6 m at depth. The value of 10^{-3} is consistent with the strains near the outer edges of the right-lateral shear zone at Kaynaşlı (Table 1). The larger strains, closer to the main rupture, would have occurred as the fault approached the ground surface and the right-lateral belt narrowed until, ultimately, the fault appeared as the main rupture, with extremely high and localized strains at the ground surface.

A sequence of events similar to that described above is, we believe, responsible for the right-lateral, process zones along faults near the ground surface.

IMPLICATIONS OF THE RESEARCH

The ground rupture near the eastern end of the 12 November 1999 Düzce earthquake crossed the alignment of the Kaynaşlı viaduct. The piers of the viaduct were a giant strain gage that produced a remarkable record of ground movement and distortion near a fault. The data indicate that along the right-lateral fault at Kaynaşlı, the fault zone consists of right-lateral movement at the main trace, a zone of right-lateral distortion near the trace, bounded by left-lateral distortion. The measurements of the Kaynaşlı viaduct, thus affirmed the forgotten conclusion from the studies of the 1906 San Francisco earthquake: earthquake ruptures and accompanying zones of distortion typically occur throughout zones or belts, rather than along linear traces or planes. Idealizations of belts of deformation are shown in Figure 6C or Figure 6D, consisting of a broad zone of right-lateral shearing, perhaps including one or more narrow zones of intense rupturing, bounded on either side by indefinite zones of left-lateral deformation.

The understanding we have gained about deformation belts and rupture belts along earthquake ruptures could be important to earthquake engineering for three reasons:

- ? An earthquake rupture at the ground surface must be considered, not as a fault line or plane, but as a fault zone with a finite width. The structural damage within the zone is caused, not primarily by ground acceleration, but by ground distortion.
- ? Deformation zones can be sites of large strains (10^{-3} to 10^{-2} or even higher) that can seriously damage foundations of structures.

- ? Deformation zones could be sinks of earthquake energy, perhaps causing ground shaking to be less—other things being equal—near the fault, than farther away from the fault.

ACKNOWLEDGEMENTS

Grant No. 85270-CMS of the National Science Foundation sponsored the project. Department of Energy, Grant No. DE FG02 98ER14886 supervised by Nick Woodward, provided some of the support for travel to Turkey. Appreciation is due ASTALDI-BAYINDIR Co. and Dr. Ç. Yılmaz for their generosity in making the data available to the project. The project is a joint effort of the Middle East Technical University and Purdue University with support from the National Science Foundation (Grant NTP1381). The other faculty who participated in the research were P. Gülkan, G. Özcebe, H. Sucuoğlu, A. Kılıç, A. Bobet, R. Frosch, K. Ridgway, G. Fernandez and J. Ramirez. Robert Fleming, USGS Ret., reviewed and improved the manuscript several times. The people of Kaynaşlı were wonderfully friendly during the fieldwork. They brought çay and ayran to drink and cool fruit to eat in the field. They shared numerous observations and insights into the earthquake. We are grateful to all these people for their help.

FOURTH OBJECTIVE: INVESTIGATE MECHANICAL ANALOGS THAT CORRESPOND WITH THE GEOMETRIC MODELS OF FAULT-RELATED FOLDING

MECHANICAL ANALYSIS OF THE GEOMETRY OF FORCED FOLDS

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ABSTRACT

We present a mechanical model of forced-folding comprised of a cover with power-law rheology overlying displaced, rigid basement blocks. We systematically investigate the influence of various parameters on the theoretical fold form: shape and dip of basement fault, basement-cover contact, and anisotropy of the cover. We show that the degree of anisotropy in the cover largely influences the geometry of the forelimb of the forced-fold. Folds produced in isotropic cover display forelimbs that taper from large dip angles near the basement-cover contact to low dip angles at the ground surface. In contrast, dips in the forelimbs of folds in anisotropic cover are nearly uniform with depth. We show that the basement-cover contact and the shape of the basement fault largely influence the geometry of the backlimb. Backlimb rotation occurs in cover welded to the basement and in cover overlying curved basement faults. In addition, the kinematic features of the theoretical folds are compared with the fold geometry generated by parallel kink and trishear kinematic models. Folds in isotropic cover overlying straight basement faults closely resemble the fold forms produced by the trishear kinematic model while fold forms in anisotropic cover more closely resemble folds produced by parallel kink geometric constructions.

INTRODUCTION

Typical structures in the Rocky Mountain Foreland of the western United States are fault-related folds that form over basement faults. Outcrops and seismic profiles show that the folds are typically asymmetric monoclines

with long, gently-dipping back-limbs and short, steeply-dipping forelimbs, overlying straight or curved fault surfaces in basement rock (e.g., Prucha *et al.*, 1965; Stearns, 1971; Reches, 1978; Stone, 1983a and b, 1985; Schmidt *et al.*, 1993). Forced-folding has been proposed as the mechanism for some of these structures. The essential features of the forced-folding mechanism are a sedimentary cover that deforms more or less passively and rigid basement blocks that are displaced along planar or listric faults (Stearns, 1978; Reches and Johnson, 1978). Evidence for undeformed, perhaps rigid, basement blocks that moved during folding of the sedimentary cover has been cited in several folds in the Rocky Mountain foreland (Prucha *et al.*, 1965; Stearns, 1971; Mathews, 1986; Erslev *et al.* 1988; Erslev and Rogers, 1993).

Efforts to explain forced-folds have followed three, largely divergent paths: theoretical analysis, experimentation, and kinematic analysis. The earliest study was a combined experimental/theoretical analysis by Sanford (1959), who experimented with clay models and performed theoretical analyses with elasticity theory. Reches and Johnson (1978) performed a similar kind of theoretical analysis, but for layers, in order to explore possible mechanisms of formation of the Palisades monocline in the Grand Canyon. Withjack *et al.* (1990) performed clay experiments of forced-folds over normal faults. Friedman *et al.* (1980) formed small forced-folds experimentally in rock specimens subjected to high pressures.

Kinematic models and geometric constructions have been the most common method of describing forced folds. The parallel kink construction of basement-involved folding by Narr and Suppe (1994) assumes that bed length and cross-sectional area is preserved, and bed thickness and limb dips are uniform (Fig. 1B). The folds are formed over faults composed of linked, straight-line segments. Mitra and Mount (1998) imagined two end-member descriptions of basement-involved structures (Fig. 1C and D) depending on whether the cover rock is welded to or detached from the basement. They assumed intuitively that deformation should be concentrated within a triangular zone if the cover is welded to basement and should be concentrated within a rectangular zone if the cover is detached from basement. A third method, the trishear kinematic model, was introduced by Erslev (1991) and later expounded by Hardy and Ford (1997), Allmendinger (1998), and Zehnder and Allmendinger (2000). Trishear was invented to produce "more nearly realistic" fold geometries with rounded hinges and variable forelimb dip angles (Fig. 1A). Trishear folds are formed by specifying constant velocities on either side of a central, triangular region (Fig. 1A) and specifying a velocity distribution within the triangular region that satisfies the continuity equation (conservation of area).

The purpose of this paper is to systematically investigate, with mechanical analysis, effects of various mechanical properties on fold form, and to compare the kinematic features, including fold geometry, produced by the theoretical analysis with those suggested by the various kinematic models. We will show that there are mechanical conditions under which parallel kink and trishear-like geometries could appear.

A MECHANICAL MODEL OF BASEMENT FORCED-FOLDING

We have developed a mechanical model of forced-folding using the approach of viscous folding theory outlined in Johnson and Fletcher (1994). The theory is for plane flow of an incompressible, nonlinear (including power-law) anisotropic material, as is indicated in the Appendix. Our analyses of forced folding are based on the following specific assumptions:

1. The basement consists of two rigid blocks separated by a fault. One block translates relative to the other along a planar fault or rotates along a curved fault.

2. The velocity normal to the contact between the basement and the cover is conserved across the basement-cover contact.
3. The sedimentary cover is adequately modeled as a single anisotropic (or isotropic) viscous layer. The anisotropy may be specified by the ratio, η_n/η_s , of the normal, η_n , to the shear, η_s , coefficients of viscosity (Fig. 2 and Appendix).
4. The sedimentary cover satisfies force equilibrium and conditions of incompressibility and continuity so that its behavior is described by a mathematical solution of a certain form of a biharmonic partial differential equation (Appendix).
5. The upper surface of the anisotropic material is the traction-free ground surface; that is, we specify that the tractions acting on the upper surface are zero.
6. The resistance to slip at the contact between the top of the basement and the bottom of sedimentary cover can be modeled by placing a thin film at the cover-basement contact that controls the amount of shear passed from the basement to the cover. By varying the ratio, $\eta_f/\sqrt{\eta_n\eta_s}$, of the film viscosity, η_f , to the effective cover viscosity, we can vary the resistance to slip at the contact ranging from free slip when $\eta_f/\sqrt{\eta_n\eta_s} \rightarrow 0$ to welded when $\eta_f/\sqrt{\eta_n\eta_s} \gg 1$.

Figure 1. Kinematic descriptions of forced-folds. Each model has a deformed cover overlying a rigid basement. The essential differences are in the deformation patterns assumed in the cover.

INFLUENCE OF ANISOTROPY ON FOLD FORM

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he development of folds in basement-cored structures could be influenced by the mechanical anisotropy of the sedimentary cover (Spang and Evans, 1988) because mechanical anisotropy and properties of contacts between

layers are known to largely control fold form in buckling of multilayers (e.g., Johnson, 1977; Johnson and Pfaff, 1989).

We can express the anisotropy of multilayered rocks in terms of a viscosity ratio, η_n/η_s , for flowage of rocks (Fig. 2). The viscosity ratio is the ratio of the viscosity for shortening or lengthening parallel to layers, η_n , to the viscosity for shearing parallel to layers, η_s . On this basis, there are three general categories of behavior. If the viscosity ratio is unity, the rock is considered isotropic. If the viscosity ratio is less than one, the rock is *anisotropic* and *soft* to *layer-parallel* normal straining (Fig. 2). If the viscosity ratio

Figure 2

Figure 2. Model parameters. Boundary conditions applied are a traction-free upper surface and a velocity distribution on the bottom interface. The anisotropy of the cover is specified by the value for the ratio of the normal to shear viscosity (η_n/η_s). A thin film is placed between the basement and cover. The resistance to slip at the basement-cover contact is specified by choosing a ratio of the viscosity of the thin film, η_f , to the effective viscosity **Error! Objects cannot be created from editing field codes.** of the cover.

is greater than one, the rock is *anisotropic* and *stiff* to *layer-parallel* normal straining²³.

We discuss the mechanical conditions that are responsible for layer-parallel soft and layer-parallel stiff in the Appendix. Here it is sufficient to note that multilayers composed of many thin layers and low resistance to sliding between layers are characterized by layer-parallel stiff ($\eta_n/\eta_s > 1$) responses to deformation (Biot, 1965; Johnson, 1977). In the same way, multilayers containing interbedded stiff and soft layers bonded together are layer-parallel stiff. Layer-parallel stiffness enhances flexural slip between stiff layers during folding and thus makes flexuring easy. In contrast, massively bedded, isotropic rock ($\eta_n/\eta_s = 1$) cannot deform by flexural slip and therefore makes bending more difficult. Bending is even more difficult if the layer is characterized as layer-parallel soft ($\eta_n/\eta_s < 1$). The layer tends to thicken or thin rather than to bend.

Figure 3 shows some examples of forced-folds produced with our mechanical model for basement faults of different types and for isotropic ($\eta_n/\eta_s = 1$) and anisotropic ($\eta_n/\eta_s = 3$) sedimentary

²³ Perhaps it would be clearer to some if we said that, if the viscosity ratio is less than one, the rock is *stiff* to *layer-parallel* shearing, whereas, if the viscosity ratio is greater than one, the rock is *soft* to *layer-parallel* shearing.

covers²⁴. In all these examples, the sedimentary cover was welded to the basement rocks ($\eta_f / \sqrt{\eta_n \eta_s} \gg 1$).

GENERAL RESULTS

The theoretical results show that the forelimb geometry is largely influenced by the anisotropy of the cover. In the isotropic cover, the forelimb tapers from steep dip angles at the bottom of the fold to more shallow dip angles at the surface, whereas the forelimb dips uniformly in the anisotropic cover. Furthermore, the width of the flexure is much less for the anisotropic than for the isotropic cover, and the width of the flexure is nearly constant for the anisotropic cover whereas it increases markedly up-section for the isotropic cover.

Figure 4 displays fold forms and strain ellipses for covers with viscosity ratios ranging from $\eta_n / \eta_s = 1/3$ to $\eta_n / \eta_s = 3$. The fold in the layer-parallel soft cover (Fig. 4A) displays more thickening in the synclinal hinge and more thinning in the anticlinal hinge than the folds in the isotropic and layer-parallel stiff covers (Figs. 4B and 4C respectively).

²⁴ Note that, if the cover were a power-law material subjected to mean shearing, the viscosity ratio would be equal to the power. For a power-law material of power three, the normal-to-shear viscosity ratio would be three. If the cover were bonded linear viscous layers of equal thickness, and the viscosity of the stiff layers were ten times that of the soft layers, the normal-to-shear viscosity ratio would be about three.

Figure 3

Figure 3. Forced-folds produced with the mechanical model. Cover welded to the basement in each example. The loading in the model is provided entirely by the displaced rigid basement blocks. The fold shape is influenced by the geometry of the fault and the anisotropy of the cover.

A. Layer-parallel soft cover ($\eta_n/\eta_s \approx 1/3$)

B. Isotropic cover ($\eta_n/\eta_s \approx 1$)

C. Layer-parallel stiff cover ($\eta_n/\eta_s \approx 3$)

Figure 4

Figure 4. Folds formed in isotropic and anisotropic cover overlying a 45° dipping reverse fault. In each example, the cover is welded to the basement. The fold in the cover with viscosity ratio $\eta_n/\eta_s = 1/3$ displays reason to expect the deformation of the forelimb to reflect the nature of the basement-cover contact.

more thickening in the synclinal hinge and more thinning in the anticlinal hinge than in the folds in the isotropic and layer-parallel stiff covers. The forelimb dips taper in the isotropic and layer-parallel soft covers but are uniform in dip in the layer-parallel stiff cover.

COMPARISON WITH EXPERIMENTS

We will investigate the role of anisotropy on forms of forced-folds using the geometry of experimental examples. Friedman *et al.* (1980) conducted experiments using lubricated layers of limestone and sandstone. Lubricated rock layers were stacked over rigid forcing blocks with a 66° dipping fault and subjected to confining pressure. Lubrication of the rock layers was intended to create an anisotropic cover medium and encourage flexural slip during folding. Figure 5A shows the deformed rock layers in Friedman *et al.*'s (1980) experiment.

We modeled the conditions of this experiment with our theory using an isotropic cover (Fig. 5B) and an anisotropic cover (Fig. 5C). The model with an anisotropic cover (viscosity ratio of $\eta_n/\eta_s=2$) (Fig. 5F) provides a better fit to the experimental fold than the model with an isotropic ($\eta_n/\eta_s=1$) cover, so the theoretical folds show that, indeed, the rock cover in the experiment behaved much like an anisotropic material.

As a counterproof, we compare our results with an experiment in which the cover should be isotropic. A forced-fold in clay material overlying a basement normal fault in an experiment by Withjack *et al.* (1990) is shown in Fig. 5D. We would expect a homogeneous block of clay to behave as an isotropic material. We modeled this normal fault using an isotropic cover (Fig. 5E) and an anisotropic cover (Fig. 5F). The results for the isotropic cover nearly perfectly fit the slopes and amplitudes of the top three layers of the clay model. In contrast, the slopes of the upper three layers in the anisotropic cover ($\eta_n/\eta_s=2$) are too steep to fit the clay layers.

COMPARISON WITH TRISHEAR AND PARALLEL KINK DESCRIPTIONS

The affect of anisotropy on fold form is shown in more detail in Fig. 6. The fold in the isotropic cover (Fig. 6A) broadens and shallows up section, whereas the fold in the anisotropic cover (Fig. 6B) has a nearly constant width and dip angle up section.

The geometry of the theoretical model with isotropic cover closely resembles the geometry produced by the trishear description, but not the geometry for the anisotropic cover. The fold in the isotropic cover (Fig. 6A) has a narrow forelimb near the faulted basement and a wide forelimb near the ground surface, much like the forelimbs in the trishear model. Thus the rounded trishear descriptions by Erslev (1991) and Allmendinger (1998) most closely represent the fold form for an isotropic sedimentary cover.

The forelimb in the layer-parallel stiff, anisotropic cover in Fig. 6B has a quite different geometry. In this fold, the cover has a viscosity ratio $\eta_n/\eta_s=3$ so that the medium deforms more easily in shear than in compression. The forelimb falls within a narrower rectangular region with uniform bed dips, much like the idealized parallel kink folds of Narr and Suppe (1994) (Fig. 1B) and the idealized, "cover-detached folds" of Mitra and Mount (1998) (Fig. 1C). The genetic terminology of Mitra and Mount is unfortunate

A. Experimental Model of Friedman et al. (1980)

D. Clay model of Withjack et al. (1990)

B. Isotropic ($\nu_n/\nu_s \approx 1$)

E. Isotropic ($\nu_n/\nu_s \approx 1$)

C. Anisotropic ($\nu_n/\nu_s \approx 2$)

F. Anisotropic ($\nu_n/\nu_s \approx 2$)

— theoretical
 experimental

Figure 5

Figure 5. Comparison of experiments with lubricated rock (A, B, C), and clay (D, E, F) and theory. The lubricated rock layers are better modeled with an anisotropic ($\eta_n/\eta_s = 2$) cover whereas the clay is better modeled as an isotropic cover.

because in the most closely corresponding theoretical fold (Fig. 5B), the cover is welded to the basement rocks so that detachment of the cover is impossible.

Thus the straight limbs and sharp hinges of the Narr and Suppe (1994) and Mitra and Mount (1998) constructions most closely correlate with the theoretical fold form that occurs in a highly anisotropic sedimentary cover.

INFLUENCE OF BASEMENT-COVER CONTACT ON FOLD FORM

Stearns (1978) documented that some folds in the Rocky Mountain foreland are characterized by sedimentary bedding that is thinned in the steep forelimb (e.g. Uncompahgre uplift), while in others the bedding maintains nearly constant thickness above the basement-cover contact (e.g. Rattlesnake Mountain). Stearns deduced that whether the cover rock is thinned depends on the basement-cover contact—sedimentary cover that is nearly constant in thicknesses through the forelimb was detached from the basement, while cover rock that is thinned in the forelimb was welded to the basement. Others, though, apparently dispute Stearn's deduction. Hodgson (1965) and Blackstone (1981) point out that there is no evidence to suggest that detachments occur at the cover-basement contacts in the Rocky Mountain foreland.

Mitra and Mount (1998) carried Stearn's deduction further, generalizing it by proposing two end-member models of forced folding depending on whether the cover units are welded (Fig. 1C) to or decoupled (Fig. 1D) from the basement. They speculated that, if the cover is welded to the basement, the forelimb has shallower dips distant from the basement fault and steeper dips near the basement fault. If the cover is detached from the basement, they suggest the forelimb dip angle is constant and the fold form resembles the parallel kink style of folding proposed by Narr and Suppe (1994).

With the mechanical model we can predict the influence of the basement-cover contact on the form of a forced-fold. We added a thin film at the basement-cover contact into the mechanical model so we can vary the resistance to slip at the base of the cover. We do so by varying the ratio of the viscosity of the film to a measure of the viscosity of the anisotropic cover through a parameter, $\eta_f/\sqrt{\eta_n\eta_s}$ (Fig. 2). A large ratio of $\eta_f/\sqrt{\eta_n\eta_s}$ simulates a cover welded to the basement (Fig. 7A). A small ratio of $\eta_f/\sqrt{\eta_n\eta_s}$ simulates a cover that is free to slip relative to the basement (Fig. 7B).

Figure 7 shows four examples of solutions for low-angle, reverse faulting of a rigid basement overlain by homogenous sedimentary cover; Figs. 7A and 7B are for a homogeneous overburden, and Figs. 7C and 7D are for an anisotropic overburden. The figure shows that the resistance to slip at the contact between the cover and basement has the largest influence on the anticlinal hinge and backlimb geometry. One difference is that the anticlinal hinge in the welded cover (Fig. 7A) is amplified, bulged upward, more than in the detached cover (Fig. 7B). This apparently is a result of layer-parallel shortening induced in the cover by the reverse faulting in the basement. The more highly amplified fold displays a slightly steeper forelimb and a slight backlimb rotation.

Indeed, the theory shows that the change in forelimb shape is a result of the rheologic properties of the overburden, and the experiments by Friedman *et al.* (1980) and Withjack *et al.* (1990) support

this conclusion. Clearly, the kink-like forms (Figs. 7C and 7D) occur in the anisotropic cover and widening-outward forms (Figs. 7A and 7B) occur in the isotropic cover, whether the cover is welded to or detached from the basement. Based on this analysis, there appears to be no sound, mechanical

A. Isotropic ($\mu_n/\mu_s \approx 1$)

B. Anisotropic ($\mu_n/\mu_s \approx 3$)

C. Anisotropic and Isotropic Solutions Superimposed

Figure 6

Figure 6. Comparison of folds formed in isotropic and anisotropic covers above 45-degree reverse fault in a rigid basement. Dashed lines in A and B are axial planes. Cover welded to basement.

Figure 7

Figure 7. Folds formed over a 30° dipping thrust fault. The folds in the welded cover display bulging of the anticlinal hinge and slight backlimb rotation. The folds in the detached cover display no backlimb rotation.

DISCUSSION AND CONCLUSIONS

In the kinematic approaches to describing forced folding, one makes various *ad hoc* assumptions about velocity or displacement distributions and then attempts to match the geometry of actual fault-related folds. The kinematic descriptions can do a remarkably good job of simulating the gross geometry of actual basement-cored, fault-related folds. The problem is that, if the geometries of the model of a forced fold and the actual forced fold resemble one another, the *ad hoc* displacement or velocity distributions are generally assumed to be valid descriptions of the kinematics of actual folding processes (e.g., Erslev, 1991; Almendinger, 1998). This is a poor assumption.

The central limitation of the kinematic approach is that kinematic descriptions do not concede understanding of how various parameters, such as the rheology of the cover rock and nature of the basement-cover contact, influence kinematics of a folding process, including fold geometry. The kinematic approach is in stark contrast with the mechanical approach to folding processes, in which one completely solves problems mathematically based on mechanical principles, boundary conditions and rheology (e.g., Johnson, 1970; Johnson and Fletcher, 1994) and from which the kinematics follows as a logical consequence. The kinematic description is a product, a result, not the *cause* of a process.

Our mechanical model of forced-folds allows us to investigate the influence of 1.) Shape and dip of basement fault, 2.) Basement-cover contact, and 3.) Rheology of cover on fold form. The forced folds shown in Figs. 3 and 7 show that the geometry of the forelimb is largely influenced by the rheology of the cover and essentially uninfluenced by the shape and dip of the basement fault and

by resistance to slip at the basement-cover contact. In each of the folds produced in an isotropic cover (left column of Fig. 3), the forelimb is thickened near the synclinal hinge and thinned between the anticlinal and synclinal hinges, regardless of the geometry of the basement fault. The degree of anisotropy in the cover, however, largely influences the geometry of the forelimb. In each of the folds produced in isotropic cover, the forelimb tapers from range from large dip angles near the basement-cover contact to low dip angles at the ground surface. In contrast, the dips of layering in the forelimb of folds in anisotropic cover are nearly uniform with depth. The forelimbs in isotropic cover closely resemble the fold forms produced by the trishear kinematic description (Fig. 1A), while the forelimbs in the anisotropic cover more closely resemble the fold forms produced by the parallel kink construction of Narr and Suppe and Mitra and Mount (1998) (Fig. 1). Theory indicates the different conditions under which fold forms resembling the two quite different kinematic descriptions can be produced. The theory also indicates the conditions for all gradations of form between the end members.

Our treatment of anisotropy closely models a cover medium comprised of many thin layers. Therefore we can see how a cover that has many thin, slipping layers will produce a tight forelimb with uniform dip, much like a kink band (Reches and Johnson, 1976). The results shown in Fig. 6 emphasize this fact.

We have also shown that the lubricated sandstone and limestone layers in the Friedman *et al.* (1980) experimental model behave much like a viscous material with anisotropy $\eta_n/\eta_s = 2$. The clay material in the Withjack *et al.* (1990) model has no slip surfaces and is best modeled with the isotropic mechanical model. Thus, the experiments produce forms predicted by the theory with the known material properties.

We should emphasize here that the forced-fold mechanical model is intended to help us understand the mechanism of forced-folding, and we do not suggest that the mechanism is appropriate for all basement-involved foreland structures. While some authors have suggested that some foreland folds form over rigid, displaced basement blocks (Prucha *et al.*, 1965; Stearns, 1971; Mathews, 1986; Erslev *et al.* 1988; Erslev and Rogers, 1993), others provide evidence that the basement is folded (Berg, 1962; Blackstone, 1983; Brown, 1984; Narr, 1993; Narr and Suppe, 1994). Basement-involved structures that fold the basement may not form under the same mechanism as forced-folds. Similarly, the forced-fold theoretical model does not include propagation of faults from the basement into the cover. The mechanical behavior of fault-propagation folds was addressed by Wei (1997).

APPENDIX

Outline of Theory

Although the folding theory used to derive the results presented here has been largely presented elsewhere, the actual implementation here is quite different, so we will explain the origin of the theory. In our most nearly general model of forced folding, we are analyzing a layer of a Reiner-Rivlin, nonlinear fluid (Johnson and Fletcher, 1994, p. 413) subjected to a mean flow and a perturbed flow. In this case, the perturbed plane flow satisfies equations of the form of eqs. (8.3.3) (op cit., p. 374),

$$(\eta_n^2 - \eta_s)(\partial^2 \tilde{v}_x / \partial x^2) + \eta_s (\partial^2 \tilde{v}_x / \partial z^2) = (\partial \tilde{P} / \partial x) \quad (1a)$$

$$(2\eta_n - \eta_s)(\partial^2 \tilde{v}_z / \partial z^2) + \eta_s(\partial^2 \tilde{v}_z / \partial x^2) = (\partial \tilde{P} / \partial z) \quad (1b)$$

in which v is velocity, P is mean stress, η_n is normal viscosity and η_s is shear viscosity. The normal and shear viscosities enter through the rheological equations for plane flow, eqs. (8.2.7) (op. cit., p. 373)?

$$\tilde{\tau}_{xx} = 2\eta_n \tilde{D}_{xx} - \tilde{P} \quad (2a)$$

$$\tilde{\tau}_{zz} = 2\eta_n \tilde{D}_{zz} - \tilde{P} \quad (2b)$$

$$\tilde{\tau}_{xz} = 2\eta_s \tilde{D}_{xz} \quad (2c)$$

in which D is deformation rate.

The velocities are given by the stream function, eqs. (8.3.3) (op. cit.),

$$\tilde{v}_x = -(\partial \tilde{\psi} / \partial z); \quad \tilde{v}_z = (\partial \tilde{\psi} / \partial x) \quad (3)$$

which is a solution to equation, eq. (8.3.4) (op. cit., p. 375)

$$\frac{\partial^4 \tilde{\psi}}{\partial x^4} + 2(2\frac{\eta_n}{\eta_s} - 1)\frac{\partial^4 \tilde{\psi}}{\partial x^2 \partial z^2} + \frac{\partial^4 \tilde{\psi}}{\partial z^4} = 0 \quad (4)$$

The controlling parameter for the type of solution of this equation is the ratio of the normal to the shear viscosity, η_n/η_s . Suppose the nonlinear material is strain-rate softening, as is a power-law material. Then, as Johnson and Fletcher (1994) show, the viscosity ratio is less than one if there is mean layer-parallel shortening of the layer and is greater than one if there is mean layer-parallel (and normal) shearing of the layer (Table I). The mean flow environment, therefore, has a major, known effect on the way the nonlinear sedimentary overburden behaves. If there is general shearing, as there is with strictly forced folding, we expect the viscosity ratio to be greater than one (Table I).

What makes the solution even more powerful is that the gross rheological properties of a viscous multilayer can also be represented in terms of the same viscosity ratio, but in this case the ratio is determined by the relative viscosities, thicknesses and contract properties of the interbedded layers (Biot, 1965; Johnson and Page, 1976; Johnson, 1977; Johnson and Fletcher, 1994, p. 419). If the layers are interbedded stiff and soft materials, the viscosity ratio, η_n/η_s , is greater than one for low slopes of layers (Table I). We note finally a result not shown in Table I, that the viscosity ratio becomes less than one as the layers become inclined to angles greater than 22.5° (e.g., Biot, 1965; Spang and Evans, 1988).

Figure 8

Figure 8. The basement-cover interface for a normal fault. The lower interface and the velocities applied to the lower interface are expressed as Fourier series. The applied velocities and the shape of the interface are determined from the assumed kinematics of the basement blocks. Because the width of the normal fault increases with time, the velocity distribution applied to the bottom is a function of time.

Table I. Anisotropic Behaviors of Materials

Material	Overall Layer-Parallel Loading	Layer-Parallel Soft (Error! < 1)	Isotropic ($\frac{\gamma_n}{\gamma_s} = 1$)	Layer-Parallel Stiff (Error! > 1)
Linear single viscous (n=1)	any		x	
Linear single elastic	any		x	
Linear multilayer (bonded)	any			X
Linear mul	any			X

tilayer (free slip)				
Power-law viscous ($n > 1$)	Shortening	x		
"	Shear			X
Power-law viscous ($n < 1$)	Shortening			X
"	Shear	x		
Non linear elastic	Shortening			X
"	Shear	x		

Finally, we would note that we have found it simplest to solve the differential equation using complex exponentials, so the general, sinusoidal solution is of the form (Johnson and Fletcher, 1994, p. 375),

$$\eta = (1/l)[C_1 \exp(m_1 lz) + C_2 \exp(-m_1 lz) + C_3 \exp(m_2 lz) + C_4 \exp(-m_2 lz)] \sin(lx)$$

where $C_1 \dots C_4$ are complex constants, and the complex roots are

$$m_{1,2} = \sqrt{(2\frac{\eta_a}{\eta_s} - 1)} \pm \sqrt{\{(2\frac{\eta_a}{\eta_s} - 1)^2 - 1\}}$$

Figure 9

Figure 9. The basement-cover interface for a reverse. The lower interface and the velocities applied to the lower interface are expressed as Fourier series. The applied velocities and the shape of the interface are determined from the assumed kinematics of the basement blocks. The box function in B. is used to approximate the interface to avoid the problem of the overhang created at the origin in A.

The rest of the solution is straightforward and as explained in many examples by Johnson and Fletcher (1994).

Treatment of Boundary Conditions along Basement-Cover Contact

As discussed in the main body of the text, we apply a velocity boundary condition at the basement-cover contact that is determined from the assumed kinematics of the rigid basement blocks. For the normal fault geometry in Fig. 8A, the shape of the lower interface of the cover material has the form shown in Fig. 8B. We assume that the top and bottom of the rigid basement blocks move parallel to the fault and that the fault surface is stationary. Thus the appropriate velocity distribution that is applied to the bottom interface of the cover is as shown in Fig. 8C. We can represent the interface shape and the velocity distributions as a Fourier series. For the normal fault, the wavelength, L , of the Fourier series is a function of time, because the fault surface grows in width (x direction) with time. This is illustrated with the two time steps shown in Fig. 8.

The shape of the basement-cover interface for the reverse fault (Fig. 9) is a bit tricky. We cannot fit a Fourier series to the reverse fault interface because of the overhang that is created at the origin in Fig. 9A. Thus we approximate the interface with the box function in Fig. 9B. While the interface shape actually models a 90° dipping fault, we can approximate the kinematics of any fault dip by applying an x -component of velocity as well as a z -component of velocity as shown in Fig. 9C.

The velocity distribution applied to the base of cover overlying a listric shaped fault is determined under the assumption that the listric fault is a segment of a circle and thus the tilting of the basement block is rigid body rotation along the circular fault (Fig 10A). Because we have the problem of overhang, exactly as we had with the reverse fault, we approximate the interface shape as in Fig. 10B. The vertical velocities in Fig. 10C are a direct result of the tilting of the basement block. The horizontal velocities applied in Fig. 10D are a result of the shortening required to move the tip of the basement fault from its original position, before tilting, to its present position.

Treatment of basement-cover contact

We place a thin film between the basement and cover to allow variable resistance to slip at the basement-cover contact. A general method for a nonlinear, power-law viscous film is outlined in Johnson and Pfaff (1989). Here we use the case of a linear viscous film. The film is assumed to be so thin that it can deform only in simple shear. For plain strain, the shear stress parallel to the film is

$$\sigma_{ns} = 2BD_{ns} \quad (5)$$

where B is a material constant and D is deformation rate. The rate of shearing deformation is given by the rate of slip, $\dot{\gamma}_s$, across the film

$$D_{ns} = (1/2)(\delta v_s / h_f), \quad (6)$$

where

$$\delta v_s = (v_s)_t - (v_s)_b, \quad (7)$$

h_f is the thickness of the film, and $(v_s)_t$ and $(v_s)_b$ are the velocities parallel to the film at the bottom and top surfaces. Thus at the bottom of the cover we have the boundary condition

$$\delta v_s \approx (h_f / \mu_f) \sigma_{ns} . \quad (8)$$

ACKNOWLEDGEMENTS

T

his work was performed under Department of Energy grant DE-FG02-98ER14886, supervised by Nick Woodward of DOE. We thank Rick Allmendinger for reviewing a preliminary copy of the manuscript.

Figure 10

Figure 10. The basement-cover interface for a listric fault. The lower interface and the velocities applied to the lower interface are expressed as Fourier series. The applied velocities and the shape of the interface are determined from the assumed kinematics of the basement blocks. To

avoid the problem created by the overhang at the origin in A, a vertical fault in B is used to approximate reverse listric faults.

MECHANICAL MODELS OF TRISHEAR-LIKE FOLDS

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ABSTRACT

Previous workers have formulated velocity descriptions of the trishear kinematic model of fault propagation folds, which are inherently nonunique. We present two complete mechanical models of fault-related folding and assess the validity of the assumptions used in the assignment of velocities in the trishear description and to eliminate the untenable situation of an infinite number of possible solutions for velocity fields. The mechanical model of forced-folding, *Forced Fold*, based on viscous folding theory, is used to derive the velocity fields in an anisotropic sedimentary cover overlying faulted and displaced rigid basement blocks. The solution of displacements around a stress-free fault in an elastic body is used to model fault-arrest folds that form around a fault imbedded in a deformed medium. The results indicate that the velocity fields assumed in the trishear model more closely resemble the velocities derived in the mechanical *Forced-Fold* model than the mechanical model of fault-arrest folding. The mechanical model *Forced Fold* also allows one to make a rational choice between the two competing descriptions of the geometry of forced folds—rounded, trishear forms on the one hand and angular, kink-like forms on the other. The theory shows that the descriptions are essentially the same mechanically; one can select one or the other or any in between simply by changing the anisotropy of the sedimentary cover in the theoretical model.

INTRODUCTION

In the last decade, structural geologists have devised various geometric descriptions of forced-folds and fault-propagation folds to reproduce the geometry of basement-involved uplifts of the Rocky Mountain foreland (Erslev, 1991; Narr and Suppe, 1994; Hardy and Ford, 1997; Allmendinger, 1998; Mitra and Mount, 1998). The curvilinear, trishear geometric description (Fig. 1A) was introduced by Erslev (1991) to replicate the geometry of fault-propagation folds more nearly accurately than the parallel kink-like (Fig. 1B) graphical constructions by, for example, Suppe (1985) and Narr and Suppe (1994). More recently, Hardy and Ford (1997) and Zehnder and Allmendinger (2000) have devised velocity methods (Waltham and Hardy, 1995) of the so-called trishear type. The velocity methods create fold forms by specifying velocities at points in the medium surrounding the fold-

producing fault. The velocities are assigned according to a set of *ad hoc* rules which are assumed to describe the folding process.

Although many authors have compared fold forms generated with the trishear method with fold forms in experiments and in the field, the validity of the assumed trishear velocity distributions has not been checked with results derived from theoretical models of fault-related folding. The need for such theoretical models was recognized by Zehnder and Allmendinger (2000), who showed that one can generate an infinite number of velocity distributions that satisfy the conditions considered in the kinematic models of Hardy and Ford (1997).

In this paper we present two mechanical models of fault-related folding and compare the velocities and displacements produced by the theoretical model with the velocity fields assumed in the trishear kinematic description.

VELOCITY FIELDS AND FOLD FORMS ACCORDING TO TRISHEAR KINEMATIC DESCRIPTION

General Conditions

The essential elements of the Erslev (1991) trishear geometric construction consist of three regions in the vicinity of a fault tip (Fig. 2A). In the velocity-based descriptions of trishear, Hardy and Ford (1997) and Zehnder and Allmendinger (2000) assume that the same three regions are regions of different velocities. They assign (assume) constant velocity fields in regions I and II. The material in region I, above the fault, moves as a rigid body in the x-direction, parallel to the fault. The material in region II, below the fault, is also rigid but does not move. Since the materials in rigid regions I and II move relative to one another, the material in the triangular, "trishear" zone (zone III in Fig. 1A), must deform. The resulting fold form is computed at multiple time steps by a simple zeroth-order numerical integration of the positions of points defining passive markers within the trishear zone; the displacements at points are calculated by multiplying the velocity at a point by the time step.

In order to describe how the material within the trishear zone moves, they assume:

1. The material does not change volume. They assume that the two-dimensional continuity equation, $\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} = 0$, applies. They have implicitly assumed plane flow, that is, the velocity in the z-direction is zero. Application of the continuity equation automatically guarantees that cross-sectional area is preserved (if density remains constant), which is a sufficient condition for balanced structural profiles.
2. The velocity normal to the fault is zero along the boundary between regions I and III and the velocity parallel to the fault is equal to the velocity parallel to the fault throughout region I. Thus one boundary condition is, $v_x = v_0, v_y = 0 @ y = x \tan \theta_1$
3. The material within trishear region III is welded to the rigid material in region II so that a second boundary condition is that, along the boundary between regions II and III, $v_x = v_y = 0 @ y = x \tan \theta_2$. Therefore, the empirical parameters that can be varied are θ_1 and θ_2 , v_0 , and the velocity distribution, v_x and v_y , within the trishear zone. Although authors vary these parameters when attempting to reproduce the geometry of folds, only one of the parameters is clear: v_0 is a rigid-body motion. It is quite unclear

what physical, mechanical conditions the parameters ϕ_1 and ϕ_2 or the velocity distribution, v_x and v_y , describe (Allmendinger, 1998). The trishear method is applied to fault-propagation folds by specifying that the fault tip propagates into the triangular region.

Erslev geometric construction

Erslev (1991) devised two types of trishear deformation, what he called homogeneous and heterogeneous trishear. Figure 3 displays the fold forms of both types. In his constructions, line

Figure 1

Figure 1. Geometric constructions of fault-related folds. A. Trishear fold (Erslev, 1991). B. Fault-propagation fold (Narr and Suppe, 1994).

segments perpendicular to the fault, called tie lines, are constructed in the triangular region. Homogeneous trishear rotates the tie lines uniformly throughout deformation, whereas in heterogeneous trishear, segments of tie lines in the center of the triangular zone rotate more than those near the triangle boundaries. The homogeneous construction generates a narrow forelimb with forelimb dips that are nearly uniform with depth (Fig. 3A). The heterogeneous construction generates a widening-upward forelimb with dips that shallow up section (Fig. 3B).

The folds in Fig. 3 were generated with a 45° reverse fault and with $\phi_1 = \phi_2 = 55^\circ$. We will show that there are mechanical conditions under which folds resembling both of these constructions occur.

Hardy/Ford Kinematic Description

Hardy and Ford (1997) divided the triangular trishear region (III, Fig. 2A)

Figure 2. A. The geometry assumed in trishear. Velocities are uniform and parallel to the fault in region I and zero in region II. A velocity distribution satisfying continuity and boundary conditions is assigned in region III. B. An example of a trishear velocity field derived by Zehnder and Allmendinger (2000) for the case $\phi_1 = \phi_2 = 15^\circ$.

A. homogeneous trishear

B. heterogeneous trishear

Figure 3

Figure 3. Two types of trishear (after Erslev, 1991). A. "Homogeneous shearing". B. "Heterogeneous shearing."

into sectors and satisfied continuity in each sector (Fig. 4). They assumed the magnitude of the velocity within each sector is constant, but decreases linearly from the top of the triangular region to the bottom. The components of velocity in the i^{th} sector are given by the expressions

$$v_x = v_i \cos(\theta_i) \quad (1a)$$

$$v_y = v_i \sin(\theta_i) \quad (1b)$$

where v_i is the net velocity in the i^{th} sector and θ_i is direction of the velocity vector, determined by solving eqs. (1) together with the so-called block contact condition

$$(v_y)_i = (v_x)_i \tan \phi = (v_y)_{i+1} = (v_x)_{i+1} \tan \phi \quad (2)$$

where f is the boundary. The expressions for the velocity components (eqs. 1) satisfy continuity.

Zehnder/Allmendinger Kinematic Descriptions

Zehnder and Allmendinger (2000) investigated more thoroughly the velocity description of trishear. They presented a general method of deriving solutions to the continuity equation that satisfy the boundary conditions. Figure 2B is one of their velocity field solutions, which satisfies the three sets of conditions outlined above. As one follows a row of grid points, such as that immediately above the fault

B.

Figure2

Figure 4

Figure 4. The trishear velocity field devised by Hardy and Ford (1997).

tip, one sees that the vectors decrease in length and turn downward more with increasing horizontal distance from the fault tip. As one follows a column of grid points, one sees that the vectors decrease more rapidly in length and turn downward more with increasing depth.

The analysis by Zehnder and Allmendinger (2000) presents a dilemma, though. How does one select a solution? They show that there is in fact an infinite number of possible velocity formulations that satisfy the continuity equation within the triangular trishear region and satisfy the boundary conditions specified by Hardy and Ford (1997). This result illustrates a fatal weakness of the trishear kinematic model (indeed, of any strictly kinematic description of a process): in kinematic descriptions there is an infinite number of ways to assign motions that satisfy the condition of continuity (area conservation). Many of these motions are physically untenable. Thus it is clear that other conditions, in addition to continuity, must be satisfied to describe the geometry of fault-related folds (or any other physical process).

VELOCITY FIELDS AND FOLD FORMS ACCORDING TO FORCED-FOLD MECHANICAL MODEL

It is well known in continuum mechanics (e.g., Love, 1944; Timoshenko and Goodier, 1951; Malvern, 1969; Johnson, 1970, 1977; Johnson and Fletcher, 1994; Fletcher and Pollard, 1999) that properly formulated mechanical models of processes, such as fault-related folding, include these other conditions. The solutions provide the velocity field uniquely for a given set of conditions. A complete mechanical model contains several basic equations: constitutive relations, compatibility of strains (elastic body) or of deformation rates (viscous body), equilibrium equations, and sufficient traction and velocity boundary conditions, all of which are necessary to determine the displacement (or velocity) and stress fields in the deforming medium. In some cases other equations, such as the continuity equations, are also invoked depending on the types of processes being modeled.

We will consider mechanical models of two very different mechanisms of fault-related folding. The first is a solution derived by Johnson and Johnson (in review) that uses modern viscous folding theory (Johnson and Fletcher, 1994) to model forced-folds. The second uses the solution

given by Tada *et al.* (1985) for the displacements around a crack imbedded in an infinite elastic body subjected to far-field stresses.

Solution of Boundary-value Problem

All of the following equations are required in order to derive a solution for problems of forced folding of a viscous anisotropic fluid. For plane flow (zero flow in the y -direction) the *constitutive equations* are given by eqs. (8.2.7) (Johnson and Fletcher, 1994, p. 373)?

$$\tau_{xx} = 2\eta_n \tilde{D}_{xx} - \tilde{P} \quad (3a)$$

$$\tau_{zz} = 2\eta_n \tilde{D}_{zz} - \tilde{P} \quad (3b)$$

$$\tau_{xz} = 2\eta_s \tilde{D}_{xz} \quad (3c)$$

in which η_n is the normal and η_s is the shear coefficient of viscosity, $\tilde{P}$ is the mean pressure, τ is stress and D is deformation rate,

$$\tilde{D}_{xx} = (\partial \tilde{v}_x / \partial x) \quad (4a)$$

$$\tilde{D}_{zz} = (\partial \tilde{v}_z / \partial z) \quad (4b)$$

$$\tilde{D}_{xz} = (1/2)[(\partial \tilde{v}_x / \partial z) + (\partial \tilde{v}_z / \partial x)] \quad (4c)$$

The condition of *no volume change* is that,

$$\tilde{D}_{xx} + \tilde{D}_{zz} = (\partial \tilde{v}_z / \partial z) + (\partial \tilde{v}_x / \partial x) = 0 \quad (5)$$

The *equations of force and moment equilibrium*, combined with the constitutive equations, give differential equations of the form of eqs. (8.3.3) (op cit., p. 374),

$$(2\eta_n - \eta_s)(\partial^2 \tilde{v}_x / \partial x^2) + \eta_s(\partial^2 \tilde{v}_x / \partial z^2) = (\partial \tilde{P} / \partial x) \quad (6a)$$

$$(2\eta_n - \eta_s)(\partial^2 \tilde{v}_z / \partial z^2) + \eta_s(\partial^2 \tilde{v}_z / \partial x^2) = (\partial \tilde{P} / \partial z) \quad (6b)$$

Introducing a stream function, ψ , such that (eqs. (8.3.3); op. cit.),

$$\tilde{v}_x = -(\partial \psi / \partial z) \quad (7a)$$

$$\tilde{v}_z = (\partial \psi / \partial x) \quad (7b)$$

we have the general, fourth-order differential equation that needs to be solved (eq. (8.3.4), op. cit., p. 375),

$$\frac{\eta^4}{\eta_x^4} + 2(2\frac{\eta_n}{\eta_s} - 1)\frac{\eta^4}{\eta_x^2 \eta_z^2} + \frac{\eta^4}{\eta_z^4} = 0 \quad (8)$$

The relevant solution to the velocities in the mechanical model has the form

$$\tilde{v}_x = \sum_{m=1}^{\infty} [a_m g_1 e^{g_1 m z} + b_m g_1 e^{-g_1 m z} + c_m g_2 e^{g_2 m z} + d_m g_2 e^{-g_2 m z}] e^{i m x} \quad (9a)$$

$$\tilde{v}_z = \sum_{m=1}^{\infty} [i a_m g_1 e^{g_1 m z} - i b_m g_1 e^{-g_1 m z} + c_m g_2 e^{g_2 m z} - d_m g_2 e^{-g_2 m z}] e^{i m x} \quad (9b)$$

$$g_{1,2} = \sqrt{2\frac{\eta_n}{\eta_s} - 1} \pm \sqrt{2\frac{\eta_n}{\eta_s} - 1} \pm 1 \quad (9c)$$

where the complex coefficients a_m , b_m , c_m , and d_m are determined by the boundary conditions. We solve for the mean pressure by substituting eqs. (9) into (6) and integrating. Then we solve for the stresses by substituting the resulting expression for mean pressure and eqs. (9) into the constitutive equations (3). Thus we have the complete solution, for velocities and stresses, throughout the body, to within values given by the boundary conditions. Note that we did not have to guess anything. The velocities and stresses come from the solution. All that remains is to evaluate the arbitrary constants in terms of the boundary conditions.

As discussed in Johnson and Johnson (in review), we apply a velocity boundary condition at the basement-cover contact, determined from the rigid motion of basement blocks along a fault, and a traction-free boundary condition on the upper ground surface (The normal and shear tractions remain zero on the ground surface). We have set up the solution so that we can study effects of several variables:

- ? Shape of the fault. The fault may be a plane or a circular cylindrical arc.
- ? Slip on the fault. Only dip slip is allowed, but faulting may be normal or reverse.
- ? Anisotropy (η_n/η_s) of the cover. The anisotropy ratio is the ratio of the coefficient of viscosity in layer-parallel and layer-normal shearing to the coefficient of viscosity in layer-parallel and layer-normal compression. Each value of anisotropy (η_n/η_s) models certain rock properties. For example, a sedimentary cover composed of interbedded stiff and soft layers that cannot slip might have $\eta_n/\eta_s = 3$. For many thin layers slipping freely, the value would be even larger.
- ? Thickness of the cover.
- ? Basement-cover contact. We can vary the resistance to slip along the basement/cover contact from completely detached (zero resistance to sliding) to completely welded (no slippage).

We have written and compiled a computer program in Matlab, *Forced-Fold*²⁵, that implements the solution with a GUI (graphical user's interface).

Planar or Listric Basement Fault

Figure 5 displays fold forms and velocities generated by *Forced Fold* which models the deformation of layered sedimentary cover overlying displaced, rigid basement blocks (Johnson and Johnson, in review). The folds shown in Fig. 5 were formed in an isotropic (linear) cover welded to the basement with different basement-fault geometries. The footwall is fixed in each example by specifying zero velocities along the basement-cover contact on the footwall, and non-zero velocities parallel to the fault along the basement-cover contact on the hangingwall.

In the case of planar basement faults, the velocities are nearly uniform in magnitude and direction over the hangingwall block and parallel to the fault. The velocities above the footwall block range from small horizontal velocities at the ground surface to zero at the basement-cover contact. Thus the velocity distributions show that the velocities follow the general pattern assumed in the Hardy/Ford and Zehnder/Allmendinger trishear kinematic description for planar basement faults (Figs. 5B and 5D) but not for listric faults (Fig. 5F).

The overall shapes of the folds are roughly monoclinial for the planar faults and asymmetric/anticlinal for the listric fault. For both normal and reverse basement faults, the monoclinial fold is narrow near the tip of the fault and broadens upward to the ground surface. For the listric fault, the shape is more strongly asymmetric near the fault tip than near the surface.

The shapes of the forelimbs of the forced folds in isotropic cover are roughly the same, regardless of the geometry of the basement fault. The forelimb thickens near the synclinal hinge and thins between the anticlinal and synclinal hinges.

Anisotropy of the cover

Thus far we have explored the effect of shape and sense of slip on the basement fault on the displacement and velocity distributions. The sedimentary cover has been isotropic and has been welded to the rigid basement blocks. Now we consider only reverse faulting and explore the effect of anisotropy of the sedimentary cover on the forms of the displacement and velocity distributions. All other conditions are the same.

Figure 6 shows the forms and velocity distributions for isotropic ($\eta_n/\eta_s = 1$) and anisotropic ($\eta_n/\eta_s = 5$) cover welded to the basement. It is clear that the anisotropy of the cover markedly influences the geometry of the forced-fold. The forelimb in isotropic cover increasingly widens upward from the tip of the fault where the limb dips are steeper to the top surface where the limb dips are shallower. The forelimb in the anisotropic cover is essentially of constant width and dip from near the fault tip to the ground surface. Thus the forelimb closely resembles a kink fold with rounded hinges.

The forelimbs in anisotropic cover resemble those in the parallel kink constructions (Fig. 3B). The forelimb width and dips are uniform with depth in the cover. The forelimb in the forced-fold

²⁵ A trial version of the computer program *Forced-Fold* may be downloaded from the Purdue University, Department of Earth and Atmospheric Sciences *Faux Pli Software* website: <http://www.eas.purdue.edu/fauxpli/>.

formed in isotropic cover resembles the forelimb in Erslev's (1991) so-called "heterogeneous" trishear folds (Fig. 3A). The kink-like forelimb in the forced-fold formed in anisotropic cover resembles the forelimb of Erslev's (1991) so-called "homogeneous" trishear folds.

With the results thus far, the forced-fold theoretical model has already explained an important condition responsible for strikingly different perceptions of the shapes of forced-folds—the sharp kink-like forced folds depicted by Narr and Suppe (1994), Mitra and Mount (1998) and many others, and the rounded, more open forced folds depicted by Erslev (1991) and others. By varying the anisotropy of the sedimentary cover one can vary the shape of the forced fold, over a wide range of shapes.

Figure 5. Fold forms and velocities calculated with the *Forced-Fold* mechanical model. Each fold is produced in an isotropic cover welded to the basement.

Figure 5. Effect of shape and sense of slip of basement fault on fold forms and velocities in sedimentary cover. Fold forms and velocities calculated with the Forced-Fold mechanical model. Each fold is produced in an isotropic cover welded to the basement.

Figure 6

Figure 6. Effect of anisotropy and degree of detachment from basement of fold form. Fold forms and velocities calculated with Forced-Fold mechanical model for reverse faulting. On left are solutions for isotropic cover. On right are solutions for anisotropic cover. Top row is for no resistance to slip, middle row for moderate resistance to slip, and lower row for very high resistance to slip between cover and rigid basement.

Nature of Basement/cover Contact

The fold form and velocity distributions are affected less dramatically by the bonding between the sedimentary cover and the basement. The bonding considered in the theoretical model ranges from complete detachment (free slip) to welding of the sedimentary cover to the basement. For both isotropic and anisotropic sedimentary cover, the degree of bonding determines the direction of displacement in the sedimentary cover over the hangingwall basement block. If detached, the sedimentary cover moves vertically, even though the faulting is reverse. If slippage is resisted but possible, the cover moves obliquely to the fault, and if completely welded, the sedimentary cover moves in the direction of faulting.

Figure 7

Figure 7. Effect of cover thickness on fold form. The dips are shallower in the 500 unit thick cover than in the cover with thickness 100 units.

Figure 6 shows that, regardless of the degree anisotropy of the cover, much of the cover above the hangingwall block has uniform velocities. The velocities above the hangingwall block are nearly uniform and parallel to the fault surface in the cover that is welded to the basement. In addition, the velocities over the footwall block are larger in the cover welded to the basement than in the cover that is partly or totally detached from the basement.

In spite of the different velocity distributions imparted by basement rock that is detached, partly bonded or welded to the sedimentary cover, the forms of the forced-folds are largely unaffected by the nature of the bonding between the sedimentary cover and the basement blocks.

Thickness of the cover

The thickness of the sedimentary cover also has a minor influence on the fold form of a forced-fold. The thickness of the cover mainly influences the dip of the forelimb. As the thickness increases, the influence of the upper, free surface becomes less significant and the limb dips decrease. Fig. 7A displays two superimposed folds formed in an isotropic cover welded to a basement with a 45° dipping fault. The only parameter varied in the two folds was the thickness of the cover. The fold in the cover with thickness of 100 units as steeper limb dips than the fold in the cover with thickness of 500 units.

Comparison with Trishear Description

We can show with theoretical examples that the assumption of a triangular zone of deformation in the Erslev trishear geometric construction and the Hardy/Ford trishear kinematic description is reasonable for the forced-fold mechanism under certain conditions. The theoretical examples also can show what the empirical variables that Erslev chose for his analysis reflect in terms that have physical meaning.

The gross form of the velocity distributions of the forced-fold and the trishear kinematic description are similar. Figure 8 shows a triangular boundary, with vertex angle of 110°, strategically placed in the velocity distribution of the theoretical fold shown in Fig. 4B; the cover is isotropic and welded to the basement. It is clear that one can imagine a roughly triangular zone within which the velocities are changing, that is, within which deformation is concentrated. Outside of the triangular zone, above the hangingwall block, the cover behaves as a nearly rigid body; the velocities are uniformly parallel to the fault. The footwall block was assumed rigid in our model, so there velocities are zero there, as in the region below the fault in trishear. The velocities within this triangular region grossly resemble the velocities in the triangular region of a trishear model. As shown in Fig. 2B, the velocities in a trishear model of Zehnder and Allmendinger range from a maximum velocity at the top of the triangular zone, over the hangingwall, to zero at the bottom.

The theoretical models, however, show differences in the velocity distributions for some conditions. We also see from the theoretical results that the trishear assumption about the movement of the hanging wall block fails if the cover is not perfectly welded to the basement. It is assumed in the trishear model that the velocities outside of the triangular region, above the hanging wall, are uniform and parallel to the fault surface. We show in Fig. 6 that this is the case in the mechanical model only when the cover is welded to the basement. When the cover is not completely welded to the basement, there is a discontinuity in velocity at the basement-cover contact.

The theoretical analysis also shows that there are conditions under which Erslev's (1991) so-called "homogeneous" and "heterogeneous" trishear-like folds can occur and that the different shapes depicted by Erslev apparently are results of different anisotropies of the sedimentary cover. As discussed in Johnson and Johnson (in review), an anisotropic cover with $\eta/\eta_s > 1$ simulates a sedimentary cover composed of many thin mechanical layers while an isotropic cover most closely simulates a sedimentary cover lacking mechanical layers.

Note that in every forced-fold that we have produced theoretically, the trishear-like triangular region of localized deformation falls within a triangle with a vertex angle of 110°. Although the vertex angle is a parameter that is free to be varied in the trishear model, it is unclear, as pointed

out by Allmendinger (1998), what physical property would control such a parameter. Our theoretical results suggest that the only appropriate vertex angle is about 110° .

We have adjusted parameters within the theoretical model until we have obtained excellent matches between the geometric constructions for "homogeneous" and "heterogeneous" trishear folds and the theoretical models (Fig. 9). The comparable theoretical model for the "heterogeneous" fold (Fig. 9A), with a forelimb that broadens upward and limb dips that taper from shallow at the ground surface to steep near the fault, is the isotropic cover welded to rigid basement (Figs. 9A and 6). The comparable theoretical model for the "homogeneous" fold (Fig. 9B), with nearly uniform forelimb dips and width, is the moderately anisotropic cover ($\eta_n/\eta_s = 2$) welded to rigid basement (Fig. 9B).

In order to extend the theoretical model to an even more nearly kink-like shape, with straight limbs and sharp hinges, we would select a higher anisotropy. If the anisotropy is increased to the larger ratio of five, $\eta_n/\eta_s = 5$, as in Fig. 6, the forelimb geometry begins to suggest the extreme parallel kink constructions (e.g., Narr and Suppe, 1994).

Figure 8

Figure 8. Solution for velocities according to Forced-Fold with superposition of hypothetical trishear boundaries. Isotropic cover welded to basement.

Figure 9.

Figure 9. Two forms of trishear according to Erslev (1991) and most nearly comparable reverse fault theoretical fold forms. A Erslev's "heterogeneous trishear" and Forced-Fold with isotropic cover and welded to rigid basement. A. Erslev's "homogeneous" and Forced-Fold with anisotropic (anisotropy = 2) cover welded to rigid basement.

FOLD FORM FOR FAULT-ARREST MECHANICAL MODEL

In the previous sections of this paper we have compared the geometry and kinematics of the trishear construction to geometry and kinematics of the forced-fold theoretical model. The trishear geometry appears to be best modeled in terms of an isotropic or anisotropic cover welded to the basement. In the forced-fold model, though, we have assumed a rigid basement. The trishear construction is supposed to describe the deformation near the tip of a fault in a body, not necessarily associated with a rigid basement. Thus we are motivated to compare the trishear construction to fold forms produced near the tip of a fault near a free surface in a homogeneous, isotropic body.

We use the solution of Tada et al. (1985) to calculate the displacements around the tip of an inclined, two-dimensional fault imbedded in an incompressible, elastic medium with a traction free surface (ground surface). The way we use this solution, the infinite elastic medium is initially loaded and then the shear stress along the imbedded crack is dropped. The fold is produced by stress drop along the fault. Here we assume a succession of stress drops so that a fold of significant amplitude is produced. The problem is inherently nonlinear and the theory of elasticity is inherently linear, but we improve the accuracy of the solution of fold form by using a Runge-Kutta²⁶, fourth-order integration of the large slip event. The displacements in an incompressible elastic material

Figure 10

²⁶ We used the same method of integration for the viscous flow problems discussed in previous pages.

Figure 10. Solutions for single faults superimposed to derive an approximate solution for an inclined fault near the ground surface. A. The shear and normal tractions must be zero on the upper (ground) surface. B. The same sense and amount of slip are applied to a fault inclined 45° and its image across the upper surface to cancel the normal and shear stresses at the upper, ground surface.

4.1 Solution of Boundary-value Problem

around a fault along which the shear stress drops by an amount τ_{xy} are,

$$u_x = \frac{\tau_{xy}}{2G} \left[l \{ \operatorname{erf} \sqrt{z^2 - a^2} \} + y \left(\operatorname{Re} \left\{ \frac{1}{\sqrt{1 - (a/z)^2}} \right\} - 2 \right) \right] \quad (1a)$$

$$u_y = -\frac{\tau_{xy}}{2G} \left[y \operatorname{Im} \left\{ \frac{1}{\sqrt{1 - (a/z)^2}} \right\} \right] \quad (1b)$$

in which Re and Im are real and complex parts, and z is the complex coordinate,

$$z = x + iy.$$

We let the x -axis be vertical and the y -axis horizontal, and draw grid lines parallel to the horizontal, y -axis (Fig. 4.17). The horizontal lines within the elastic body are passive markers.

While this elastic solution provides displacements rather than velocities, as in the viscous solution for forced-folding, we can compare the fold forms. After integrating the velocities of the trishear kinematic description, we compare the displacement field of the trishear model with the displacement field of the fault in the elastic medium.

In order to derive an approximate solution for the effect of the traction-free ground surface, we superimpose solutions for two faults. The solution needs to satisfy the condition that the shear and normal tractions are zero on the upper (ground) surface. In order to model a traction free ground surface, we reflect the fault and put its center an equal distance into the reflected medium above the ground surface (Fig. 10) and apply the same sense and amount of slip on both faults. If the fault and its image are inclined at 45° as shown in Fig. 4, the shear and normal tractions on the surface sum to zero, so the superposition solves the problem approximately. Although the approximation is satisfactory for our purposes, more nearly exact solutions can be obtained by other methods²⁷.

FOLD FORM NEAR TIP OF NEAR-SURFACE FAULT

Figure 11 shows a 45° dipping fault in an elastic medium that is initially loaded in horizontal compression. Straight horizontal markers were drawn after loading. The medium is initially loaded such that, after the stresses along the fault are dropped, and the fault slips, the maximum displacement along the fault is 0.2 times the fault length. The load is maintained, and the markers remain deformed.

The fold is a fault-arrest fold, because the fault was not allowed to propagate. It simply slipped, maintaining its length. The shape of the fold in the passive markers is an asymmetric fold (Fig.

²⁷ E.g., see Pollard and Holzhausen, (1979).

11B). Near the ground surface the passive layers are folded nearly symmetrically. Near the fault tip the folding is nearly monoclinical, with steeply-dipping layers in the forelimb and gently-dipping layers in the rotated backlimb. The displacements around the fault are shown in Fig. 11C. Away from the fault the displacements are roughly horizontal. Closer to the fault, the displacements increase and become parallel with the fault. Uplift is generated just behind the fault tip where the displacements are largely vertical.

The forelimb geometry of the fault-arrest fold is similar to the forelimb geometry of the forced-fold (Figs. 5 and 6) in isotropic media. In each fold, the forelimb dips steeply near the fault tip and dips more shallowly at the ground surface. In both models, the layering is thinned in the forelimb and thickened in the anticlinal hinge. A characteristic of the fault-arrest fold that is strikingly different from the forced-folds overlying straight basement faults is the existence of backlimb rotation. The back limb dips gently and nearly uniformly to the left in the fault-arrest fold. The forced-folds overlying straight basement faults (Figs. 5 and 6) display little backlimb rotation.

The deformation in the fault-arrest fold is clearly different from trishear deformation. While the forelimb geometry resembles the forelimb geometry of trishear folds—with steeply dipping layers near the fault tip and decreasing dips towards the ground surface—the backlimb is folded. Under the assumptions of the trishear construction, a change in dip of the fault is necessary to create a folded backlimb (Allmendinger, 1998). For a straight fault, trishear assumes that the backlimb is undeformed. As a result of the backlimb rotation, there is no trishear-like triangular zone of localized deformation in the fault-arrest fold.

CONCLUSIONS

A complete mechanical solution, *Forced-Fold*, provides a unique solution for each set of parameters and eliminates the untenable situation of an indefinite number of possible solutions for velocity fields recognized by Zehnder and Allmendinger (2000).

We have demonstrated in this paper that the assumptions used in the trishear description—specifically, the triangular zone of flow and the boundary conditions—are more nearly appropriate for the forced-fold mechanism than the fault-arrest fold mechanism. Based on the results of the mechanical analyses of velocities and displacements in fault-related folds, *Forced-Fold* produces uniform velocities above the hangingwall of the basement fault, much like the uniform velocity distribution assumed above the hangingwall and outside the triangular region of the trishear construction. Furthermore, in the case that the cover is welded to the basement, the velocities above the hangingwall block are parallel to the fault, as is assumed in trishear. A triangular region of concentrated deformation with vertex angle of about 110° , similar to the trishear triangular zone, can be imagined in the forced-folds. As illustrated in Fig. 8, the lower boundary of the triangular region would lie on the footwall side of the basement-cover contact, below which the velocities are zero. The upper boundary of the triangular region would lie above the hangingwall side of the fault.

An added benefit of the *Forced-Fold* mechanical model is that a change of a single parameter, anisotropy, produces fold forms that range from the rounded shapes of “homogeneous” and “heterogeneous” trishear of the Erslev (1991) geometric construction to the angular, kink-like shapes

Figure 11. A. Geometry of the fault and passive markers in the mechanical model of a fault imbedded in an elastic medium. B. Fold form produced by dropping the stresses on the crack in the initially loaded medium. The maximum displacement on the fault is 0.2 times the fault length. C. Displacements in the elastic medium after the fault slips. of the Suppe (1985) fault-propagation fold and the Narr and Suppe (1994) and Mitra and Mount, (1998) forced folds.

We have also compared the displacements around a fault in an elastic medium with the trishear displacements because the trishear description is supposed to describe fault-propagation folding. The trishear assumptions are clearly irrelevant to the problem of a fault imbedded in a deformed medium, such as the fault-arrest elastic model. The fault-arrest model produces a complicated field of displacements that cannot be produced with trishear. The backlimb of the fault-arrest fold is deformed and thus there is not a distinct triangular region of concentrated deformation around the tip of the fault, as is assumed in the trishear model.

DISCUSSION

For the last 20 years there has been a redoubling of efforts in structural geology to invent geometric constructions of a wide variety of fold shapes. The efforts initiated far back in geology, with the early, kink-like constructions of Holland (1914; also, Badgley, 1959) and the concentric, circular-arc constructions of folds by Busk (1929). Suppe (1980, 1983) applied a Holland-type construction (fault-bend-fold) to produce detachment-ramp folds with sharp hinges and straight limbs.

Over the years the same geometric approach has been applied to forced-folds and fault-propagation folds to produce the so-called parallel kink folds. In the kink-like geometric constructions of Suppe (1985), Narr and Suppe (1994), and Mitra and Mount (1998), the deformation is assumed to be concentrated within a rectangular or triangular zone above the fault tip, in which the forelimb has angular hinges and straight limbs, and the area is conserved. The forelimb dip is constructed by making certain arbitrary assumptions about the forelimb geometry. The parallel kink constructions assume that the anticlinal and synclinal axial planes are parallel to the fault and that the layers do not change thickness.

The Erslev (1991) trishear geometric model was invented to produce "more nearly realistic" fold geometries with rounded hinges and variable forelimb dip angles (Fig. 1A). The trishear kinematic velocity scheme introduced by Hardy and Ford (1997) and Zehnder and Allmendinger (2000) increased the credibility of the parallel kink constructions by showing that a more rounded fold form could be the result of a specific velocity distribution. The trishear models improved on the geometric approaches by removing the arbitrariness by which the forelimb geometry is constructed and by assuming velocity fields within the triangular zone that satisfy the continuity equation. The assignment of the velocity field is still arbitrary, however, as there are infinitely many legitimate velocity fields that satisfy continuity, and the trishear method provides no other conditions to limit the possible velocity distributions. Thus the kinematic trishear schemes are nonunique. For this reason, and reasons also asserted by Fletcher and Pollard (1999), such *ad hoc* kinematic schemes of the types put forward by Hardy and Ford (1997) and Zehnder and Allmendinger (2000) for trishear folds are inadequate descriptions of physical processes. Furthermore, there may be no physical process that can produce an idealized fold created with an arbitrarily selected trishear velocity distribution.

While we have demonstrated in this paper that the assumptions used in the trishear description—specifically, the triangular zone of flow and the boundary conditions—are reasonable for some forced-folds, it seems quite unclear what one learns from the trishear description. We find ourselves asking many of the questions Fletcher and Pollard (1999) were asking. Why not describe a fold in a straightforward way? If one wants to understand how a forced-fold formed, why would one assume the kinematics as is done in the trishear geometric construction or the trishear kinematic description? The kinematics are a *result* of a physical process, not the *cause*. Theory is available and easily implemented to solve problems with much richer diversity and with understandable physical parameters. What does one learn, for example, with the trishear description alone, that the sedimentary cover somehow developed "heterogeneous" flow or "homogeneous" flow?

The simple *Forced Fold* model, with consideration of anisotropy, and based on a series of clear assumptions and physical parameters, produces fold forms that range from the relatively rounded folds resembling the trishear model to sharp, kink-like folds. One changes the anisotropy ratio from 1 for a rounded form to 3 or higher for a kink-like form. Furthermore, there are physical reasons behind the selection of the ratio. For example, interbedded stiff and soft layers of equal thickness welded together but with elasticity moduli or viscosity coefficients different by a factor of 10 will have a ratio of normal-to-shear viscosity coefficient (or elasticity modulus) of about 3. If slippage is easier, the ratio will be even higher. If the cover of a forced-fold behaves as a nonlinear,

power-law material (subjected to shearing), the ratio of effective normal-to-shear viscosity will be on the order of the power itself, say 3 or 5 (see Johnson and Johnson, in review, for more discussion of effective anisotropy).

With the theoretical model, we can readily determine the effect of the thickness of the sedimentary cover, simply by changing the thickness in the *Forced-Fold* program. We see in this way that the fold broadens with increasing thickness. Likewise, we can readily determine the effect of drag at the basement/cover contact by varying a parameter that reflects the connection or detachment of the cover to the basement.

Clearly the theory can generate a rich variety of fold forms. None of these parameters and none of this understanding of the physical processes of forced-folding are contained in the various kinematic and geometric constructions.

STUDY OF MONOCLINE ON EAST SIDE OF SAN RAFAEL SWELL

Kaj Johnson did his M.S. thesis on the structure of the monocline on the eastern side of San Rafael Swell, Utah. Following are some of the manuscripts that resulted from his study:

TYPES OF INTERNAL DEFORMATION IN THE SAN RAFAEL MONOCLINE

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SPOTTED WOLF SECTION

The study area is on the eastern side of the San Rafael Swell (Figure 1) along a section of the monocline forming the eastern flank. This section of the monocline, called the Spotted Wolf section, extends 10 km south of Spotted Wolf Canyon (I 70) (Figure 2). This section was chosen for study because it includes a wide range of dip angles in the monoclinal limb, within the same rock units. The monocline at locations 1 and 7 (Figure 2) is relatively broad, where the central limb dips 30° or less. At locations 3 through 6 the fold is much tighter and, in places, the beds are overturned. The radius of curvature of the synclinal hinge is about 760 m at location 1, near I 70, whereas it is 40 m at location 4, near midlength of the Spotted Wolf section, where beds dip 60°.

Internal deformation of the rocks in the Spotted Wolf section occurs by faulting, jointing, and kink folding. In this chapter I wish to provide a brief overview of the types and significance of the internal deformation observed as well as motivate the various methods of structural analysis developed in following chapters.

DEFORMATION BANDS

What They Are

Much of the faulting along the Spotted Wolf Section is in the form of deformation bands²⁸, which are ~1 mm wide tabular bodies of highly localized cataclasis in porous sandstones (Figure 3). Small amounts of displacement (mm) are accommodated across single bands. Larger displacements (1 to 10 cm) can occur along zones where many bands form side by side (Aydin and Johnson, 1978). Deformation bands are distinctive features in sandstones. Because the quartz grains within the band have been crushed and the porosity has been greatly reduced, the bands form resistant white ridges in sandstone (Figure 3).

A.

B.

Figure 1

²⁸ Also known as Atilla faults, micro faults, seal faults, granulation seams. See description by Aydin (1978).

Figure 1. Spotted Wolf Section of east side of San Rafael monocline, southeastern Utah.

Figure 2. Geologic setting of Spotted Wolf section of San Rafael monocline, Utah.

Deformation bands have been studied in detail by several investigators (Aydin, 1977, 1978; Aydin and Johnson, 1978, 1983; Jamison and Stearns, 1982; Underhill and Woodcock, 1987; Cruikshank et al., 1991; Zhao and Johnson, 1991; Antonelli et al., 1994).

While slickenlines are usually not present on deformation bands³⁰, the amount and sense of offset (the slip vector) can sometimes be determined with measurements on two offset planes. The orientation of two planes (in this case cross-beds or deformation bands) offset by the deformation band is measured in the field as well as the amount and sense of offset along the deformation band. With these measurements the slip vector can be calculated. I have written a MATLAB® m-file to automate this process (see Appendix).

Where They Occur

Deformation bands at the San Rafael Swell occur within the Coconino, Wingate Sandstone, Navajo Sandstone, and sandstone beds of the Carmel Formation. I studied deformation bands within the Navajo and Carmel where they are most prominent and easiest to work with. When slip direction and sense can be determined, the deformation bands are excellent indicators of stress-state, as a stress inversion algorithm can be implemented to estimate the local stress tensor.

As discussed by Davis (1999), deformation bands are pervasive throughout sandstone units in southern Utah. Deformation bands are most highly concentrated near large structures such as the San Rafael Swell where the rock is highly deformed, but surprisingly, deformation bands formed

³⁰ Slickenlines are occasionally found in zones of deformation bands, where many single bands have formed side by side.

even in flat lying, otherwise non-deformed rocks far from any major geologic structure. At San Rafael Swell, the deformation bands are most highly concentrated within the Spotted Wolf section, where the limb of the fold dips steeply. Relatively few deformation bands form in the upper and lower limbs where beds dip shallowly. I mapped and determined slip directions, sense, and amount of slip on hundreds of deformations at the seven locations denoted in Figure 1.2 with the purpose of calculating the stress state(s) at each of these locations. This process and the results are discussed in chapter 3.

Figure 3. Single deformation band in Entrada Sandstone offsetting cross-bedding by about 5 mm. Deformation bands form a resistant (typically) white rib in sandstones. (Photo taken from Aydin, 1977)

Cross-Bedding-Parallel Bands

For mechanical purposes, it is advantageous to think of the faulting at San Rafael Swell as belonging to one of two groups of faults: 1. Layer-parallel faults and 2. Faults cutting across layering. The layer-parallel faults formed as deformation bands along the cross-bedding in the Navajo Sandstone (the cross-bedding is actually the mechanical "layering" in the Navajo) and along bedding in the Wingate.

CONTRACTIONAL AND EXTENSIONAL FAULTS

Here the terms contractional and extensional are used with reference to the layering. Contractional faults accomplish thickening and shortening of the layer with the maximum compression direction

oriented roughly parallel to layering. Extensional faults accomplish thinning and lengthening of the layer with the maximum extension direction oriented roughly parallel to layering.

Small contractional faults formed in limestone beds of the lower part of the Carmel Formation. As shown in Figure 4, these contractional faults are usually small, accommodating around one centimeter of displacement, and form in conjugate pairs. Using Anderson's (1951) approach, the maximum compression direction, which bisects the angle between the conjugate set of faults, is sub-parallel to layering (Figure 4B). Thus, these contractional faults record layer parallel compression. Fischer (1997, 1998)

Figure 4. Small contraction faults in Carmel Formation. A. Short thick arrow points to exposed fault surface with white calcite slickenfibers. Long thin arrow points to fault trace on the exposed vertical face. Thin arrow shows reverse sense of displacement. Scale is ten centimeters long. B. Schematic sketch depicting conjugate contractional faults and layer-parallel maximum compression.

suggested that these contractional faults record early layer-parallel compression, before the limbs reached significant dip angles. Extensional faulting occurs in steeply dipping ($>45^\circ$) Moenkopi siltstone beds and Coconino Sandstone. In places the thinly bedded (<10 cm) siltstone beds of the Moenkopi Formation are the most highly deformed rocks at the San Rafael Swell. Near location 5 (Figure 1.2), where the Moenkopi beds dip generally $50-60^\circ$, the siltstone beds change in steeply dipping Coconino beds were "squeezed" normal to layering much like the overlying Moenkopi beds.

This "squeezing" of beds is expected in steeply dipping beds near the tip of the underlying fault in a fault-related fold. Accurate profiles of monoclines (e.g. Reches, 1978a), experimental models (e.g. Friedman et al., 1980), and theoretical models all show layer-normal thinning of the steeply dipping layers near the tip of the fault (Figure 8). This observation is indirect evidence of an

underlying fault tip somewhere near the base of the Moenkopi and the top of the Coconino Sandstone.

A.

B.

Figure 5. Listric extensional fault in Moenkopi Formation. A. View is to the north. Beds above fault surface dip $\sim 50^\circ$ to the east and beds below the fault dip as steeply as 80° to the east. B. Interpretation of photo. Arrows show layer-parallel extension and layer-normal compression.

Figure 6. Idealized listric normal fault.

A.

B.

Figure 7. Extensional faults in Coconino Sandstone. A. View to the west of the top of an eastward dipping bed of Coconino. The faults with a zigzag trace dip to the west such that the faulting extends the bedding. B. Interpretation of photo highlighting the faults and showing the relative sense of displacement. Notice that the top blocks moved "out of the page" in a reverse sense.

Figure 8. Theoretical model of a fault-tip fold (see chapter 4 for details). High layer-parallel extension occurs in steeply dipping beds near the tip of the fault (shaded in gray).

JOINTS

Layer-normal joints with various orientations formed within the Moenkopi, Navajo Sandstone, and Carmel Formation. As noted by Fischer (1997), cross-strike and strike-parallel joint sets indicate a complex history of strike-normal and strike-parallel compression during the formation of the San Rafael Swell. As suggested by Fischer (1997) these multiple joint sets may be related to lateral propagation during the growth of the monocline. I chose not to analyze joints because of the difficulty in interpreting multiple joint sets.

KINK FOLDS

Small-scale kink folds formed within the cross-bedding of steeply dipping Navajo Sandstone (Figure 9). Deformation bands formed along cross-bedding in the Navajo, and in places this slip is localized into narrow bands creating kink folds. The small folds are only about 10 cm wide. The kink folds are good indicators of stress state because the geometry of the folds can be used to estimate the direction of maximum compression responsible for the folds. The kink folds are discussed and analyzed in detail in chapter 5.

Figure 9. A. A kink band in the cross-bedding of Navajo Sandstone. B. Solid black lines show the cross-beds and the dashed lines show the walls of the kink band. The cross-beds are re-oriented within the walls of the kink band to create a monoclinial fold.

LOCALIZATION OF DELAMINATION OF BEDDING IN SAN RAFAEL SWELL, UTAH AND OTHER MONOCLINAL FOLDS

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Journal of Structural Geology, (2000) 22:1455-1468

ABSTRACT

Layers being bent by folding can become delaminated, so that a single layer in an early stage of folding becomes divided into many thinner layers in a later stage of folding. One can explain the delamination in terms of internal layer-parallel shear stresses that are generated by bending. An analysis of equilibrium of layer-parallel shear stresses and fiber normal stresses indicates that the maximum shear stress increases with the square of the thickness and the change in curvature, or coil, at middepth of the layer. The maximum shear stress is at middepth in the layer, so the layer will tend to fault into two layers of half the original thickness as it is bent. Using the method of computation of coil outlined here, we have been able to show that the coil predicts the location and sense of shear of delaminations in several field examples. The method illustrates how a few careful field measurements of layer geometry can provide valuable information about conditions of folding.

Key Words: Delamination, bedding-plane faults, coil, curvature, change of curvature, splitting of layers, monocline, kink band.

INTRODUCTION

In the course of a structural study of the monocline along the eastern margin of the San Rafael Swell, Utah, we noticed in some places, within the inner limb of the fold, that massive sandstone beds contain many layer-parallel faults whereas in other places they do not. Preliminary observations indicated that, in some way, the presence of the bedding-parallel faults within the inner limb correlates with the tightness of the fold. Where the fold is quite broad—radius of

curvature of 760 m—there are scant bedding-parallel faults. Where the fold is quite tight—radius of curvature of 40 m—bedding-parallel faults are abundant.

A.

B.

Fig. 1A. View of San Rafael Swell, Spotted Wolf section. I-70 is out of view to the right. White face of flatirons is Navajo Sandstone. B. Aerial view of southern end of Spotted Wolf Section. Sudden change in dip is evident with beds dipping steeply in the bottom of the picture and shallowly in the top.

Other researchers have observed slip surfaces parallel to bedding in steep limbs of folds, so our observations are not unique. Pfaff (1986) and Aytuna (1984) mapped bedding-parallel faults in kink bands ranging in width from a few meters to a few hundred meters in the central

Appalachians. Cook and Stearns (1975) mapped faults parallel to cross-bedding in a monoclinical fold at Dinosaur National Monument, Utah. Bevan (1984) noted the existence of slip planes parallel to bedding within the steep limb of the Isle of Wight monocline, England.

It appears, however, that no one has explained the formation of the bedding-parallel faults or why, for example, they should form in the limb rather than in the hinges of a monoclinical fold.

The formation of bedding-parallel faults is important to understanding of folding because the faults allow a stiff, single, thick layer to delaminate³¹ and, therefore, to bend easily; through delamination the layer is locally converted to a multilayer. As a result, the faults can cause localized folding of a layer at wavelengths much shorter than the thickness of the original layer.

The purposes of this paper are to describe some examples of delamination, explain why delamination occurs, identify parameters that affect delamination, and show how to make field measurements of the parameters.

FIELD EXAMPLES

Spotted Wolf Section of San Rafael Swell

The San Rafael Swell, located in central Utah, is an asymmetric, domical uplift stretching about 50 km east and west and about 120 km north and south. Beds over the center and on the west edge of the uplift dip gently westward. A magnificent monoclinical fold, with an inner limb dipping eastward, dominates the eastern edge of the Swell. The limb is characterized morphologically by dramatic flat-irons of Navajo Sandstone (Fig. 1). The Spotted Wolf section of the monocline extends 10 km from Spotted Wolf Canyon (at Interstate Highway 70) in the north to Iron Wash in the south.

The Spotted Wolf section was chosen for study because the limb has a wide range of dip angles and radii of curvature within the same rock units. The radius of curvature of the synclinal hinge ranges from 760 m, near I 70, to 40 m near midlength of the section. At both ends of the Spotted Wolf section, the fold hinges are wide open, the monocline is broad, and dips are relatively shallow, 30° or less (Fig. 1A). In the middle part of the section, the fold hinges are much tighter, the monocline is much narrower and the beds within the inner limb dip much more steeply and are locally overturned (Fig. 1B).

In places along the Spotted Wolf section small faults formed parallel to bedding in the Wingate Sandstone and parallel to cross-bedding in the Navajo Sandstone (Johnson, in prep). The Navajo Sandstone is a massive, cross-bedded sandstone. Bedding planes are difficult to locate, in part because beds can be as thick as several tens of meters, but large low-angle cross-beds give the Navajo a layered appearance. The Wingate Sandstone is also an intricately cross-bedded sandstone but it contains many beds as thin as about three meters.

³¹ Delamination is the local or general subdivision of a single mechanical layer into two or more mechanical layers. A single mechanical layer may contain many or no sedimentary layers. For example, a slab of granite bounded by joints is a mechanical layer. A varved clay could be a single mechanical layer if the strength of contacts is high or many mechanical layers if the strength of contacts between varves is low.

Fig. 2. Shikellamy Fold, Pennsylvania. Faults are mapped in detail in the lower picture. Right lateral bedding parallel faults are localized within the inner limb.

The frequency and occurrence of faults parallel to bedding or low-angle cross-bedding vary from place to place along the Spotted Wolf section and appear to correlate with tightness of the synclinal hinge and with thickness of bedding in sandstone units. The frequency of cross-bed-parallel faults where radius of curvature of the synclinal hinge is on the order of 50 m is much higher than where the radius of curvature of the synclinal hinge is on the order of 1000 m. For example, small cross-bed-parallel faults are scarce in large roadcuts where I70 provides nearly perfect exposures in the Navajo Sandstone at the northern end of the Spotted Wolf Section. In contrast, they are abundant in steeply dipping Navajo beds, near the synclinal hinge, where the fold is tight.

Small layer-parallel faults occur within thinly bedded Wingate Sandstone in the steep part of the Spotted Wolf section, however, they generally occur at bed contacts. They did not form parallel to cross-bedding within the Wingate.

Shikellamy Fold

The spacing of bedding-parallel faults was documented in a railroad cut in the Shikellamy fold (Fig. 2), a monoclinical fold in the Sherman Creek Member of the Catskill Formation, along the northwest bank of the Susquehanna River at Northumberland, Pennsylvania (Aytuna, 1984). A down-plunge cross section of the area shows that the Northumberland syncline is flat-bottomed (about one km wide) with open hinges on both sides. The Shikellamy fold is near the southern side of the flat bottom. Its inner limb is about 150 m long.

The Shikellamy fold is characterized by bedding-parallel faults that are sparse in the outer limbs and numerous in the inner limb. The spacing of bedding-parallel faults ranges from 0.25 to 0.7 m in sandstones and from 0.45 to 0.55 m in siltstones within the inner limb. In contrast, the spacing is 2 to 4 m in the southern outer limb and 25 m in the northern outer limb of the fold. As viewed down the fold axis (Fig. 2), the sense of shear parallel to bedding was right-lateral for faults in the steep inner limb and left-lateral for faults in the shallow outer limbs on the right and left sides.

Amity Hall Fold

Pfaff (1986) studied monoclinical folds in interbedded mudstone and siltstone in central Pennsylvania. One of them (Fig. 3) is exposed in a roadcut near Amity Hall, along highway 22/322

along the Juniata River, about 35 km northwest of Harrisburg, Pennsylvania. The rocks are part of the Trimmers Rock and Brallier formations, of Devonian age. The fold is shown via a sketch of traces of fractures; bedding surfaces were not mapped. Thus the spacing of the lines indicates the spacing of fractures. The sketch shows that the inner limb of the fold, which is about 2 m long, is more thoroughly delaminated than the two outer limbs (Fig. 3A). The number of fractures at least doubles in the inner limb.

The extra fractures in the inner limb extend only as far as the fold hinges, as shown in a detailed sketch (Fig. 3B) of the synclinal hinge marked with a rectangle in Fig. 3A. The sketch shows several bedding-parallel faults that accommodated about 1 to 2 cm of right-lateral slip and terminated near the hinges of the fold. The (mechanical) hinges separate right-lateral from left-lateral shearing on the upper and lower surfaces of the layer.

A.

B.

Fig. 3 (from Pfaff, 1986) A. Map of traces of fractures in Amity Hall fold. Localization of bedding parallel fractures in steep limb (Mapped by A.M. Johnson). B. Detail map of hinge marked with a box in A. (Mapped by V. Pfaff).

THEORETICAL ANALYSIS

The problems

The three monoclinical folds show well-developed delamination of the inner limbs. In two of the three field examples—the narrow, 2 m wide, Amity Hall monoclinical fold and the 150 m wide, Shikellamy fold—the folds show the formation of much more closely spaced bedding-parallel faults within the short inner limbs of the folds than in the long outer limbs on both sides. We could map the entire width of the inner limb in the example of the Amity Hall fold, and in this case we could see that the bedding-parallel faults extend essentially from hinge to hinge. In the much larger—1500 m width—monocline on the east side of San Rafael Swell we have mapped delaminated sandstone in the inner limb near the synclinal hinge.

Based on the field observations, we choose to ask the following questions about delamination of sedimentary rocks:

- How does a bent layer become delaminated by a layer-parallel fault into two layers? And, how do multiple layer-parallel faults form?
- Why did the massive Navajo Sandstone at San Rafael Swell delaminate but the bedded Wingate Sandstone did not?
- Why did the Navajo Sandstone delaminate near the tight synclinal hinge where the radius of curvature is 40 m but not near the broad synclinal hinge where the radius of curvature is 760 m?

Shear stress within a bent layer

The mechanism which we believe to be responsible for the process of delamination in bent rock layers has been known for many years in engineering mechanics (e.g., Marin and Sauer, 1954; Timoshenko and Goodier, 1961). In elementary mechanics, the bending of a layer is described solely in terms of a distinctive triangular distribution of normal stresses and strains parallel to the walls of the layer. For the positive curvature shown in Fig. 4A, the stresses are compressive at the top and tensile at the bottom and become zero at middepth (Fig. 4B). These stresses are known as the fiber stresses. We assume that they have the linear form,

$$\sigma_f = B \kappa n, \quad (1a)$$

in which B is an elasticity modulus

Fig. 4. A. Bent beam. B. Element of beam showing fiber stress distribution, layer-normal shear stress, $\tau(n)$, distribution, and contact stress, τ_{cont} , on top and bottom surfaces. C. Small piece of element with fiber stresses, σ_f , and layer normal shear stresses.

$$B = \frac{E}{(1 - \nu^2)} \quad (1b)$$

in which E is an Young's modulus, ν is Poisson's ratio, κ is curvature at middepth of the layer, and n is the radial distance from middepth. It is also well known that if one integrates the moments produced by fiber stresses with respect to middepth ($n=0$) one obtains the bending moment, M , for a beam or layer of rectangular cross section,

$$M = B\kappa \int_{-T/2}^{T/2} bn^2 dn = B\kappa I. \quad (1c)$$

where $I = bT^3/12$ is the moment of inertia. The equation for the bending moment relates the curvature, κ , of a layer to applied loads and moments. The equation for the bending moment, however, provides no understanding of the delamination of layers by layer-parallel faulting.

In more advanced engineering mechanics it is demonstrated that shear stresses are generated within a bent layer and that the shear stresses are caused by a change in bending moment (e.g., Marin and Sauer, 1954; Timoshenko and Goodier, 1961). We believe that these shear stresses are responsible for delamination of rock layers.

The shear-stress distribution within the layer depends on the boundary conditions as well as the change in fiber stresses along the layer. The top and bottom boundaries of the layer (Fig. 4B) are subjected to a contact shear stress, τ_{cont} . This stress is an interaction between the layer and its surroundings. For example, if slip between layers is left-lateral, the layers are subjected to left-lateral contact shear, τ_{cont} . If the layers slip freely past each other, τ_{cont} is zero. The element of the layer (Fig. 4B) is also subjected to shear stresses, $\tau(n)$, along its sides.

In order to derive an expression for the shear stress distribution through the element shown in Fig. 4B, we select the piece of the element shown in Fig. 4C that includes the upper surface of the layer

and a thickness δn of the element. Acting on each side of the piece are the linear fiber stress distributions and the unknown shear stress distributions, $\tau(n)$. Acting on the top is the contact stress, τ_{cont} , and on the bottom is $\tau(n_o)$. The fiber stresses change across the piece.

The fiber stresses at the left-hand end of the piece of the element are given by eq. (1a),

$$\sigma_\phi = B\kappa v,$$

whereas the fiber stresses at the right-hand of the piece are,

$$\sigma_f + \frac{\partial \sigma_f}{\partial s} \delta s = B \left[\kappa + \frac{d\kappa}{ds} \delta s \right] n. \quad (1d)$$

As indicated in (1d), the change in curvature, $\frac{d\kappa}{ds}$, across the element results in a change in the fiber stresses across the element. It is the change of fiber stresses that generates the shear within the layer.

We define $\tau(n_o)$ as the shear stress generated at the bottom of the piece of the element. Its relation to the change of fiber stresses, and curvature, is obtained by summing forces equal to zero in the horizontal direction,

$$[\tau_{\text{cont}} - \tau(n_o)] \delta s b - \frac{d\kappa}{ds} \delta s B b \int_{n_o}^{T/2} n dn = 0. \quad (2a)$$

Thus the shear stress along the base of the piece of the element is,

$$\tau(n_o) = \tau_{\text{cont}} - \frac{d\kappa}{ds} B \int_{n_o}^{T/2} n dn$$

$$\tau(n_o) = \tau_{\text{cont}} - \frac{d\kappa}{ds} (B/2) [(T/2)^2 - n_o^2] \quad (2b)$$

in which the sign ($\pm$) of τ_{cont} is determined from the slip direction of the beds (τ_{cont} is positive for right lateral slip). Eq. (2b) also gives the shear stress, $\tau(n_o)$, on the vertical sides of the piece, adjacent to the base (Fig. 4C).

Eq. (2b) indicates that the relation between shear stress, $\tau(n_o)$, and distance from middepth, n_o , is parabolic (Fig. 4B). Thus the shear stress on the base of the tiny element increases from the boundary value, τ_{cont} , at the top of the layer ($n_o = T/2$) to its maximum value, τ_{mid} , at midheight ($n_o = 0$) in the layer:

$$\tau_{\text{mid}} = \tau_{\text{cont}} - \frac{d\kappa}{ds} \frac{BT^2}{8}. \quad (2c)$$

Since eq. (2b) is in the form of a shear stress-strain relation, we write it in terms of the shear modulus, G , times a strain,

$$G = \frac{E}{2(1+\nu)}$$

so that

$$B = \frac{2G}{1-\nu} \quad (2d)$$

and eq. (2b) becomes

$$\frac{\tau(n_o)}{2G} = \pm \frac{\tau_{cont}}{2G} + (\epsilon_{sn})_{coil}, \quad (2e)$$

in which the shear strain due to coil is

$$(\epsilon_{sn})_{coil} = - \frac{dk}{ds} \left(\frac{T}{2} \right)^2 \frac{1}{2(1-\nu)} \left[1 - \left(\frac{2n_o}{T} \right)^2 \right]. \quad (2f)$$

If the material is incompressible, $\nu = 0.5$ so eq. (2f) simplifies to

$$(\epsilon_{sn})_{coil} = - \frac{dk}{ds} \left(\frac{T}{2} \right)^2 \left[1 - \left(\frac{2n_o}{T} \right)^2 \right]. \quad (2g)$$

Why a bent layer delaminates

The solution for the shear-stress distribution within a layer provides an answer to the first question, why a bent layer might delaminate. Suppose the contact stress, τ_{cont} , is negligible; that is, suppose that layers slip freely over one another. Then eq. (2c) indicates that the shear stress is proportional to the coil (change in curvature at middepth, $\frac{dk}{ds}$), to the elasticity modulus, B , and to the square of the thickness of the original layer, T^2 . If the coil is negative, the shear stress, τ_{mid} , is positive. If the coil is positive, the shear stress is negative.

The parabolic distribution indicates that the maximum shear stress occurs at middepth in the layer. If the shear strength parallel to bedding planes is constant, the first place the shear stress will equal the shear strength will be at middepth. A bedding-plane fault will form at middepth in the layer. Furthermore, the shear strength parallel to laminations tends to be less than the shear strength across laminations. Therefore if the laminations are parallel to the walls of the bed, the fault will have a strong tendency to propagate parallel to the layering at middepth. This is a reason sedimentary bedding tends to be split by bedding-parallel faults when it is bent.

Other things being equal, the stiffest, thickest layer in a stack of layers will tend to delaminate first as magnitude of the coil is increased. Suppose all the layers are of the same stiffness (same elastic modulus, B) but of different thicknesses, and the thickest layer has thickness T . Suppose the stack of layers is bent and the shear stresses parallel to layering increase until the shear stress at

middepth of the layer of thickness T reaches the shear strength. The shear stress will be less than the shear strength at other levels within the layer of thickness T and within other layers thinner than T . Thus the layer of thickness T will delaminate first. Now suppose all the layers are of the same thickness but of different stiffnesses. Then the stiffest layer will tend to delaminate first.

In either case, we now have two layers of thickness $T/2$. Each layer has a parabolic distribution of shear stress, but the maximum shear stress has been reduced by a factor of four. As the coil increases, the shear stress increases and each of these layers, in turn, becomes separated by a bedding-parallel fault. Now we have four layers of thickness $T/4$, and the layer originally of thickness T is a delaminated layer.

In order to explain the effect of contact strength we define the shear stress generated by the coil, τ_{coil} , such that

$$\tau_{coil} = -\frac{d\kappa}{ds} \frac{BT^2}{8} = -2G \left(\frac{d\kappa}{ds} \right) \left(\frac{T}{2} \right)^2 \frac{1}{2(1-\nu)}. \quad (3a)$$

Equation (2e) then can be written

$$\tau_{mid} = \tau_{cont} + \tau_{coil}. \quad (3b)$$

In this form we can visualize the effect of the contact stress, τ_{cont} , on intralayer stress at midheight, τ_{mid} . Whether the contact stress adds to or subtracts from the intralayer shear stress at midheight depends on the sense of interlayer slip. By definition, right-lateral shear stress is positive. Right-lateral slip, therefore, increases τ_{mid} when τ_{coil} is positive (negative change in curvature) and decreases τ_{mid} when τ_{coil} is negative (positive change in curvature).

Why greater delamination in Navajo Sandstone

It appears as though we can understand the greater delamination of the Navajo Sandstone as compared to the Wingate in terms of bedding thicknesses of these formations. The Wingate Sandstone contains many beds with thicknesses of a few meters. The Navajo Sandstone on the other hand, contains only a few beds with thicknesses of ~100 m, but has numerous cross-bedding surfaces.

If contact stress is neglected, the midheight shear stress, τ_{mid} , is proportional to the square of the thickness and directly proportional to the coil, eq. (2c). Assuming similar coils in the two rock units, the midheight shear stress in the massive Navajo Sandstone should be 400 to 10,000 times larger than the midheight shear stress in the bedded Wingate Sandstone. Thus there is a much higher tendency for layer-parallel faulting in the Navajo than in the Wingate.

Critical coil

For a bed to delaminate, the coil in curvature at midheight must be large enough for τ_{mid} to overcome the internal shear strength, τ_{int} , of the layer. Thus, there is a critical value for the coil at which delamination occurs. This critical value is obtained from eq. (2c) after substituting τ_{int} for τ_{mid} :

$$\left(\frac{d\kappa}{ds}\right)_{\text{crit}} = \frac{\tau_{\text{cont}} - \tau_{\text{int}}}{2G} \left(\frac{2}{T}\right)^2 2(1-\nu) . \quad (4a)$$

In dimensionless form, eq. (4a) can be written as

$$\frac{1}{2(1-\nu)} \left(\frac{d\kappa}{ds}\right)^*_{\text{crit}} = \frac{\tau_{\text{cont}} - \tau_{\text{int}}}{2G} , \quad (4b)$$

in which the nondimensional coil (or nondimensional change in curvature) is

$$\left(\frac{d\kappa}{ds}\right)^*_{\text{crit}} = \left(\frac{d\kappa}{ds}\right)_{\text{crit}} \left(\frac{T}{2}\right)^2 . \quad (4c)$$

For the special case of an incompressible material

$$\left(\frac{d\kappa}{ds}\right)^*_{\text{crit}} = \frac{\tau_{\text{cont}} - \tau_{\text{int}}}{2G} . \quad (4d)$$

The nondimensional coil is the negative of the shear strain at midheight in the layer (eq. (2g)).

The coil and the thickness of the layering being bent are clearly the most important geometric parameters that we need to measure in order to understand where delamination will occur. These parameters combine to form the nondimensional coil, eq. (4c), a quantity that we can compare from fold to fold at all scales.

METHODS OF MEASURING CURVATURE AND COIL

We need to be able to determine the sign and magnitude of the change in curvature along folded layers in order to determine where the layers will tend to delaminate.

Curvature and coil

In previous pages we have used the terms curvature and coil, but did not define what they are. Curvature,

$$\kappa = d\theta / ds , \quad (5a)$$

is a measure of the amount of bending of a curve at a point in a plane. We find that the radius of curvature,

$$\rho = 1 / \kappa , \quad (5b)$$

at a point has a more intuitive geometrical meaning than curvature. A circle can be constructed at a point along a curve with the same curvature and tangent as the curve (Fig. 6). The radius of the circle is called the radius of curvature, ρ , at the point.

Fig. 5. Curve in the xy -plane. A point on the curve has tangent vector T , horizontal vector H , and angle θ between them. Curvature is defined as the change in θ with the distance, s , along the curve, $d\theta/ds$.

Fig. 6. Radius of curvature, ρ , at a point on a curve is the radius of the circle that has the same tangent and curvature as the point on the curve.

An alternative expression for the curvature is derived in the Calculus in terms of Cartesian coordinates, x and y

$$\kappa(x) = \frac{y''(x)}{[1 + y'(x)^2]^{3/2}} \quad (5c)$$

Here $y''(x)$ is the second derivative and $y'(x)$ is the first derivative (slope) of the curve at a point.

The derivative of curvature with respect to distance, s , at a point is defined as the coil at that point,

$$\text{coil} \equiv \frac{d\kappa}{ds} = \frac{d^2\theta}{ds^2} \quad (6a)$$

The derivative, ds , is always taken as positive along the curve s in the counterclockwise direction.

Fig. 7. A. Coil at point 2 can be visualized with change in radius of curvature from point 3 to 1. B. The coil along a circle is zero since $\rho_1 = \rho_2$ for any two points 1 and 2 on a circle.

The coil can be visualized with two nearby radii of curvature. Figure 7A shows a curve with an open bend that becomes progressively tighter with distance along the curve. The open bend has a radius of curvature ρ_1 at point 1, and the tight bend a distance δs away has a much smaller radius of curvature ρ_3 at point 3. In terms of radius of curvature, the coil given by eq. (6a) at point 2 becomes

$$coil \equiv \left[\frac{d^2\theta}{ds^2} \right]_2 = \left[\frac{d}{ds} \left(\frac{1}{\rho} \right) \right]_2 \approx \frac{\frac{1}{\rho_3} - \frac{1}{\rho_1}}{s_3 - s_1} = \frac{-1}{\rho_3 \rho_1} \left(\frac{\rho_3 - \rho_1}{s_3 - s_1} \right). \quad (6b)$$

An extreme value is obtained from a circle (Figure 7B). It has zero coil everywhere because

$$\frac{1}{\rho_2} - \frac{1}{\rho_1} = 0$$

everywhere. The coil of a straight line is also zero because a straight line is the extreme of a circle with an infinite radius of curvature.

A. positive coil

B. negative coil

Fig. 8. Negative and positive coil. A. Positive Archimedes' coil has positive coil at every point. B. Negative Archimedes' spiral has negative coil at every point.

The sign of the coil at a point is determined by eq. (6b) or with the aid of Fig. 8. Figure 8 shows two Archimedes' spirals, one with positive coil at every point on the curve and the other with negative coil at every point on the curve. While traversing a curve in a clockwise direction, if at a point the curvature changes from larger to smaller, the coil is negative, as is the case at every point on the negative spiral in Fig. 8B. While traversing a curve in a counterclockwise direction, if at a point the curvature changes from smaller to larger, the coil is positive, as is the case at every point on the positive spiral in Fig. 8A. In the example of Fig. 7A, the coil is positive at point 2 since the curvature changes from smaller at point 1, to larger at point 3 as we traverse the curve in a counterclockwise direction.

Measurement of curvature and coil with field measurements

We will develop a method for calculating curvature and coil along a contact in a cylindrical fold using field measurements. The method requires five dip measurements along a line normal to the strike of the cylindrical fold, and measurements of the distance between each of these points, measured along the fold (Fig. 9). The curvature, $d\theta/ds$, is calculated at points (s_2, θ_2) , (s_3, θ_3) , and (s_4, θ_4) , and then the coil is calculated at point (s_3, θ_3) with these $d\theta/ds$ values.

Fig. 9. Hypothetical profile of a fold and five points spaced unevenly along the fold. Coil can be approximated at (s_3, θ_3) with eqs. (4.7).

The method presented here (after Nakamura, 1996) allows for uneven spacing of the five points. We want to allow for uneven spacing of points because the field data may not be available everywhere on the fold. We first outline the procedure delineated in eqs. (7) before deriving the steps.

The angle, θ , is a function of the directed distance, s , along the curve. The curvature, $d\theta/ds$, is calculated at a point (s_i, θ_i) using the three points (s_{i-1}, θ_{i-1}) , (s_i, θ_i) , and (s_{i+1}, θ_{i+1}) . The derivative, $d\theta/ds$, is assumed to have the form

$$\left(\frac{d\theta}{ds}\right)_i = a_i \theta(s_i) + a_{i-1} \theta(s_{i-1}) + a_{i+1} \theta(s_{i+1}), \quad (7a)$$

in which the coefficients are the solution to the equations,

$$\begin{aligned} a_i + a_{i-1} + a_{i+1} &= 0 \\ 0 + a_{i-1} \{s_{i-1} - s_i\} + a_{i+1} \{s_{i+1} - s_i\} &= 1, \\ 0 + a_{i-1} \{s_{i-1} - s_i\}^2 + a_{i+1} \{s_{i+1} - s_i\}^2 &= 0 \end{aligned} \quad (7b)$$

or,

$$\begin{bmatrix} a_i \\ a_{i-1} \\ a_{i+1} \end{bmatrix} = \begin{bmatrix} \frac{(s_{i-1} - s_i)^2 - (s_{i+1} - s_i)^2}{(s_{i+1} - s_i)(s_{i-1} - s_i)(s_{i+1} - s_{i-1})} \\ \frac{s_{i+1} - s_i}{(s_{i-1} - s_i)(s_{i+1} - s_{i-1})} \\ \frac{-(s_{i-1} - s_i)}{(s_{i+1} - s_i)(s_{i+1} - s_{i-1})} \end{bmatrix}. \quad (7c)$$

The coil, $(d^2\theta/ds^2)_i$, is determined at (s_3, θ_3) using the curvature values calculated at (s_2, θ_2) , (s_3, θ_3) , and (s_4, θ_4) . The coil is calculated as a derivative of first derivatives of the form:

$$\left(\frac{d^2\theta}{ds^2}\right)_3 = a_3 \left(\frac{d\theta}{ds}\right)_3 + a_2 \left(\frac{d\theta}{ds}\right)_2 + a_4 \left(\frac{d\theta}{ds}\right)_4. \quad (7d)$$

Here the coefficients a_2 , a_3 , and a_4 are the same as in eqs. (4.7c) with $i=3$ since the spacing is the same.

Now we will explain how eqs. (7) are derived. The method deals with unequally spaced points by expressing the functions $\theta(s)$ and $d\theta/ds$ in eqs. (7a) and (7d) as a Taylor expansion. This can be illustrated with eq. (7a); eq. (7d) is expanded in the same way.

Recognizing that a function, $\theta(s)$, can be written as a Taylor expansion about a point s_i ,

$$\theta(s) = \sum_{n=0}^{\infty} \frac{\theta^{(n)}(s_i)(s-s_i)^n}{n!},$$

eq. (4.7a) can be expressed:

$$\begin{aligned} \left.\frac{d\theta}{ds}\right|_i &= a_i \theta(s_i) + a_{i-1} \left[\theta(s_i) + \theta'(s_i)\{s_{i-1} - s_i\} + \frac{\theta''(0)}{2!}\{s_{i-1} - s_i\}^2 + \frac{\theta'''(0)}{3!}\{s_{i-1} - s_i\}^3 + \dots \right] \\ &+ a_{i+1} \left[\theta(s_i) + \theta'(s_i)\{s_{i+1} - s_i\} + \frac{\theta''(0)}{2!}\{s_{i+1} - s_i\}^2 + \frac{\theta'''(0)}{3!}\{s_{i+1} - s_i\}^3 + \dots \right] \end{aligned} \quad (8a)$$

After reorganizing terms and dropping third order terms and higher,

$$\begin{aligned} \left.\frac{d\theta}{ds}\right|_0 &\approx \theta(s_i)[a_i + a_{i-1} + a_{i+1}] + \theta'(s_i)[a_{i-1}\{s_{i-1} - s_i\} + a_{i+1}\{s_{i+1} - s_i\}] + \\ &\frac{\theta''(s_i)}{2!}[a_{i-1}\{s_{i-1} - s_i\}^2 + a_{i+1}\{s_{i+1} - s_i\}^2] \end{aligned} \quad (8b)$$

To solve for the coefficients, we set the coefficients of $\theta(s_i)$, $\theta'(s_i)$, and $\theta''(s_i)$ in eq. (8b) to 0, 1, 0 respectively:

$$\begin{aligned}
 a_i + a_{i-1} + a_{i+1} &= 0 \\
 0 + a_{i-1} \{s_{i-1} - s_i\} + a_{i+1} \{s_{i+1} - s_i\} &= 1. \\
 0 + a_{i-1} \{s_{i-1} - s_i\}^2 + \{s_{i+1} - s_i\}^2 &= 0
 \end{aligned}
 \tag{8c}$$

Thus we have derived equations equivalent to eq. (7b).

The error of the solution can be estimated with the higher-order term that is not used to determine the coefficients, $\frac{\theta'''(0)}{3!} [a_{i-1}s_{i-1}^3 + a_{i+1}s_{i+1}^3]$. Because we neglect terms with third and higher derivatives, the result is a third-order approximation.

Coil measurements on a profile: Palisades monocline

As an example of measurements of coil on a structure profile, we select a strike view of the Palisades monocline, Grand Canyon, Colorado, constructed by Reches (1978). The Palisades monocline is a branch of the East Kaibab monocline near the confluence of the Colorado and Little Colorado rivers. The canyon of Palisades Creek is about 1200 m deep and offers an unusual opportunity to examine the actual geometry within a monocline, based on first-hand observations. The upper part of the canyon was mapped by a combination of photogrammetry and field observation, and the lower part was mapped with plane-table methods, so the map, and the derivative strike view (Fig. 10) are highly accurate. The result is a clear view of a monocline, from the fault zone in Precambrian, basement rocks to the Permian, Coconino Sandstone and Kaibab Limestone at the canyon rim (Fig. 10).

The strike view shows that the Palisades monocline is asymmetric. There is a near-vertical fault in Precambrian rocks below. The upper and lower contacts of the Bright Angel shale have slightly different forms. The lower contact is adjacent to the basement fault and the beds are steep to overturned there. The upper contact is about 100 m above the end of the basement fault.

The bottom of the Redwall limestone has a distinct synclinal hinge but no anticlinal hinge; it is probably outside the view to the right. The top of the Redwall is broadly folded.

We can describe the shapes of the contacts between formations in terms of coil. Figure 10A displays normalized coil, expressed units of 10^{-2} , calculated at locations along four contacts. A thickness of $T = 200$ m was used to obtain the normalized coil.

The largest values for coil in the Palisades monocline occur adjacent to sharp hinges (Fig. 10A). The contact at the base of the Bright Angel Shale (Cambrian) has relatively sharp synclinal and anticlinal hinges. The coil is large adjacent to both hinges. The contact between the top of the Bright Angel shale and the bottom of the Muav Limestone (Cambrian) has a sharp synclinal hinge and an open anticlinal hinge. The coil is large only adjacent to the sharp synclinal hinge. The coil is small and highly localized near a contraction fault in the contact at the base of the Redwall Limestone (Mississippian) and the top of the Temple Butte Limestone (Cambrian).

The coil decreases rapidly and the fold broadens upward, away from the fault. The broad part of the fold does not have sharp hinges.

The reason the coil is large near narrow hinges in folds and relatively straight limbs can be understood in terms of change of radius of curvature (Fig. 10B). Radius of curvature is large on each limb and small in the hinge. A small radius of curvature in the tight hinge changes abruptly to a large radius of curvature in the adjacent straight limb.

A. Coil

B. Radius of Curvature

Fig. 10. Accurate composite cross section of the Palisades Monocline. A. Coil was calculated at each of the indicated locations. The largest changes in curvature at each interface are located adjacent to the sharp hinges. B. Radius of curvature at the locations in A.

The locations of the largest coils correspond with locations of largest τ_{coil} . In the case that τ_{cont} is zero or has the same sign as τ_{coil} , the locations of largest coil correspond to the locations of largest τ_{mid} and correspondingly, locations where delamination is most likely to occur. The shaded parts of Fig. 10A show locations of large coil. The arrows show sense of shear. Of course, coil does not completely determine whether or not a bed delaminates, as delamination depends also on the thickness of the bed and the shear strength of the rock. We can, however, determine directly from these measurements where thick beds are most likely to delaminate in a fold and where they likely will not delaminate.

The bedding-parallel faults were not mapped, however, when the monocline was studied so we cannot check the predictions with this example. Rather, we use this example to indicate that the regions of high coil and thus higher layer-parallel shearing are near the hinges in this particular monocline.

LOCATION OF LAYER-PARALLEL FAULTS IN FIELD EXAMPLES

Amity Hall monoclinial fold

Layer-parallel faults in the Amity Hall Fold (Fig. 3) are in both the short inner and long outer limbs. The faults within the inner limb are about 2 m long. The shape of the fold is characteristic of a monoclinial kink band (Reches and Johnson, 1976). It has relatively sharp hinges, with radii of curvature of about 60 cm, and relatively straight limbs, with larger radii of curvature of three or four meters.

There is a relatively high concentration of layer-parallel faults in the inner limb, adjacent to the sharp hinges (Figure 3A). There are fewer faults near mid-length of the inner limb.

Fig. 11. Calculations of radius of curvature and coil at ten points along a bedding contact in the Amity Hall fold. The largest coil is located adjacent to the sharp hinges.

In this example we know the spacing of faults and we can measure coil, so we should be able to understand why the bedding-parallel faults are distributed as they are. We measured the coil at ten locations along bedding-parallel faults in the fold (Fig. 11). We can see that the location of the high concentration of faults adjacent to the hinges corresponds with the locations of largest magnitude of coil (-1.6 and $+1.6 \text{ m}^{-2}$ that is, per square meter", near the anticlinal hinge and -1.9 m^{-2} at the synclinal hinge). Normalizing coil using $T = 0.2 \text{ m}$, the normalized coils are -1.6 and $+1.6$ near the anticlinal hinge and -1.9 at the synclinal hinge. The coil adjacent to the sharp hinges is about 5 times larger than the coil in the center of the inner limb.

These results suggest that we might expect a concentration of layer-parallel faults on the two outer limbs and the inner limb, adjacent to the hinges. We observe that the faults are concentrated, though, only on the inner limb. The lack of faults, however, on the outer limbs can be understood if the Amity Hall fold did indeed form as a kink band and we consider the effect of the contact stress, τ_{cont} .

To obtain the kink band form, it is necessary that there be frictional or cohesive contact strength between layers. This is so because, for an isolated kink band to form (without a train of folds), it is necessary at the edge of the band that the bending moment, M , be zero where there is no bending. Honea and Johnson (1976) have shown that, for this to occur, the contact stress, τ_{cont} , eq. (2b), must be some non-zero critical value. Consider now eqs. (3) and, in the case of our view of the Amity Fold, a positive value for τ_{cont} applied to the layers. Then the effect of τ_{cont} on the midheight layer-parallel shear stresses, τ_{mid} , is to increase τ_{mid} in the inner limb where the coil is negative (τ_{coil} is positive), and decrease τ_{mid} on the outer limbs where the coil is positive (τ_{coil} is negative). It is therefore likely that an effect of the contact stress was to reduce the intralayer shear stresses on the outer limbs to the extent of preventing layer-parallel faults from forming there, that is, preventing delamination.

San Rafael Swell monocline

The question of why cross-bedding faults are abundant in the Navajo Sandstone at San Rafael Swell near the tight fold hinges, whereas they are scarce near the broad fold hinges, can be addressed with the result in eqs. (2c) and (3b). First, we assume that the shear stress is solely a result of change in bending. Then eq. (2c) shows that the shear stresses at midheight of the layer, when $\tau_{\text{cont}} = 0$, are proportional to the coil, $d\kappa/ds$, of the bent layer. The coil near the tight synclinal hinge, where cross-bed-parallel faults were observed, is larger than near the synclinal hinge of the broader part of the fold, where cross-bed-parallel faults were not observed.

Fig. 12. Cross section of steep section of San Rafael Monocline. Radius of curvature of the synclinal hinge was measured in the field near the bottom of the Entrada Sandstone. Radius of curvature in the upper anticlinal hinge was measured from a topographic map. Dark shaded section is exposed in the field. The lighter shaded section is inferred.

The radius of curvature was measured in several places within the Carmel Formation, which overlies the Navajo Sandstone, and in the Entrada Sandstone, which overlies the Carmel. At a location where the synclinal hinge of the Entrada Sandstone is tight (Fig. 12), the radius of curvature in the hinge is $\rho = 40 \text{ m}$ ($\kappa = 25 \times 10^{-3} \text{ m}^{-1}$). At a location within the Carmel Formation where the synclinal hinge is broad, the radius of curvature is $\rho = 760 \text{ m}$ ($\kappa = 1.3 \times 10^{-3} \text{ m}^{-1}$). The difference is more than an order of magnitude.

We can roughly estimate the coil if we assume the curvature is zero on the inner limb of the fold and both of the measured hinges are the same distance, Δs (measured in meters), from the location of zero curvature, then the coil adjacent to the broad hinge is

$$\frac{d\kappa}{ds} = \frac{\frac{1}{760} - 0}{\Delta s} = \frac{1}{760\Delta s} (\text{m}^{-2})$$

and the coil adjacent to the tight hinge is

$$\frac{d\kappa}{ds} \approx \frac{\frac{1}{40} - 0}{\Delta s} = \frac{1}{40\Delta s} (\text{m}^{-2}).$$

Therefore the coil is nearly 20 times larger at the location with the tighter hinge. Assuming the bed thickness at both locations is the same (200 m for Navajo Sandstone), the midheight shear stress within the Navajo at the location with tighter hinge would have been nearly 20 times the midheight shear stress at the other location (eq. (2c)). Furthermore, if the spacing of the two measurements of curvature are an arc distance, $\Delta s = 500 \text{ m}$ apart, the normalized coil in the tighter hinge is about 0.5 whereas that in the open hinge is about 0.026. Thus the normalized coil in the open syncline at

San Rafael Swell is about the same magnitude as the normalized coil in the Amity Hall fold. The normalized coil in the tight syncline at San Rafael Swell is about the same magnitude as the larger values obtained in the Palisades monocline.

DISCUSSION AND CONCLUSIONS

Analysis of beam and plate theory shows that layer-parallel shear stresses within a bent layer should be maximal at middepth of the layer and should increase with coil. Assuming that layer-parallel faulting is caused by the high shear stress, we have a viable explanation for the delamination of relatively thick layers. We presented several examples where delamination has occurred in monoclinical folds and showed that coil is maximal in these locations. In this sense, coil predicts sites of delamination in a layer.

Although the concept of coil is rather abstract, it can be understood in terms of familiar geometric concepts. The curvature, κ , is defined as the change in slope angle, θ , with a change in arc distance along a line, s , such that

$$\kappa = \frac{\delta\theta}{\delta s} = \frac{1}{\rho},$$

in which ρ is the radius of curvature. Accordingly, the coil at point 2, between points 1 and 3, can be written in terms of the curvatures ρ_1 and ρ_3 at the two adjacent points. The coil at a point can thus be calculated by measuring the curvature at two adjacent points a distance s_3-s_1 apart:

$$\left[\frac{d^2\theta}{ds^2} \right]_2 \approx \frac{-1}{\rho_3\rho_1} \left(\frac{\rho_3-\rho_1}{s_3-s_1} \right)$$

We have presented two methods for calculating change in curvature, or coil, with field data. In one method, one measures altitudes and positions (e.g., x , y and z) of five points with equal spacing (7 for unequal spacing) along a line of curvature within a bedding surface. The coil at the middle point can be computed with these data. Alternatively, one can calculate coil along a line of curvature if one measures apparent dips at three points equally spaced (5 points unequally spaced along the line), including the altitudes and positions (x , y and z) of the points along a line of curvature. The general measurement techniques for unequally spaced points are explained in the Appendix.

Problems that we have addressed inadequately are:

- Why do the faults form parallel to cross-beds and not parallel to bedding in the Navajo Sandstone?
- Why are there many more cross-bed-parallel faults in the Navajo Sandstone than in the Wingate Sandstone?

The planes of weakness in the Navajo Sandstone apparently are largely restricted to large cross-beds whereas those in the Wingate include cross-beds and bedding contacts. Since the maximum shear stress occurs on surfaces parallel to bedding, the Wingate tends to fault at bedding contacts; there are few cross-bed-parallel faults. In the Navajo, the maximum shear stress at midheight, τ_{mid} , is parallel to bedding, but the surfaces of weakness are cross-beds, at acute angles to bedding. The component of shear stress resolved onto the cross-beds (τ_{x-bed} , Fig. 13) is smaller

than the shear stress parallel to bedding, but the shear strength parallel to cross-beds is apparently lower than that parallel to bedding.

Fig. 13. A cross-bedded sandstone bed. When folded, the maximum shear stress (τ_{mid}) acts parallel to bedding at middepth of the layer. While the shear stresses on the inclined cross-bed surfaces are less than the maximum shear stress ($\tau_{x-bed} < \tau_{mid}$), the shear strength is apparently less on cross-bed surfaces than on planes parallel to bedding. Faulting is observed parallel to cross-beds, not parallel to bedding.

Fig. 14. Hypothetical profile of a fold and five points spaced unevenly along the fold. Coil can be approximated at x_4 with eqs. (8).

In this way the existence of cross-bedding faults in the thick beds of the Navajo Sandstone and the lack of them in the more thinly-bedded Wingate Sandstone can be explained in terms of eq. (2c).

In the case of the Wingate, there are sufficient, weak bedding-parallel surfaces so that the shear stresses do not become large enough to fault along cross-bedding in these units.

In a companion paper (Johnson and Johnson, in prep.) we predict locations of delamination in a variety of theoretical monoclinical folds, including layers overlying an idealized laccolithic intrusion, in kink folds and in drape folds. The pattern of delamination and coil should provide a way of determining how a fold was formed.

ACKNOWLEDGEMENTS

This research was made possible by the generous efforts of Kaj's field assistant and wife, Cyndi Johnson, and financial support for field work by the Donald Levandowski memorial award the Cedric J. Newby award, Department of Earth and Atmospheric Sciences, Purdue. The remainder of the field expenses were paid by Department of Energy grant DE-FG02-98ER14886, supervised by Nick Woodward of DOE. We appreciate the support of these organizations.

APPENDIX

The general ideas behind approximating the coil at a point were explained in the text and are summarized in eqs. (7). A problem with using the central difference approximation in eqs. (7b) and (7c) is that the approximations are based on the assumption that measurement points are spaced evenly with equal incremental arc distances of Δs along the curve. This assumption often cannot be satisfied in the field or on maps because the data may not be available everywhere. For this reason we use a different method to perform the calculations including those reported in the text.

In this Appendix we develop a method using seven points which is a slightly more complicated, but a much more powerful, numerical method which allows for unequal spacing of points (after Nakamura, 1996). We will first present the results of the method as eqs. (8), before deriving the equations.

- The angle θ_i is calculated at the five points, x_2 through x_6 , as shown in Fig. 14 with the approximation

$$\theta_i \approx \tan^{-1} \left(\frac{dy}{dx} \right)_i. \quad (8a)$$

We approximate the derivative dy/dx to third-order accuracy with a method that accounts for unequal spacing of points. To calculate the derivative at (x_i, y_i) we use the three points (x_{i-1}, y_{i-1}) , (x_i, y_i) , and (x_{i+1}, y_{i+1}) . The derivative is assumed to have the form

$$(y')_i = a_i y_i + a_{i-1} y_{i-1} + a_{i+1} y_{i+1}, \quad (8b)$$

in which the three coefficients are determined by solving the three equations,

$$\begin{aligned}
a_i + a_{i-1} + a_{i+1} &= 0 \\
0 + a_{i-1}x_{i-1} + a_{i+1}x_{i+1} &= 1 \\
0 + a_{i-1}x_{i-1}^2 + a_{i+1}x_{i+1}^2 &= 0
\end{aligned} \tag{8c}$$

The solution is

$$\begin{bmatrix} a_i \\ a_{i-1} \\ a_{i+1} \end{bmatrix} = \begin{bmatrix} \frac{x_{i-1}^2 - x_{i+1}^2}{x_{i+1}x_{i-1}(x_{i+1} - x_{i-1})} \\ \frac{x_{i+1}}{x_{i-1}(x_{i+1} - x_{i-1})} \\ \frac{-x_{i-1}}{x_{i+1}(x_{i+1} - x_{i-1})} \end{bmatrix} \tag{8d}$$

These results are substituted into eq. (8b) and the resulting equation is substituted into eq. (8a). In this way θ_i is determined at each of the five points.

- The curvature, $(d\theta/ds)_i$, is calculated at x_3, x_4 , and x_5 .

Because, however, we are taking a derivative with respect to s , we are no longer working in the x - y coordinates, but instead in (s_i, θ_i) coordinates where s_i is the distance along the curve from the origin to (x_i, y_i) . Again, to account for unequally spaced points along the curve, we calculate the derivative $(d\theta/ds)_i$ in a similar way to $(dy/dx)_i$. The derivative has the form

$$\left(\frac{d\theta}{ds} \right)_i = b_i \theta_i + b_{i-1} \theta_{i-1} + b_{i+1} \theta_{i+1}, \tag{8e}$$

in which the coefficients are the solution to the equations,

$$\begin{aligned}
b_i + b_{i-1} + b_{i+1} &= 0 \\
0 + b_{i-1}s_{i-1} + b_{i+1}s_{i+1} &= 1, \\
0 + b_{i-1}s_{i-1}^2 + b_{i+1}s_{i+1}^2 &= 0
\end{aligned} \tag{8f}$$

or,

$$\begin{bmatrix} b_i \\ b_{i-1} \\ b_{i+1} \end{bmatrix} = \begin{bmatrix} \frac{s_{i-1}^2 - s_{i+1}^2}{s_{i+1}s_{i-1}(s_{i+1} - s_{i-1})} \\ \frac{s_{i+1}}{s_{i-1}(s_{i+1} - s_{i-1})} \\ \frac{-s_{i-1}}{s_{i+1}(s_{i+1} - s_{i-1})} \end{bmatrix} \tag{8g}$$

Lastly, the coil, $(d^2\theta/ds^2)_i$, is determined at (s_4, θ_4) using the curvature values calculated at (s_3, θ_3) , (s_4, θ_4) , and (s_5, θ_5) . The curvature is calculated as a derivative of first derivatives of the form:

$$\left(\frac{d^2\theta}{ds^2}\right)_4 = b_4\left(\frac{d\theta}{ds}\right)_4 + b_3\left(\frac{d\theta}{ds}\right)_3 + b_5\left(\frac{d\theta}{ds}\right)_5. \quad (8h)$$

Here the coefficients b_3 , b_4 , and b_5 are the same as in eqs. (8g) with $i=4$ since the spacing is the same.

Now we will explain how eqs. (8) are derived. The method deals with unequally spaced points by expressing the functions $y(x)$, $\theta(s)$, and $d\theta/ds$ in eqs. (8b), (8e), and (8h) as a Taylor expansion. This can be illustrated with eq. (8b); the other equations are expanded in the same way.

It is convenient first to shift the coordinates in the x direction so that the three points are centered about $x=0$. Then the three x coordinates become $\chi_i \rightarrow 0$, $\chi_{i-1} \rightarrow \chi_{i-1} - \chi_i$, and $\chi_{i+1} \rightarrow \chi_{i+1} - \chi_i$. Using a Taylor expansion, eq. (8a) can be expressed:

$$\begin{aligned} y'(0) = & a_i y(0) + a_{i-1} \left[y(0) + y'(0)\chi_{i-1} + \frac{y''(0)}{2!}\chi_{i-1}^2 + \frac{y'''(0)}{3!}\chi_{i-1}^3 + \dots \right] \\ & + a_{i+1} \left[y(0) + y'(0)\chi_{i+1} + \frac{y''(0)}{2!}\chi_{i+1}^2 + \frac{y'''(0)}{3!}\chi_{i+1}^3 + \dots \right] \end{aligned} \quad (9a)$$

After reorganizing terms,

$$\begin{aligned} y'(0) = & y(0)[a_i + a_{i-1} + a_{i+1}] + y'(0)[a_{i-1}\chi_{i-1} + a_{i+1}\chi_{i+1}] + \frac{y''(0)}{2!}[a_{i-1}\chi_{i-1}^2 + a_{i+1}\chi_{i+1}^2] \\ & + \frac{y'''(0)}{3!}[a_{i-1}\chi_{i-1}^3 + a_{i+1}\chi_{i+1}^3] + \dots \end{aligned} \quad (9b)$$

To solve for the coefficients, we set the coefficients of $y(0)$, $y'(0)$, and $y''(0)$ in eq. (9b) to 0, 1, 0 respectively:

$$\begin{aligned} a_i + a_{i-1} + a_{i+1} &= 0 \\ 0 + a_{i-1}\chi_{i-1} + a_{i+1}\chi_{i+1} &= 1. \\ 0 + a_{i-1}\chi_{i-1}^2 + a_{i+1}\chi_{i+1}^2 &= 0 \end{aligned} \quad (9c)$$

Thus we have derived equations equivalent to eq. (8c). In the same way, using s and θ , eqs. (8f) for the b_i -coefficients is derived.

The error of the solution can be estimated with the higher-order term that is not used to determine the coefficients, $\frac{y'''(0)}{3!}[a_{i-1}\chi_{i-1}^3 + a_{i+1}\chi_{i+1}^3]$. Because we neglect terms with third and higher derivatives, the result is a third-order approximation.

ANALYSIS OF MULTIPLE SETS OF BAND FAULTS IN POROUS SANDSTONE, SAN RAFAEL MONOCLINE, UTAH

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ABSTRACT

Heterogeneous populations of small band faults at San Rafael Swell, Utah, can be sorted into three distinct stress events using measurements of orientation and net slip on faults, offset relations observed in the field, and a stress inversion method. The timing of the stress events relative to uplift of the Swell was approximated using orientations of bedding at the time of faulting, together with crosscutting relations. Stress events occurring before, during and after uplift of the structure can be identified with the method. The stress event that occurred during uplift has a maximum compression direction that is 20° off the normal to the general strike of the monoclinical fold forming the eastern flank of San Rafael Swell.

INTRODUCTION

As part of my field study of the Spotted Wolf section of the San Rafael monocline on the eastern side of San Rafael Swell, Utah (Johnson, 1999; Johnson and Johnson, in prep. A), I measured the kinematics of many small faults in Navajo Sandstone: strike and dip of a fault surface, direction and net amount of lateral shift across the fault. This paper presents the stress-inversion method used, the results of analyses of the faults at seven sites along the San Rafael Monocline (Fig. 2), as well as the method used to determine timing of the faulting relative to uplift of the San Rafael Swell.

I developed a stress-inversion method in order to analyze the field measurements and to reconstruct the paleostress history of my field area. I use a stress-inversion technique to sort the heterogeneous fault populations into multiple stress events, then I deduce the stress history of the faulting with the concurrent use of relative age observed in the field and the stress-inversion method.

Previous researchers have used small scale structures such as faults to estimate the stress state associated with Laramide uplifts of the Colorado Plateau and Rocky Mountain foreland (Reches, 1978a; Anderson and Barnhard, 1986; Varga, 1993; Davis and Tindall, 1996).

Perhaps the greatest difficulty in reconstructing the paleostress history of the faults at San Rafael Swell is the fact that the faults are on the flank of the large uplift where the beds range from horizontal to dip angles up to 60°. Analysis therefore requires not only determination of the stress state(s) at the time of faulting, but the orientation of the beds at the time of faulting. It is conceivable that faulting occurred over a long time period, from before uplift of the structure when the beds were horizontal, through formation of the structure when the beds were being tilted, and after uplift when the beds were in their present position.

The study area is on the eastern side of the San Rafael Swell, a major Laramide uplift of the Colorado Plateau (Kelley and Clinton, 1960) that is believed to have formed initially in latest Campanian time (late Cretaceous) and into early Tertiary (Lawton, 1983). The Swell is an

asymmetric, domical uplift stretching ~50 km east and west and ~120 km north and south. Beds over the center and on the west edge of the uplift dip gently westward. The eastern edge of the Swell is dominated by a magnificent monoclinial fold where the beds dip steeply eastward, and are even overturned locally, in the central limb of the monocline. The steeply dipping central limb is characterized by dramatic flat-irons of Navajo Sandstone (Fig. 1). The eastern limb of the monocline dips gently to the east.

A.

B.

Figure 1

The Spotted Wolf section of the monocline extends 10 km from Spotted Wolf Canyon (I 70) in the north to Iron Wash in the south (Figure 2). This section was chosen for study because it includes a wide range of dip angles in the monoclinial limb, within the same rock units. The monocline at locations 1 and 7 (Figure 2) is relatively broad, where the central limb dips 30° or less. At locations 3 through 6 the fold is much tighter and, in places, the beds are overturned. The radius of curvature of the synclinal hinge is about 760 m at location 1, near I 70, whereas it is 40 m at location 4, near midlength of the Spotted Wolf section, where beds dip 60° .

Figure 2.

Geologic setting of Spotted Wolf Section of San Rafael Swell, Utah.

The central limb of the monocline is wonderfully exposed and is therefore an ideal location to study the internal deformation of the central limb of a monocline. Deep, transverse, narrow canyons run perpendicular to strike, cutting through much of the steeply dipping beds of the Jurassic and Triassic section. Furthermore, a longitudinal valley, running parallel to strike through the easily weathered Moenkopi siltstones and between the Permian, White Rim Sandstone and Triassic sandstones, provides great exposure along strike of the central limb. The lower synclinal hinge in Jurassic, Entrada and Carmel sandstones is exposed in cross section within washes a few meters deep in the Spotted Wolf section. The anticlinal hinge and upper limb are preserved in the White Rim (Coconino) Sandstone (Figure 1).

BAND FAULTS

Much of the faulting at the San Rafael Swell is in the form of deformation bands³², which are very thin, tabular bodies of highly localized cataclasis in porous sandstones³³. Small amounts of

³² Also known as Atilla faults, micro faults, seal faults, granulation seams. See description by Aydin (1978).

displacement (mm) are accommodated across single bands. Larger displacements (1 to 10 cm) can occur along zones where many bands form side by side (Aydin and Johnson, 1978). Deformation bands have been studied in detail by several investigators (Aydin, 1977, 1978; Aydin and Johnson, 1978, 1983; Jamison and Stearns, 1982; Underhill and Woodcock, 1987; Cruikshank and others, 1991; Zhao and Johnson, 1991; Antonelli and others, 1994).

Deformation bands were mapped at various locations along the Spotted Wolf section in the Entrada, Navajo and Wingate sandstones. The Entrada Sandstone is a relatively thick-bedded, medium-grained, cross-bedded sandstone. Cross-bed thicknesses range from five meters to a half meter. The Navajo Sandstone is very thick-bedded to massive but in other ways similar to the Entrada Sandstone. Bedding planes are difficult to locate in the Navajo, in part because the beds can be as thick as several tens of meters, but large low-angle forsets give the Navajo a layered appearance. The Wingate Sandstone, like the Navajo, is an intricately cross-bedded sandstone with bed thicknesses of several meters.

Two distinct groups of band faults were identified within the sandstones. One group formed parallel to cross-bedding surfaces in the Navajo Sandstone (Figure 3). The second group of faults cut cross-beds and formed in numerous orientations. The cross-bed-parallel faults occur in the Navajo Sandstone at locations 3 through 6 in Fig. 2, where the beds range in dip from 45° to 60°. The faults parallel to cross beds almost never (there was one exception) offset the faults that cut cross beds, suggesting that the cross-bed-parallel faults formed earlier than the second group of faults.

Jamison and Stearns (1982) have also documented deformation bands parallel to cross-beds in the Wingate Sandstone at Colorado National Monument. They too observed the same age relationship with cross-bed faults offset by faults not parallel to cross-beds and deduced that the cross-bed-parallel faults were active only in the early stages of deformation of the Wingate Sandstone.

In this paper only faults in the second group, which offset cross-beds, will be analyzed. The cross-bed parallel faults of the first group do not offset cross-beds or other faults, so it was impossible to determine the slip directions necessary for the stress inversion. Also, we have shown (Johnson and Johnson, in prep A, in prep B) that the faults parallel to cross-bedding at San Rafael Swell most likely formed in response to bending of the strata during uplift of the Swell. Thus they would be excluded from an analysis of the regional stress state.

³³ Jamison and Stearns (1982) have also documented deformation bands parallel to cross-beds in the Wingate Sandstone at Colorado National Monument. They too observed the same age relationship with cross-bed faults offset by faults not parallel to cross-beds and deduced that the cross-bed-parallel faults were active only in the early stages of deformation of the Wingate Sandstone.

Figure 3. White lineations are deformation bands in Navajo Sandstone along cross-beds at location 5 (see Figure 2). Lowest band is along bottom of upper unit. Other bands are along broad, foreset cross-bedding. The band faulting is clearly localized along the cross-stratification. Lens cap on left for scale.

Seven locations (numbered 1 through 7 in Fig. 2) along the synclinal hinge of the Spotted Wolf section of San Rafael Swell were selected for mapping deformation bands. A small outcrop of deformation bands was chosen for study at each location (largest outcrop is ~20 m by 20 m). Figure 4 is a photo and block diagram of part of the outcrop at location 4, Fig 2.

The white, raised ridges in the rock in Fig. 4A are traces of deformation bands. There are two general orientations of faults at this location (Fig. 4B). Slickenlines are not present on most fault surfaces so the slip direction was obtained by measuring two independent apparent displacements on offset surfaces (cross beds or other deformation bands). The net slip, direction, and sense of slip can be obtained with the apparent displacement measurements. Only the direction and sense of slip are used in the stress-tensor-inversion method.

A.

1 m

B.

N80W

1 m

Figure 4. A. Photo of typical outcrop of deformation bands. The deformation bands are the thin resistant ridges that form white lineations. B. Block diagram of 4A.

METHOD OF STRESS INVERSION

The fact that multiple sets of faults can form, even in an isotropic medium, under a single stress state has been recognized at least since the experiments by Oertel (1965) on clay cakes subjected to three-dimensional deformations. Until that time, faults were generally considered to be two-dimensional structures (e.g., Anderson, 1942). Aydin (1977, 1978) recognized and described multiple sets of deformation bands and zones of deformation bands in Entrada Sandstone in the San Rafael Desert, Utah (Aydin and Reches 1982). Reches (1978b) described multiple fault sets in a monocline in the Grand Canyon and showed (Reches, 1978a) that multiple sets are required to accommodate three-dimensional strain.

In the case of newly-formed faults in a isotropic material the principal stress directions can be inferred directly from orientations and slip on regular sets of fault planes (e.g. Anderson, 1942; Johnson, 1994). There are two sets if the strain is two-dimensional and four sets if the strain is three-dimensional (Oertel, 1965; Reches, 1978a; Johnson, 1994).

The stress-inversion method developed here is independent of the assumption of isotropy—faults or other fractures may pre-exist, there may be bedding or cross-bedding planes of weakness—but it does depend upon upon certain assumptions that I will enumerate here and discuss and incorporate in the theory in the Appendix:

- The stress state is constant throughout the time the faults slipped to produce the measured net slip.
- Slip on one fault does not affect the slip on another fault. (The faults do not interact.)
- The stress field does not change with position in the vicinity of the faults.
- The fault surfaces are planar where slip data are obtained.
- The slip vector is parallel to the maximum shear traction on the fault plane.
- The slip vector and the maximum shear traction point in the same direction, that is, the slip has the same sense as the shear traction.
- Slip on a pre-existing fault or on a newly-formed fault follows Coulomb's friction law

$$|\tau| = C + \sigma_n \tan \phi . \quad (1)$$

where τ is maximum shear traction and σ_n is normal traction (compression is positive) on the fault, and ϕ is the friction angle. For a sandstone, ϕ is generally 30 to 40°.

I have made a fresh derivation of the stress-inversion method in order to determine the stress state corresponding to a group of faults that slipped under the same stress state, but stress-inversion methods are now widely used in structural geology. The inversion method was apparently first used to determine stress states from multiple faults by Carey and Brunier (1974). Carey and Brunier adopted the first five assumptions listed above. They computed only a deviatoric stress tensor, that is, the stress tensor minus the mean normal stress; furthermore, they compute the components of deviatoric stress normalized by a stress of unknown magnitude, such as the lithostatic stress. Many other researchers have modified the Carey and Brunier (1974) method to devise quantitative, computer-aided analyses to reconstruct the local tectonic stress tensor from fault orientations and slip directions (Etchecopar et al., 1981; Angelier et. al., 1982; Angelier,

1984; Gephart and Forsyth, 1984; Gephart, 1990). None of them made assumptions 6 and 7, that slippage is according to Coulomb's law and the senses of traction and slip are the same.

Reches (1987) have made independent, fresh derivations of the stress-inversion method. The derivations are rather different from those based on the work by Carey and Bruhier (1974), but they are similar to each other. The general approaches by Reches and me are very similar. Reches (1987) derived stress-inversion equations based on assumptions 1 through 5, as did Carey and Bruhier (1974), but he added assumption 7. He did not use assumption 6 and he did not use senses of slip to compute the stress tensor. Reches included Coulomb's yield criterion and normalized with one of the principal stresses. In this way he was able to determine not only directions of principal stresses but normalized values of principal stresses. After determining the normalized principal stresses, he computes the maximum resolved shear stress on each fault plane. If the direction or sense of slip for a fault deviate markedly from the direction of maximum shear stress the data for this fault is eliminated and the stress tensor is redetermined.

The method I have developed explicitly uses senses of slip on faults to obtain the normalized stress tensor. The method follows all seven assumptions listed above. The stress components are normalized with the minimum principal stress. The method depends on accurate measurements of attitudes of fault planes locally as well as net slip on those faults. I do not know how to use the measurements of magnitude of slip although I assume that the slip vector and traction vector point in the same direction.

The stress inversion method is described in the Appendix. It can be implemented with a MATLAB® program to which I have attached a user-friendly interface. Currently the program requires the MATLAB® programming environment, but it will soon be available as a stand-alone product for Microsoft WINDOWS 95® and 98® environments.

The aim of all of the various stress-inversion methods that have been devised over the years is to reconstruct the stress tensor, using orientations of fault planes and orientations of slickenlines or other indicators of direction of slip on the faults. The principal stress directions can be determined from the stress tensor.

A least-squares method is used to find the best fit stress tensor for a population of faults (see Appendix). The method determines the stress tensor that best fits the observed slips on all the faults. Then, for each fault, the observed slip (labeled "os" in Fig. 5) is compared to the direction of maximum shear stress on the fault plane, or calculated slip direction (labeled "cs" in Fig. 5). The angle between the observed and calculated slip directions, the misfit angle, serves as a measure of the quality of the solution for each fault and the average misfit angle serves as a measure of the quality of the solution for the collection of faults analyzed.

Figure 5.

MISFIT OF SOLUTIONS

When calculating the stress state for a group of faults, the misfit may be attributed to deviations from any of the seven assumptions used in the method. Assumptions 3 and 4 appear to be valid for our study, as the deformation bands are locally planar features, and faults were mapped in small areas (20 m by 20 m) where regional variations in the stress field should not be a problem. Assumption 2, that the faults do not interact, is made for convenience. Pollard et al., (1993) have analyzed effects of fault shape and proximity to traction free boundaries and have concluded that the slip directions of faults can significantly deviate only as a result of fault interaction. They reported that the slip direction on a fault can deviate (up to about 20°) from the direction of the maximum resolved, far-field shear stress. I did not study fault interaction in my field sites, so I have no information with which to assess interaction. However, a solution is considered satisfactory if all of the misfit angles are less than 30° .

I will assume in this study that, if there is large misfit, assumption 1, that the faults slipped under a single stress state, is violated. Thus large misfits between the observed and calculated slip directions are due to multiple stress states. It is this assumption that is the basis of our method of sorting faults according to stress state. In the process of presenting the results of our fault analysis at San Rafael Swell, the general method used to sort fault populations according to stress state will be discussed.

ANALYSIS OF STRESS STATES IN FOLDED BEDS—SORTING OF FAULTS

Figure 6 displays the results at each of the seven locations along the San Rafael monocline. At each location an equal area lower hemisphere projection of the faults, rakes³⁴, and principal stress directions is provided, as well as the stress ellipsoid. The orientations of the axes of the stress ellipsoid give the orientation of the principal stresses. The shape of the ellipsoid provides the relative magnitudes (absolute values) of the principal stresses.

³⁴ Rake angles are defined as in Fig. 5. Oblique slip faults with right-lateral and reverse (normal) components of slip have rakes between 0 and 90° (0 and -90°). Oblique slip faults with left-lateral and reverse (normal) components of slip have rakes between 90 and 180° (-90 and -180°).

Figure 6. Stereonet plots and stress ellipsoids of all stress-states determined at each of the seven locations shown in Fig. 3.4. Stereonet plots are equal area, lower hemisphere projections. Poles to the fault planes as well as the rakes are plotted for each fault. The orientations of the axes of the stress ellipsoids give the orientation of the principal stresses. The shape of the ellipsoid provides the relative magnitudes (absolute values) of the principal stresses, labeled σ_1 , σ_2 , σ_3 ($\sigma_1 \geq \sigma_2 \geq \sigma_3$).

Figure 7. Directions of maximum compression and timing relative to uplift at the seven locations in the Spotted Wolf section of San Rafael Swell.

Figure 7 displays the maximum compression directions at the seven locations along the Spotted Wolf section. It was possible to determine the timing of the faulting relative to uplift of the San Rafael Swell at sites one through six. Figure 6 displays the results of the stress inversion at each of the seven locations along the Spotted Wolf Section of the San Rafael monocline. An equal angle,

lower hemisphere projection of the faults, rakes³⁵, and principal stress directions is provided, as well as the stress ellipsoid. The orientations of the axes of the stress ellipsoid give the orientation of the principal stresses. The shape of the ellipsoid provides the relative magnitudes (absolute values) of the principal stresses.

As can be seen in Fig. 2 the faults are in beds which have a wide range of dip angles. The beds at site 7 at the southern end of the study area dip 10° or less and the beds at site 5 dip 60° . This presents a special complication to the fault analysis because a decision must be made as to the position of the beds at the time of faulting. It is of course possible that faulting occurred before, throughout, and after uplift of the Swell. If the position of the beds at the time of faulting can be estimated, then the timing of the faulting relative to uplift of the Swell may be determined. The method of determining the position of the bedding at the time faulting is best illustrated using fault data at site 6, Fig. 2.

Figure 8 Sketch map of traces of faults at site 6 where the fracture pattern is very simple and the beds dip at about 50° . Whereas slip data could be collected on only thirteen of the seventeen faults, all of the fractures visible at this location were mapped. There are two sets of nearly vertical faults trending about N50W and N80W. The sense of the strike-slip component of the net slip is shown with arrows.

Figure 8 is a sketch map of traces of faults at site 6 where the fracture pattern is very simple and the beds dip about 50° . Upon initial inspection, the faults shown in Figure 8 appear to be a classical example of two sets of high-angle conjugate faults (2-dimensional faulting). If this were the case, Anderson's theory of faulting presumably could be applied to obtain a compression direction that bisects the acute angle between the two sets. The faults, however, are not purely strike-slip faults; some of the faults have rake angles as large as 40° (140° for left-lateral) (see Table 1).

³⁵Rake angles are defined as in Fig. 3.10. Oblique slip faults with right-lateral and reverse (normal) components of slip have rakes between 0 and 90° (0 and -90°). Oblique slip faults with left-lateral and reverse (normal) components of slip have rakes between 90 and 180° (-90 and -180°).

Table 1. Fault slip data for site 6 (Fig. 7)

Fault #	Orientation	Rake angle	Degrees rotated
1	N75W 89SW	-140	0
2	N66W 89SW	-178	0
3	N58W 68NE	-20	10
4	N52W 84NE	-37	0
5	N70W 89SW	165	0
6	N45W 75NE	-30	0
7	N70W 89NE	160	0
8	N48W 77NE	-26	30
9	N44W 75NE	-32	30
10	N60W 89NE	171	0
11	N62W 55NE	-10	30

Results were initially computed for 13 faults in their present position and orientation showed a large misfit angle of 74°. I assumed that some of the faults had slipped early and had subsequently been rotated. Two of the faults, though, were inconsistent with the others, regardless of orientation, so they were set aside.

Of the remaining 11 faults, four were assumed, on the basis of rake angle, to have formed earlier, when the beds were dipping 20 to 40° (Table 1). These angles suggests that the faults at site 6 formed as the layering was oriented at 20° then 40° and finally 50°. The orientations of the fault planes and slip vectors of the four earlier-formed faults were obtained by rotating the planes and slip vectors about the local strike of bedding. The stress state was then obtained using the data from seven non-rotated faults plus the data from the four rotated faults.

Location 6 in Table 2 summarizes the measurements and results. The dip of bedding (=DIP) is about 50°. Thirteen (=NM) faults were mapped. Eleven faults were used (=NU) for the analysis, the friction angle ($=\phi$) was assumed to be 30°, seven faults were not rotated (=R0), one fault was rotated 10° (=R10) and three faults were rotated 30° (=R30). The angles of rotation, the attitudes of the faults and the rake angles are presented in Table 1.

The interpretation that the faults slipped when the beds had different orientations is supported by the fact that a single stress tensor with a small mean misfit of 14° was calculated for the 11 faults. Furthermore, offsets of faults indicate that the faults that were rotated occurred before the faults that were not rotated in the analysis. According to this analysis, fault numbers 3, 8, 9, and 11 in Fig. 7 would have formed earlier than the other faults, which is consistent with the offsetting relationships that can be observed in Fig. 7.

The solution obtained for site 6 suggests that the faults formed under a constant stress state while the beds were reorienting during uplift of the San Rafael Swell.

This rationale for rotating bedding was also applied in the analysis of sites 1-1, 2, 4, 5, and 6 and the results are summarized in Table 2.

SITE	DIP	NM	NU	MM
1-1	20	30	7	10°
1-2	20	30	10	15°
1-3	20	30	11	11°
2	20	13	9	18°
3	50	18	10	13°
4	30	9	7	14°
5	60	14	9	14°
6	50	13	11	14°
7	10	14	7	6°

Table 2. Summary of stress tensor inversion data for the sites in Fig. 6*

**The locations of the numbered sites are in Fig. 6. DIP is the dip of the beds at the location (strike is N20-30E at each location); NM is the number of faults mapped; NU is the number of faults used in the inversion; MM is the mean misfit angle; ϕ is the internal friction angle; R0 is the number of faults rotated back with the bedding 0° from present orientation; R10 is the number of faults rotated back with the bedding 10° from present orientation; etc.*

DEDUCING MULTIPLE STRESS EVENTS

As explained in earlier paragraphs, I assume that large misfit angles (greater than 30°) for individual faults in their present position, are inconsistent with a single stress state or slipped when the beds had different orientations. Thus large misfit angles are attributed to heterogeneity in the fault population.

Inversion of all the faults at each of the sites resulted in large misfit angles, suggesting multiple stress events are responsible for the faulting. As can be seen in Table 2, it was impossible at any of the sites to fit a single stress tensor to all of the fault data, so an attempt was made to sort the faults at each site into multiple stress events.

This is not the first attempt to sort heterogeneous fault populations into distinct stress events or to reconstruct the paleostress history using numerical stress inversion methods. The technique of calculating the best fit stress tensor for fault populations is widely used in recent literature in tectonic studies (e.g. Vittori et al., 1994; Henry, 1998). Etchecopar et al. (1981) and Angelier

(1984) have developed algorithms designed to iteratively separate several superimposed tectonic events for heterogeneous fault populations, resulting from multiple stress states. Nemcok and Lisle (1995) and Lisle and Vandycke (1996) have also developed a statistical stress-inversion technique based on cluster analysis to sort fault-slip data into coherent sets.

Figure 9

At site 1 (Figure 9, Table 3) in San Rafael monocline, three distinct stress events can be identified. Offsetting relationships from field maps were used together with the stress inversion method to sort faults into the three events. The observations of offsetting relationships were limited, but turned out to be essential in the process of analyzing the faults at site 1.

Table 3. Maximum compression directions

Site	σ_1 bearing	Degs. from perp.	Timing relative to uplift
1-1	N45W	25 (N)	during
2	N55W	15 (N)	during
4	N70W	0	during
5	N40W	30 (N)	during
6	N53W	17 (N)	during

7	N45W	25 (N)	ambiguous
1-2	S75W	35 (S)	early
1-3	S67W	43 (S)	late
3	S75W	35 (S)	late

In Fig. 9 a block diagram has been constructed from a field sketch of a vertical face exposing a group of high angle faults at site 1. The sketch shows the sense of strike-slip component of movement with arrows and the sense of dip slip movement, if any, with offsets. Faults 1 through 8, in their present, non-rotated, position have large components of strike slip movement as the rake angles are all between 0 and 30° for right-lateral faults and between 160 and 180° for left-lateral faults. Fault 9 has a rake angle of 60°. Figure 8 was used to qualitatively sort all thirty faults at site 1 (including those not pictured) into two groups; one group of faults, including 1,2,4, and 6, that appear to be consistent with a generally NW/SE trending compression direction, and a second group, including 3,5,7 and 8 that appear to be consistent with a compression direction generally trending SW/NE. In this simplest scenario, the SW/NE compression would have occurred before the NW/SE compression as fault 5 is offset by fault 6. All of the faults at site 1 were subsequently placed into one of the two groups so that an inversion could be computed to determine the stress tensors. The fault planes and slip vectors in the SW/NE compression group were initially rotated 20° (the local dip) about the local strike of bedding, while the faults in the NW/SE compression group were initially left in their present position. Large misfit angles were obtained, however, for both groups. To obtain a satisfactory stress tensor solution (misfit angles <30°) for the NW/SE compression group (site 1-1, Fig. 6) it was necessary to rotate the fault planes and slip vectors of some of the faults about the local strike of bedding (see Table 2). To obtain a satisfactory solution for the SW/NE group it was necessary to sort the faults into two groups (sites 1-2, and 1-3). A satisfactory solution was obtained for the faults at site 1-2 with the beds rotated to horizontal, while it was necessary to leave the faults in their present position to obtain a satisfactory solution for the faults at site 1-3. Any observable offsetting relationships were used to constrain these results.

Table 4. Fault slip data for site 1

Orientation	Rake	Degrees rotated
Group 1-1		
N45W 89NE	0	0
N85W 89SW	177	0
N35W 89SW	20	0
N24W 70NE	3	20
N80E 86SE	-173	10
N90W 70S	-171	20
N80W 86SW	-165	0
Group 1-2		
N82W 36NE	-19	20
N57E 56SE	-143	20
N29E 12SE	151	20
N42E 43SE	-170	20
N11E 38SE	155	20

N67E 70NW	-20	20
N90E 60S	9	20
N85E 70S	-18	20
N8E 41SE	135	20
N3W 24NE	160	20
Group 1-3		
N35E 72NW	-28	0
N55E 84NW	23	0
N25W 60NE	180	0
N40E 89SE	-7	0
N20E 89SE	11	0
Group 1-4		
N70W 89SW	-60	0
N60W 72SW	-75	0
N70W 86SW	-103	0
N15W 64NE	-54	0
N40E 63SE	-93	0

We therefore have a local stress history that accounts for 28 of the 30 measured faults at site 1, and using the orientation of the bedding at the time of faulting, we can determine the timing of the stress events relative to the uplift of the San Rafael Swell. According to our analysis, the faults in group 1-2 formed before or in the early stages of uplift, when the beds were horizontal. Since some of the faults in group 1-1 were rotated with the bedding and others were left in their present position, the stress event recorded at site 1-1 must have occurred during uplift, when the beds were being tilted. The stress event at site 1-3 must have occurred during the late stages of uplift or after the cessation of uplift, when the beds were in their present position.

Table 5. Summary of stress inversion data for site 1

Group	DIP	NM	MM	R0	R10	R20	R30
1-1	20	7	10	30	4	1	2
1-2	20	10	15	30			10
1-3	20	5	5	30	5		
1-4	20	5	11	30	5		

RESULTS OF THE ANALYSES OF BAND FAULTS IN SAN RAFAEL MONOCLINE

All of the results summarized in Table 2 and Fig. 6 were obtained using the techniques described above. With these techniques it was possible to identify one stress event at six locations (sites 2 through 7), and three stress events at one location (site 1). Furthermore, it is possible to approximate the timing of a stress event relative to the uplift of the Swell using the position of the bedding at the time of faulting.

There is evidence to suggest the faulting at sites 1-1, 2, 4, 5, and 6 occurred during uplift of the San Rafael Swell. At these five sites (marked with a gray box in Fig. 6), a single stress state consistent with many of the faults at each site was found only after rotating some of the faults back to an earlier position. As discussed above, the faulting at sites 1-3 and 3 apparently occurred late or after uplift and the faulting at site 1-2 occurred early or before uplift. Timing of faulting relative to uplift is ambiguous at site 7 as the beds here dip only 10° or less.

The direction of maximum compression (σ_1) is remarkably consistent (NW trend) at each location marked with a gray box in Fig. 6 where the faulting apparently occurred during uplift. The stress states at sites 1-2 and 1-3, which apparently occurred before or after uplift, are clearly different than those marked with a gray box. The N45W/S45E compression direction that is recorded during uplift of the structure may be the regional compression responsible for uplift of the structure. Interestingly, while the bearings of the maximum compression directions are fairly consistent, they are as much as 30° off of the normal to the general strike of the monocline which is N20E (Table 3). Furthermore, the compression direction at sites of early or late faulting is 15-20° off the normal to the monocline.

This observation of deviation of maximum compression from the normal to the monocline is not unique to the San Rafael Swell. Reches (1978) obtained a compression direction that is roughly 25° off the normal to the general trend of the Palisades Monocline, Grand Canyon using small folds and faults and calcite twinning. He analyzed small faults in limestones in the Hazera and Hathira monoclines in the northern Negev, Israel, and determined that the maximum compression direction deviated from the normals to the axes of the monoclines by about 45° (Reches, 1976). Anderson and Barnhard (1986) used a stress inversion technique on faults in the Cocks Comb monocline, Utah, to obtain a maximum compression direction that is 20° or more off the normal to that monocline.

Table 6. Maximum compression directions

Site	bearing	angle (°) from perpendicular	timing relative to uplift
1-2	S75W	40 (S)	early
1-1	N45W	2 (N)	concurrent
2	N55W	10 (N)	concurrent
4	N70W	5	concurrent
5	N40W	25 (N)	concurrent
6	N53W	12 (N)	concurrent
1-3	N15E	80 (N)	late
1-4	vertical	--	late
3-1	S72W	38 (S)	late
3-2	N62W	3 (N)	late
7	N45W	20 (N)	ambiguous

Anderson and Barnhard (1986) suggested that a pre-existing basement fault, rather than the compression direction, might have controlled the orientation of the fold. Davis (1978) discusses thoroughly the role of pre-existing basement faults in determining the orientation of structures on the Colorado Plateau. The assumption of pre-existing basement faults is in accordance mapping by Walcott (1890) and Huntoon (1974) that demonstrated that the major fault zones underlying monoclines in the Grand Canyon area are Precambrian in age and were reactivated during the Laramide orogeny.

The results presented in this paper suggest that the orientation of the San Rafael Swell may also be controlled by a pre-existing weakness in the basement.

IMPLICATIONS TO LARAMIDE TECTONICS

The result of compression sub-normal to the trend of the San Rafael is consistent with stress-inversion results for other uplifts on the Colorado Plateau and in the Rocky Mountain foreland of northern Colorado and southern Wyoming (Reches, 1978; Anderson and Barnhard, 1986; Wise and Obi, 1992; Narr, 1993; Varga, 1993; Davis and Tindall, 1996; Bump, 1999; Davis, 1999). Figure 3.7A is a compilation map of the results of the stress inversions from eight Laramide uplifts. The maximum compression directions determined from faults on the eight uplifts are all roughly normal to the trends of the uplifts.

Coney (1976) proposed, based on plate reconstructions, that the Laramide uplifts of the Colorado Plateau and Rocky Mountains foreland formed in response to northeast-southwest compression resulting from collision of North America with oceanic plates to the west. Coney suggested that as North America was moving southwesterly away from Europe, collision with oceanic plates induced northeast-southwest compressive stresses that were transmitted into the thermally weakened backarc area. Based largely on this report, many authors have suggested that the Laramide uplifts formed in response to maximum compression that was oriented northeast/southwest throughout the Laramide orogeny (e.g. Sales, 1968; Stone, 1969; Brown, 1988; Blackstone, 1990). Other authors have considered the varying orientations of the Laramide uplifts and have suggested that the maximum compression direction rotated throughout the Laramide Orogeny (e.g. Chapin and Cather, 1983; Gries, 1990; Bergerat et al., 1992).

If the maximum compression directions in Fig. 3.7, determined from stress inversion of fault data, are regional compressions, then this data appear to support the argument that the maximum compression direction rotated throughout the Laramide Orogeny; the stress states calculated at San Rafael Swell, Owl Creek Mountains, and Casper Mountain are not consistent with northeast-southwest directed maximum compression. However, if one wants to deduce regional tectonic information from this fault data then one must first be able to answer a very important question: did the uplifts form in response to the stress state calculated from the fault data, or did the stress states form in response to the formation of the uplifts?

A.

Figure 10. Maximum compression direction obtained from inversion of fault data from eight Laramide uplifts in the Rocky Mountain foreland and Colorado Plateau. Maximum compression directions are consistently normal to the trend of the individual structures, suggesting the stress states are local and induced by the formation of the uplifts. A. Compiled using data from Kelley (1956), Reches (1978),

It is likely that the stress states calculated at these uplifts are not regional stress states at all, but rather local stress states induced by the formation of the uplifts. This can be understood mechanically by looking at the stress state in a fold that forms passively over a basement fault. In this case I mean by the expression, "passively forming over a fault", that the strata overlying a basement fault are not loaded with a far-field stress, and thus all of the stress within the overlying material is induced by folding over the displaced basement. The simplest mechanical model of a drape fold is shown in Figure 11. This figure shows the stress state at points in a viscous half space that is "draped" over a displaced rigid "basement". This model was created by applying a velocity boundary condition at the overburden/basement interface to create the effect of normal fault. The overlying viscous material is a halfspace, thus it is not layered; the lines drawn to represent layers are actually passive markers. As seen in Figure 11, the passive

Figure 11. Stress state at points in a viscous half space that is "draped" over a displaced rigid "basement". The displacement of the basement is created by applying a velocity boundary condition at the overburden/basement interface to create the effect of normal fault. The drawn "layers" are actually passive markers.

deformation of overburden over a vertical normal fault generates sub-horizontal maximum compression within the steep "limb" of the deformed passive markers near the fault, and in the

“synclinal hinge” and “lower limb” of the passive markers farther above the fault. This region of sub-horizontal maximum compression is shaded in gray.

A theoretical model of a drape fold with actual layers is shown in Figure 12, published by Reches and Johnson (1978). The monocline in Figure 12A was created by applying vertical displacements at the base of the stack of elastic multilayers. This model simulates stratigraphic layers passively folded over a vertical fault in a rigid basement. Figure 12B shows the orientations of the axis of maximum shortening. Similar to the results for the viscous half space, the layers in the vicinity of the central limb and synclinal hinge are subjected to sub-horizontal shortening.

The stress states calculated at San Rafael Swell were calculated within the steep central limb, lower synclinal hinge, and lower limb of the fold, which are areas that correspond with the shaded region in Figure 11 and the horizontal compression in Figure 12. The faults used to calculate the stress states are highly localized within these areas of the Spotted Wolf section. Very little internal deformation occurred within the strata in the exposed parts of the upper and lower limbs of the monocline. This observation, together with the theoretical illustrations in Figure 11 and Figure 12 suggest that sub-horizontal maximum compression in the overburden at San Rafael Swell and other uplifts could be a local stress state formed in response to the draping of layers over a fault. Therefore it is not logical to assume that the stresses in Fig. 3.7 are regional stresses.

Figure 12. Forced folding (“drape folding”) of a multilayer composed of five layers, with bonded contacts, subjected to localized vertical displacements at its base. $G_1, \dots, G_5$ indicate shear moduli. Stippled layers are the stiff layers. A. Displacements of contacts between layers. Displacements are exaggerated by 3 orders of magnitude. B. Orientations of the axis of maximum shortening in the passively folded multilayer in the undeformed configuration. From Reches and Johnson (1978, Figure 6). See Reches and Johnson for details.

CONCLUSIONS

A new derivation of the stress-inversion method for multiple faults that slipped during a single stress state incorporates most of the information required to describe a fault: The attitude of the fault and the direction and sense of net slip. The new derivation (presented in Appendix) provides a tool for analyzing heterogeneous faults, such as the deformation bands at San Rafael Swell, Utah. The deformation bands were successfully sorted into three distinct stress events using a

combination of the stress-inversion method and the crosscutting relationships mapped in the field. Furthermore, the timing of the stress events relative to uplift of the Swell was approximated using orientations of bedding at the time of faulting, together with crosscutting relationships.

A stress event with maximum compression oriented N70E/S70W appears to have caused faulting either in the early stages of folding or before uplift of the Swell. At five sites along the study area, a stress event with maximum compression directions clustered around N50W/S50E appears to have occurred during uplift of the structure. The compression during last event was in the same direction as the compression during the first event (N70E/S70W), apparently during late folding stages or after uplift.

The orientations of the maximum compression directions are not normal to the trend of the monocline, but rather are generally 30° and as much as 43° off the normal. This result is similar to findings at the Palisades and Cocks Comb Monoclines of the Colorado Plateau. In both cases the researchers determined the maximum compression direction to be 20-25° off the normal to the monoclines.

ACKNOWLEDGEMENTS

This research was made possible by the generous efforts of my field assistant and wife, Cyndi Johnson, and financial support for field work by the Donald Levandowski memorial award the Cedric J. Newby award, Department of Earth and Atmospheric Sciences, Purdue. The remainder of the field expenses were paid by Department of Energy grant DE-FG02-98ER14886, supervised by Nick Woodward of DOE. I appreciate the support of these organizations.

APPENDIX: THEORY OF THE STRESS-INVERSION METHOD

The complete theory of the stress inversion method used in this paper is provided here. Two coordinate systems are first defined as illustrated in Fig. 9. The global coordinates (x_1, x_2, x_3) are chosen to correspond with the compass directions. The local coordinate system ($\bar{x}_1, \bar{x}_2, \bar{x}_3$) is chosen, for convenience, with $\bar{x}_1$ and $\bar{x}_2$ in the fault plane, with $\bar{x}_1$ parallel to the strike and $\bar{x}_2$ perpendicular to strike of the fault plane. The third axis, $\bar{x}_3$, is normal to the plane (Fig. 9).

The total traction vector acting on the fault plane is commonly expressed in terms of a component, σ_n , directed normal to the plane and the maximum shear stress vector, τ , directed parallel to the plane. The maximum shear stress vector is assumed to parallel the slip vector of movement of the hanging wall of the fault surface.

A positive rake angle, θ , is counterclockwise and is assigned to slip vectors above the strike line. A negative rake angle is clockwise and is assigned to slip vectors below the strike line. Reverse and oblique reverse faults therefore have a positive rake angles and normal and oblique normal faults have negative rake angles.

Figure A1.

Figure A2.

The aim of all the stress-inversion methods (e.g., Carey and Brunier, 1974; Reches, 1987) is to construct the stress tensor,

$$T = \begin{bmatrix} T_{11} & T_{12} & T_{13} \\ T_{12} & T_{22} & T_{23} \\ T_{13} & T_{23} & T_{33} \end{bmatrix} \quad (1),$$

(Fig. A2) using the slip vectors and orientations of the fault planes measured in the field.

The principal stress directions are then calculated using the stress tensor. To do this, the measured orientation of τ in the $(\bar{x}_1, \bar{x}_2, \bar{x}_3)$ coordinate system is related to the six components of the stress tensor (Fig. A2) and the global coordinate system, (x_1, x_2, x_3) . The tensor transformation

equations (change of coordinate equations) relate the normal stress and shear stress on the fault plane to the stress components of the global coordinate system by the equations:

$$\tau \sin \theta = a_3^1 a_2^1 T_{11} + a_3^2 a_2^2 T_{22} + a_3^3 a_2^3 T_{33} + (a_3^1 a_2^2 + a_3^2 a_2^1) T_{12} + (a_3^1 a_2^3 + a_3^3 a_2^1) T_{13} + (a_3^2 a_2^3 + a_3^3 a_2^2) T_{23} \quad (2a)$$

$$\tau \cos \theta = a_3^1 a_1^1 T_{11} + a_3^2 a_1^2 T_{22} + a_3^3 a_1^3 T_{33} + (a_3^1 a_1^2 + a_3^2 a_1^1) T_{12} + (a_3^1 a_1^3 + a_3^3 a_1^1) T_{13} + (a_3^2 a_1^3 + a_3^3 a_1^2) T_{23} \quad (2b)$$

$$\sigma_n = a_3^1 a_3^1 T_{11} + a_3^2 a_3^2 T_{22} + a_3^3 a_3^3 T_{33} + (a_3^1 a_3^2 + a_3^2 a_3^1) T_{12} + (a_3^1 a_3^3 + a_3^3 a_3^1) T_{13} + (a_3^2 a_3^3 + a_3^3 a_3^2) T_{23} \quad (2c)$$

where $a_k^r = \cos(\bar{x}_k, x_r)$ are the direction cosines.

The Coulomb yield criterion is,

$$|\tau| = C + \tan(\phi), \quad (3)$$

where C is cohesion and ϕ is the internal friction angle. Equation (2c) may then be substituted into Eqs. (2a) and (2b) using Eq. (3) to obtain the two equations:

$$\begin{aligned} & \sin \alpha \tan \phi [a_3^1 a_1^1 T_{11} + a_3^2 a_1^2 T_{22} + a_3^3 a_1^3 T_{33} + (a_3^1 a_1^2 + a_3^2 a_1^1) T_{12} + (a_3^1 a_1^3 + a_3^3 a_1^1) T_{13} + (a_3^2 a_1^3 + a_3^3 a_1^2) T_{23}] \\ & = a_3^1 a_2^1 T_{11} + a_3^2 a_2^2 T_{22} + a_3^3 a_2^3 T_{33} + (a_3^1 a_2^2 + a_3^2 a_2^1) T_{12} + (a_3^1 a_2^3 + a_3^3 a_2^1) T_{13} + (a_3^2 a_2^3 + a_3^3 a_2^2) T_{23} \end{aligned} \quad (4a)$$

$$\begin{aligned} & \cos \alpha \tan \phi [a_3^1 a_1^1 T_{11} + a_3^2 a_1^2 T_{22} + a_3^3 a_1^3 T_{33} + (a_3^1 a_1^2 + a_3^2 a_1^1) T_{12} + (a_3^1 a_1^3 + a_3^3 a_1^1) T_{13} + (a_3^2 a_1^3 + a_3^3 a_1^2) T_{23}] \\ & = a_3^1 a_3^1 T_{11} + a_3^2 a_3^2 T_{22} + a_3^3 a_3^3 T_{33} + (a_3^1 a_3^2 + a_3^2 a_3^1) T_{12} + (a_3^1 a_3^3 + a_3^3 a_3^1) T_{13} + (a_3^2 a_3^3 + a_3^3 a_3^2) T_{23} \end{aligned} \quad (4b)$$

After normalizing each stress component with T_{33} , eqs. (4) can be expressed in the matrix form

$$A \begin{bmatrix} T_{11}^* & T_{22}^* & T_{12}^* & T_{13}^* & T_{23}^* \end{bmatrix}^T = B, \quad (5)$$

where

$$A = \begin{bmatrix} a_3^1 a_2^1 + (a_3^1)^2 \tan \phi \sin \theta & a_3^2 a_2^2 + (a_3^2)^2 \tan \phi \sin \theta & a_3^1 a_2^2 + a_3^2 a_2^1 + 2a_3^1 a_3^2 \tan \phi \sin \theta & a_3^1 a_2^3 + a_3^3 a_2^1 + 2a_3^1 a_3^3 \tan \phi \sin \theta & a_3^2 a_2^3 + a_3^3 a_2^2 + 2a_3^2 a_3^3 \tan \phi \sin \theta \\ a_3^1 a_1^1 + (a_3^1)^2 \tan \phi \cos \theta & a_3^2 a_1^2 + (a_3^2)^2 \tan \phi \cos \theta & a_3^1 a_1^2 + a_3^2 a_1^1 + 2a_3^1 a_3^2 \tan \phi \cos \theta & a_3^1 a_1^3 + a_3^3 a_1^1 + 2a_3^1 a_3^3 \tan \phi \cos \theta & a_3^2 a_1^3 + a_3^3 a_1^2 + 2a_3^2 a_3^3 \tan \phi \cos \theta \\ \vdots & \vdots & \vdots & \vdots & \vdots \end{bmatrix}$$

$$B = \begin{bmatrix} -a_3^3 a_2^3 - (a_3^3)^2 \tan \phi \sin \theta \\ -a_3^3 a_1^3 - (a_3^3)^2 \tan \phi \cos \theta \\ \vdots \end{bmatrix}$$

and $T_{ij}^* = T_{ij} / T_{33}$. A least squares solution is applied to the system to determine the best fit stress components, T_{ij}^* .

The principal stress directions are given by the eigenvectors, n , of the normalized stress tensor T^* (solutions to the system $(T - \lambda I)n = 0$).

Conveniently, the tractions and principal stresses can be expressed in the form of an ellipsoid, called the stress ellipsoid. The form of the ellipsoid is (Fung, 1969)

$$\frac{T_1^2}{\sigma_1^2} + \frac{T_2^2}{\sigma_2^2} + \frac{T_3^2}{\sigma_3^2} = 1 \quad (3.6)$$

where T_i are the components of the traction vector acting on a plane and σ_i are the principal stresses. Therefore the lengths of the three axes of the stress ellipsoid are proportional to the principal stresses.

This completes the stress inversion.

To implement the stress inversion method outlined above, I have written a MATLAB® program to which is attached a (GUI) user-friendly interface (the code is in Appendix). The GUI interface is illustrated in figures 3.13 through 3.16. The user enters the strike, dip, dip direction, and rake in the boxes in the lower right-hand corner, and then clicks on the "CALCULATE" button to calculate the stress tensor. The eigenvectors, giving the principal directions, the magnitudes of the principal stresses; and the angular errors (sd; one standard deviation) are displayed in the box in the lower left-hand corner. The stress ellipsoid, the geometrical information of the faults, and the mean misfit angle are displayed at the top of the interface. The stress ellipsoid is rotated such that the axes are oriented in the direction of the principal stresses.

PROPAGATION OF DEFORMATION BANDS IN POROUS SANDSTONES ©

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First Draft Spring 1999

For Fracture Mechanics Course

Revised Dec. 2001

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INTRODUCTION

Atilla Aydin's (1977, 1978) classic study of a type of faulting in porous rocks provided the first well-documented mechanisms of faulting in rock. Atilla discovered deformation bands, shear zones and slip surfaces, which are related kinds of fault-like localized deformation in porous sandstones that had gone uncharacterized theretofore. Atilla pointed out that the shear zones occur in two forms: *deformation bands* and *zones of deformation bands*. In addition, he recognized that faults occur within some zones of deformation bands as *slip surfaces*.

The shear zones³⁶ known as deformation bands in porous sandstones are described at various levels to detail elsewhere (Aydin, 1977, 1978; Aydin and Johnson, 1978, 1983; Pittman, 1981; Aydin and Reches 1982; Jamison and Stearns, 1982; Smith 1983; Underhill and Woodcock, 1987; Zhao and Johnson, 1991; Antonellini and others, 1994). Here the descriptions will be brief,

³⁶ I call them *shear zones* because the shearing is distributed across the thickness of a very narrow band, rather than concentrated on a surface. One might well argue that this is only a matter of scale of observation, but, in fact, one cannot observe anything resembling a fault surface. Aydin (1977, 1978) calls them "faults," as I have, elsewhere (e.g., Aydin and Johnson, 1978; Zhao and Johnson, 1991). In this paper and its companion, though, I need to distinguish surfaces of displacement discontinuity from shear zones, so it makes sense to maintain the traditional usage of the term *fault*.

rather we will see deformation bands, zones of deformation bands and slip surfaces in an atlas of photos.

Note: I have put in smaller type a few passages that, perhaps, would be edited out of a published version of this document, if this document were ever submitted for publication. Thus, if you see a passage like this—this explanation by Jim is hardly likely and, as such, represents some of the fuzziest thinking and exuberant nonsense I have seen since the early days of plate tectonics—take my word for it and skip the passage. You probably can live without my assessment of this issue!

ATLAS OF PHOTOS

This manuscript concerns the growth of *deformation bands*, a type of highly localized deformation in porous sandstones, first carefully described in the San Rafael Desert and Arches National Park, Utah by Atilla Aydin (Aydin, 1977, 1978; Aydin and Johnson, 1978, 1983).

A single deformation band (or perhaps a pair) highlighted by evening sun near Goblin Valley, Utah.

A single deformation band viewed with a low-power microscope. Band offsets fine-grained and very-fine grained sandstone layers.

Much closer view of a single deformation band in Entrada Sandstone.

Several deformation bands near Goblin Valley, Utah.

Idealization of a single deformation band according to Atilla Aydin (1977). There is extreme (1:5000) exaggeration of scale here. The horizontal length of this body is perhaps ten meters. The vertical length is about two mm. Nevertheless, the diagram shows the essential features of a deformation band in medium-grained sandstone.

- The host rock, several mm on either side of the band, is a lightly cemented sandstone with, say, 24% porosity.
- The grains in the host rock are relatively uniform in size and generally touching at point contacts.
- This deformation band would appear to be about one mm wide in the field.
- The trace of the deformation band may be a few meters or many tens of meters.
- In thin section, one sees an inner zone, about half that width, of crushed sandstone grains, with a few surviving full grains "floating" therein, as well as two outer zones, accumulating about half the width in which grains have been damaged and pushed closer together.
- The porosity in the outer zones has been reduced somewhat. The porosity in the inner zone has been reduced to about five percent.

More pretty deformation bands in same area.

Deformation bands many occur in one set or many sets. Here there appear to be four sets. Two sets have a common strike trending vertically in the view but have opposite dips. The other two sets have a common strike trending horizontally in the view but, again, with opposite dips. There may be other sets. It is unusual to have the strikes nearly at right angles. Photo taken near Goblin Valley, Utah.

A wide zone (about 30 cm wide) of deformation bands near Goblin Valley, Utah. The zone is composed of many (white) deformation bands, side-by-side. This zone accommodated almost pure dip shift. It is part of a normal fault zone in San Rafael Desert. (Photo by Atilla Aydin).

Atilla points out that, interestingly, *deformation bands* almost never occur singly; rather, they occur as a member of a *zone of deformation bands*, composed of two or more deformation bands that share the same average strike and dip (Aydin and Johnson 1978). A zone may be narrow or wide and contain several bands tightly packed side-by-side or may contain loosely packed, inosculating and wavy bands. The thickness of a zone depends on the number and packing of bands it contains.

Coalescence of two narrower zones of deformation bands in same area. Part of a normal fault zone.

Multiple sets of zones of deformation bands. Near Molly's Castle, Utah.

Multiple sets of narrow zones of deformation bands.

Slip surface near Molly's Castle. Slip surface is wall-like structure in center. Red sandstone bed to left has been down-dropped several meters from its position shown to right. Dave Pollard, left foreground. Paul Segall next to him.

Another view of the slip surface. Amos Nur (right), George Thompson (center) and a San Rafael Bum discuss the weather.

Magnificent view of a magnificent slip surface at same location. Paul Segall examining the fault.

A *slip surface*, occurs within some zones of highly concentrated deformation bands, as a discrete surface of discontinuity in displacement with well-developed striations and grooves. The slip is much larger across slip-surfaces than across bands or zones of bands (Aydin and Johnson 1978).

View along strike of the slip surface, showing adjacent thick zone of deformation bands (about 25 cm thick). Thinner zones to right. Slip surface is on one side of the thick zone.

Close-up view of slip surface. Horizontal lines are small conjugate faults that strike roughly at right angles to the slip direction on the slip surface.

GENETIC RELATIONS OF DEFORMATION BANDS, ZONES OF DEFORMATION BANDS AND SLIP SURFACES

Spatial relations among deformation bands, zones of deformation bands and slip surfaces indicate that these three structures form sequentially, that the individual deformation band is a primary structural element, and that the formation of a band marks an early stage in the process of faulting of these rocks.

The deformation is highly localized within a narrow band. It is similarly highly localized within a zone of deformation bands.

- Permanent deformation in a band is by fracturing and displacement of grains, by distortion of the matrix and by reduction of pore volume.
- There is both thickness decrease and shear offset across a band. The magnitude of the volumetric strain is at least 0.2, and the average shear strain is on the order of 1 to 10, so the ratio of shear strain to volumetric strain within a band is on the order of 5 to 50 (Aydin 1978).
- The deformation bands form side by side within a zone. This raises questions as to why further deformation is accommodated by the formation of a new band, rather than by continued displacement on a pre-existing band, and why the new band forms immediately adjacent to the pre-existing band.
- The trace of *an individual deformation band tends to be straight*, but the traces of deformation bands within a zone are wavy and inosculate in a plane normal to the direction of shear.
- *Each type of fault occurs in networks composed of multiple sets.* In the dip- and oblique-slip faulting in San Rafael Desert and at Arches National Park, there are more than two, and typically four, sets (Aydin and Reches 1982), whereas in the strike-slip faulting at Arches there are typically two sets (Zhao and Johnson 1991a,b). Each type of fault, however, can occur in domains containing a single set of faults.

OTHER NAMES AND STUPID THINGS LIKE THAT

Others had *seen* the three kinds of structures brought to light by Atilla, of course, but had not *studied* them. Terry Engelder did his Ph.D. research on the same kinds of shear zones at about the same time but did not understand what he was studying (Engelder, 1974). It is unclear whether Terry's Ph.D. thesis advisor, Dave Stearns (see Jamison and Stearns, 1982), understood the significance of these structures either. Later Terry called them "braided shear fractures" (Engelder, 1987), avoiding the clear terminology introduced in Atilla's papers³⁷. The deformation bands commonly have the appearance of being "braided" when they are collected in zones of

³⁷ Even worse, he identified as "pinnate joints" and "non-systemmatic joinings" structures that are dead ringers for deformation bands in the Cooma complex, Australia (Engelder, 1989, Figure 2.13, 2.15).

deformation bands, although Atilla's more careful observations indicated that the deformation bands in such "braided" zones actually inosculate rather than cross as they would have to to make a braid. Thus even the term "braided" is as misleading as the term "gouge." It is, furthermore, unclear whether the term "fracture" really is appropriate for what one observes. Certainly at the grain scale one sees fractured grains. But the deformation bands themselves are not obvious fractures, nor are the zones of deformation bands. Finally, Terry's characterization totally missed the vital connection that Atilla recognized, that deformation bands begat zones of deformation bands and zones of deformation bands begat slip surfaces—there is a clear sequence of events that leads through these three types of localized deformation.

The kind of fault known in the petroleum industry as a *seal fault*, (e.g., Hardman and Booth, 1991), is apparently the same as the deformation bands and zones of deformation bands, and may also include slip surfaces; I wonder whether people who talk about "seal faults" have read Atilla's lucid description of these structures. The name "seal fault," though, is a good example of undescriptive "oil company talk!"

Bernabe and Brace (1990) used a series of terms, apparently also in order to avoid Atilla's clearly crafted terminology: the deformation bands are called "shear fractures," "shear bands," "fracture" and "fracture zone." Scott and Nielsen (1991) called deformation bands produced experimentally "brittle shear fracture" and "cataclastic shear zone."

What lengths will people go to avoid giving an excellent scientist credit for work well done? Atilla's descriptions are unmatched and his terminology is logical. Somehow, it is as though the other investigators make every attempt to point the light their own way!

PHYSICAL PROPERTIES OF POROUS SANDSTONES

Rudnicki and Rice (1975) developed a theory of rupture of a generalized Coulomb, strain hardening, elastic-plastic material and determine the conditions under which one would expect deformation bands to form in such materials. The materials are characterized by the usual elastic properties and moduli plus angle of internal friction, cohesion, dilatancy and hardening modulus. Aydin and Johnson (1983) tried to apply the Rudnicki and Rice theory to the formation of deformation bands in porous sandstones. Their efforts were frustrated by lack of appropriate experimental measurements for rocks. One had to guess most of the parameters, so the results were highly speculative.

Later on, Bernabe and Brace (1990) ran experiments with Berea Sandstone at various confining pressures and Scott and Nielsen (1991) ran experiments with several other sandstones. The experiments were quite interesting because the researchers clearly had produce deformation bands. The experiments were quite frustrating, though, because they had not been designed to extract useful information that could be used in theoretical analysis. The latter aspect of the experiments is discussed at some length in Appendix A.

Nevertheless, the experiments did show that deformation bands could be produced in porous sandstone for a range of confining pressures though.

PEAK AND RESIDUAL STRENGTHS (OF GRANULAR SOLIDS)

A casual comparison of stress-strain relations for porous sandstones indicates that the relations are remarkably similar to those of soils tested extensively by geotechnical engineers. In particular, for both rocks and soils, one can identify what are known as *peak* and *residual strengths*.

Figure 1. Idealized stress-strain and volumetric-strain shear-strain relations for samples of overconsolidated and normally consolidated granular solids in a compression apparatus (After Lambe and Whitman, 1969, p. 302). A. Overconsolidated materials show a high strength—the peak strength—followed by a low strength—the residual strength. Normally consolidated soils show a steady increase in strength that levels out at the same or nearly the same residual strength. B. The overconsolidated granular solid is characterized by dilatancy—an increase in volume—during shearing. For low shear stresses, the material decreases in volume elastically. Then it begins to dilate. When the shear strain becomes large enough the volumetric remains constant at a value that is characteristic of the confining pressure in a test. The normally consolidated and the underconsolidated soils are characterized by decreasing volume, both elastically and plastically, that approaches a constant value.

A typical stress-strain curve shows a steep rise in stress with strain, generally reflecting elastic deformation (Figure 1A). At a certain stress level the curve forms a sharp peak—this is the peak strength—and dives down again to a lower stress, where it levels off for even large deformations—this is the residual strength. Certain soils, particularly those responsible for rapid failures in eastern Canada, Alaska and Scandinavia, known as “quick clays,” are characterized by high peak and low residual strengths. For a typical “quick clay” the ratio of peak to residual strengths is 50 but values may range from 10 to 100 or even higher. We will see that the ratios of peak to residual strength for Dobbs Valley Sandstone and Berea Sandstone range from about 1.1 to 1.3, much lower than

those of quick clays. In this respect we might say that the quick clay soil is much more brittle (Mitchell, 1976, p. 290) than the sandstone.

According to Mitchell, brittle soils, which have distinct peak and residual strengths, are characterized as cemented or overconsolidated soils in geotechnical engineering. In contrast, a peak strength is lacking in remolded samples of the same soil; remolding breaks cement between grains and generally loosens the soil. The remolded soil is thus *normally consolidated* at atmospheric pressure. If one determines the stress-strain curve for the normally consolidated soil, we see that the slope of the stress-strain gradually reduces and the shear stress approaches the residual strength of the soil. The residual strength is the same for the same soil, whether the soil is overconsolidated, normally consolidated or underconsolidated.

The presumed analogy between the behaviors of overconsolidated soils and rock might provide a useful framework for discussing behavior of intact samples of porous sandstones tested experimentally by Scott and Nielsen (1991) and Bernabe and Brace (1990). Both tested porous sandstones in triaxial apparatus and derived relations between axial compression and axial strain as a function of confining pressure. Bernabe and Brace (1990) also determined volume changes as a function of axial compression, so they provided valuable information for assessment of dilatancy during deformation of the rock they tested.

THE ROCKS

Scott and Nielsen (1991) tested samples of sandstones with porosities ranging from about 8% to 30% and at low to moderate confining pressures. Here we will examine their data for Dobbs Valley Sandstone. The Dobbs Valley Sandstone has high porosities, as do the Navajo Sandstone and the Moab tongue of the Entrada Sandstone. The sandstones are very fine to fine grained and moderately well sorted. Cement is generally quartz. The samples with the highest porosity, about 28%, contained about 10% clay. The others, with a porosity of about 21%, had small amounts of clay and rock fragments (op. cit., p. 406).

The Berea Sandstone tested by Bernabe and Brace (1990) consists primarily of subangular, well-sorted quartz grains about 150-200 μ (0.15 to 0.2 mm) in maximum dimension. About 20% of the sandstone is composed of dolomite cement and grains of other compositions. The porosity of samples tested ranged from 18 to 19%, about the same as the less porous samples of Dobbs Valley Sandstone.

STRESS-STRAIN RELATIONS FOR SANDSTONES

Scott and Nielsen (1991) experimentally tested samples of sandstones with porosities ranging from 6 to 28% at confining pressures at 0.1 to 130 MPa³⁸ in order to determine the gross behaviors of several sandstones at different confining pressures. The aim of their experiments was to specify the conditions under which intact samples failed in different ways, and to compare the stress-strain behavior with samples containing a sawcut (also, see Evans, Fredrich and Wong, 1990).

³⁸ These and following lithostatic pressure values are based on assumption that specific gravity of rock is 2.5 and the unit weight is 25000 N/m³, that is, 25 kPa/m. Thus 0.1 MPa at 4 m; 1 MPa at 40 m; 10 MPa at 400m; 100 MPa at 4 km. Pressures of 0.1 and 130 MPa are equivalent to pressures at depths of about 40 m and 2 km, respectively.

DOBBS VALLEY SANDSTONE

We begin our review of the stress-strain behavior of Dobbs Valley Sandstone, following the general approach within the field of rock mechanics. Some results for two samples of the Dobbs Valley Sandstone are shown in Figure 2. In Figure 2A are stress-deformation³⁹ curves (solid) for an intact sample and a sample with a sawcut at 35° to the core axis at confining pressures of 2.5, 10 and 30 MPa. The initial porosity was about 28%.

Sawcut

The three curves (dashed) for sawcut samples show stress-strain relations that are roughly of the same form regardless of confining pressure. The slope of the curves smoothly lowers with increasing strain and appears to approach a constant stress that increases with confining pressure. In many respects the behavior of sawcut samples is similar to that of a normally consolidated soil (Figure 1A).

Intact

There is a strikingly different stress-strain relation for intact samples under low (2.5 to 10 MPa) confining pressures. As these samples were loaded, the stress increased nearly linearly with strain, but as the stress approached a peak value, the slope of the stress-strain curve gradually lowered and then abruptly approached zero at the *peak strength*. As straining continued, the load the sample could support decreased, and the stress-strain curve has a strong, negative slope. As straining continued further, the slope remained negative, but became smaller in magnitude. As the strains became quite large (0.02 to 0.03 for these samples), the slope was nearly zero so the stress was independent of strain, and the shear stress became the *residual strength*.

Thus, for samples of sandstone of 28% porosity, confined at pressures of 2.5 and 10 MPa, the peak and residual strengths are evident. For the lower confining pressure the stress-strain curve rises sharply to a peak, at about 21 MPa for a strain of 0.005, and then reduces abruptly to an essentially constant value of 13 MPa for strains larger than 0.02. Thus the peak strength would be about 21 MPa and the residual strength would be about 13 MPa. For the higher confining pressure, the peak is 37 MPa at a strain of about 0.007 and reduces to about 33 MPa at a strains larger than 0.02. The peak strength for 10 MPa confining pressure is about 37 MPa and the residual strength is about 33 MPa.

Perhaps one of the more important results of the testing by Scott and Nielsen (1991) is the fact that the ultimate strengths of the intact and sawcut samples of Dobbs Valley Sandstone are essentially the same. This suggests that the residual strength can be obtained in two quite different ways. The residual strength is apparently approached as the intact samples grow new fractures and as the sawcut surfaces become rougher with displacement. Thus the residual strength of an intact sample can be used to estimate the strength on a sawcut of the same sample.

For higher confining pressures (11 to 70 MPa in Scott and Nielsen's paper), an increase in axial strain results in an ever increasing stress difference, but the material is *strain-softening*. That is, the

³⁹ For the whole sample the strain is simply called strain by the authors, so I interpret it to be axial strain, that is, change in length per unit of sample length. For the sawcut samples the strain is called shear strain. It is quite unclear what the authors mean by this term, so I assume "shear strain" also indicates axial change in length per unit length. The general conclusions of the research do not depend on this assumption.

higher the strain the lower the slope of the stress-strain curve. At 30 MPa confining pressure the stress-strain curves for sawcut and intact samples are essentially the same. The behavior of sawcut samples is similar to that of a normally consolidated soil.

From Thurman E. Scott and K.C. Nielsen
JGR, 96:405-414, 1991.

Figure 2. Stress-strain relations for two porous sandstones. The results are for a range of confining pressures, ranging from 2.5 to 30 MPa. The solid curves are for intact samples⁴⁰. The dashed curves are for samples containing a sawcut fracture inclined at 35° from the largest compressive stress. (The data used to compute these curves were provided in the paper by Scott and Nielsen.)

A.

⁴⁰ Unfortunately, it appears as though Scott and Nielsen misidentified their diagrams. Note that the identification of the two sets of curves described above is opposite to that presented by Scott and Nielsen in their figures 7 and 4. As they label the figures, their graphs make no sense: The main argument is that the sawcut samples must have strengths that are less than or equal to those of intact samples. Another argument is that the curves presented by Bernabe and Brace (1990) for low confinement of intact samples of Berea Sandstone show distinct peak strengths, as one would expect.

B.

Figure 3. Stress-strain and axial strain/volumetric strain curves for Berea Sandstone subjected to (effective) confining pressures ranging from 12 to 67 MPa (data from Bernabe and Brace, 1990, Figure 4). According to the conventions of engineering mechanics, positive stress and strains are extensile; negative stress and strains are compressive. Thus the results are plotted on the negative axes. A. Stress difference (axial stress minus confining stress) on vertical axis and axial strain on horizontal axis. B. Relation between axial strain and volumetric strain.

BEREA SANDSTONE

The stress-strain curves shown in Figure 3A for Berea Sandstone are much like those presented by Scott and Nielsen (1991). At low confining pressure, 12 MPa, there is a strong trough, representing the "peak" strength and there is a short segment of low strength, presumably representing the "residual" strength. The ratio of the peak to residual strengths is about 2.4. At a confining pressure of 37 MPa, the ratio of peak to residual strength is about 1.5. For larger pressures the ratios are 1.2 and 1.1 for confining pressures of 52 and 67 MPa, respectively.

Figure 3A shows relations between stress difference and axial strain for Berea Sandstone according to Bernabe and Brace (1990), for confining pressures ranging from 12 to 67 MPa (equivalent to 0.5 to 2.7 km depth, respectively). Figure 3B shows relations between maximum shearing strain and volumetric strain during the tests. The tests were all compression tests so the stresses and most of the strains are shown as negative quantities. A positive volumetric strain indicates a volume increase and a negative volumetric strain indicates a volume decrease, for example. The normal strains are also positive if extensile. Figure 3B shows that all samples decreased in volume during initial stages of loading, but samples confined at 12 or 37 MPa eventually dilated, increasing their volumes.

The results shown in Figure 3 are not particularly useful for an analysis of any process of deformation for several reasons:

- The results are not cast in terms of stress and strain invariants. In order to understand properties of isotropic materials, one must work with invariants (e.g., see Rivlin and Saunders, 1951; also, see Appendix A herein).
- For elastic and elastic-plastic materials there are three stress invariants and three strain invariants that one must consider.
- In order to investigate the properties of any isotropic material, one must vary, for example, one stress invariant while maintaining the other two constant. For example, for

one series of tests, one must determine the relations between stress and strain while varying the pressure (average normal stress) on the sample but keeping the second and third invariants constant.

- As indicated in Appendix A, the tests run by Bernabe and Brace (1990) and by Scott and Nielsen (1991) vary all three stress and strain invariants simultaneously. This results in the experimental equivalent of absolute chaos. In the experiments the confining pressure is held constant, but the problem is that the confining pressure is not invariant, so such a test is extremely difficult to interpret, at best. An attempt to analyze the results of Bernabe and Brace (1990) is presented in Appendix A.
- Finally, the quantities shown in stress-strain curves do not relate well to the processes of deformation or rupture in samples. For example, at low stresses, the deformation is presumably elastic, or nearly so. For such conditions the average axial stress and the average change in length of a sample may be good measures of what is happening inside the experimental apparatus. At higher stresses there may be both elastic and permanent deformation at some time. Still, average measures might be relevant because the deformation might be homogeneous.
- However, then something horrible happens: the deformation becomes localized in some fashion. Now the overall measures of change in length of the specimen do not nicely reflect what is happening in the apparatus. Also, if the deformation is localized on an inclined surface, one must know the geometry of that surface. The stress-strain curve is no longer easily interpreted.

Unfortunately, these are the typical problems with high pressure experiments run on rock samples. The necessary observations are very difficult to make and so they are rarely made.

The way we did it years ago (Peng and Johnson, 1972), when we discovered how granite specimens actually failed in lab tests, was to use a series of nearly identical samples and to load to different levels each of several samples covered with strain gauges. Then we cut the samples and examined and mapped the microscopic and macroscopic fracturing. In this way we discovered, for example, that the typical way for a uniaxial test specimen to fail in a proper compression test was by axial splitting. Before those experiments, the party line in rock mechanics was that such samples failed along inclined fractures, considered to be faults.

However, we have been provided with the results of Bernabe and Brace (1990) and Scott and Nielsen (1991), so we are stuck with the interpretation of confusing results.

DESCRIPTIONS OF WHOLE SAMPLE DEFORMATION

According to Bernabe and Brace (1990), the deformation of samples of Berea Sandstone under various confining pressures changes drastically. At relatively low confining pressures, the samples fail unstably along a single deformation band. The definitions of the terms are:

Low confining pressure: 10-15 MPa
Intermediate confining pressure: 16-40 MPa
Moderately high confining pressure: 41-60 MPa
High confining pressure: 61+ MPa.

At intermediate confining pressures, the samples fail along a few, distinct, conjugate deformation bands. The number of deformation bands increases in the sample as the confining pressure is increased. At high confining pressure, there are no deformation bands, rather the whole sample

becomes, in effect, a broad deformation band. The sample as a whole bulges and the rock, in a sense, flows.

According to Bernabe and Brace (1990), the grains in the Berea Sandstone rotated, cracked and slid whatever the confining pressure on the samples. For low confining pressure, though, the grains within the deformation band are crushed on a zone about one grain dimension wide. In zones the same width on either side, some of the grains are cracked. These observations are consistent with the observations of deformation bands in Entrada Sandstone by Aydin (1978). Scott and Nielsen indicate that some of the grains are cracked within 3 to 10 grain diameters from the center of the deformation band.

In fact one of their examples of "a sample deformed in the brittle regime" (their Figure 6) is reminiscent of the *deformation bands* described by Aydin (1978) in Entrada and Navajo Sandstones. The photograph illustrating "brittle" deformation is strikingly similar to Aydin's (1978) classic drawings and photographs of these structures. The Berea Sandstone contains mostly (perhaps 90%) grains 100 to 200 μ (0.1 to 0.2 mm) in maximum dimension. There is a band of shattered grains, about a grain-size wide, that trends diagonally, at an angle of about 20° to the direction of maximum compression. Within the band the grains have been broken into sharp-cornered fragments that appear to be at least bimodal, with a common size of about 100 μ and another at a size of about 10 μ . Outside the band (actually still inside the band, as defined by Aydin, 1978), within about 200 μ of either side of the crushed zone, some of the larger grains have been fractured by one or

At low confining pressures, the stress-strain curves show that there is compaction early in the deformation, followed by recovery of the volume and then increase in volume through dilatation. The dilation starts before deformation bands initiate and continues as deformation bands grow. These results, though, may reflect more the dilation of the entire sample rather than dilation within a deformation band. The early dilation prior to band formation appears to be a result of sliding at grain contacts. The compaction that we observe in deformation bands, though, is not reflected in the stress-strain curve. That compaction is clearly a result of fracturing and compaction of grains with a deformation band. At moderate to high confining pressures the stress-strain curves show only compaction; never dilation. Bernabe and Brace (1960) point out that this compaction is like that within a deformation band; it is a result of fracturing and rearrangement of grains within the rock.

The deformation bands in Berea Sandstone form at a high angle to the direction of maximum compression in the samples. At low confining pressures the angle between the maximum compression and the bands was about 20° (actually it ranged from about 20 to 30° in various samples). At intermediate confining pressure the angle was generally 30 to 40°, and at high pressure the angle approaches 45°.

ANALYSIS OF PROPAGATION OF DEFORMATION BANDS

In many respects, deformation bands must propagate much as fractures. Like fractures, they are characterized by great length-to-thickness ratios, so that a small amount of shift is sufficient to propagate a long deformation band. We have shown in previous pages that the relation between axial compression in cylindrical samples of Berea Sandstone required to cause slip within a deformation band is a highly nonlinear function, that there is a phenomenon like peak strength of a soil. After the peak strength is reached, the strength drops precipitously. In this respect the deformation band behaves mechanically analogously to fractures in soils and in some rocks (porous sandstones, at least).

We are going to apply an analysis that was developed to describe propagation of mode II fractures, mainly by Jim Rice, to the propagation of idealized deformation bands. Palmer and Rice (1973) and Rice (1980) have shown that the G_{II} or K_{II} can be evaluated for a propagating, mode II fracture by performing an appropriate integration of the post-failure, stress-displacement curve. We will adapt the method to analyze propagation of a deformation band.

In order to do so, we need to make some assumptions about deformation within the samples of Berea Sandstone subjected to relatively low confining pressures. We assume that, as the peak of the stress-strain curve is approached (Figure 4A), the deformation becomes localized within a single band. This assumption is consistent with the descriptions of samples by Bernabe and Brace (1990). Thus, we compute the displacement accommodated by a single deformation band by assuming that the additional shortening, following the peak, is accommodated not by the sample as a whole, but entirely by shift across the newly formed deformation band.

The amount of displacement across the band is computed from that additional shortening. We assume that the deformation band is inclined 25° to the axis of maximum compression. Bernabe and Brace indicate that, for low confining pressure, the range is from 20 to 30° . Therefore, for every millimeter of further shortening of the entire sample there is $1 \text{ mm}/\cos(6^\circ)$ of shift, δ , across the width of the band. Thus we know the amount of shift. We know the shear stress parallel to the band, because it is

$$\tau = [(\sigma_a - \sigma_c)/2]\sin(50^\circ)$$

The result is the curve in Figure 4A for a confining pressure of 12 MPa and Figure 4B for a confining pressure of 37 MPa.

Before proceeding with the analysis, we need to know something about Rice's J-integral.

THE J-INTEGRAL OF RICE

The analysis of propagation of a deformation band depends on the J-integral of fracture mechanics (Rice, 1968). The J-integral has three parts. Let x_1 and x_2 be coordinate axes so that a straight deformation band lies parallel to the x_1 axis. Let there be plane strain deformation in the x_1 - x_2 plane.

- Then, the stress-strain behavior of the material outside the deformation band is such (generally linearly elastic) that the strain energy (energy per unit of volume), W , at any strain ϵ_{pq} , is independent of the path of integration,

$$W(\epsilon_{pq}) = \int_0^{\epsilon_{pq}} \sigma_{ij} d\epsilon_{ij} \quad (1a)$$

A.

B.

Figure 4 Relations between shear stress and displacement for Berea Sandstone for two confining pressures.

Figure 5. An integration path that evaluates the J-integral around a fracture tip. T is the fracture tip. Γ is the path of integration. n_1 and n_2 are components of the normal in the x_1 and x_2 directions, respectively.

The second part derives from the tractions across the path of integration,

$$T_i = \sigma_{ij}n_j \quad (1b)$$

The third part derives from body force per unit of volume, f_i . In these terms Rice's J-integral for a fracture is,

$$J_P = \int_{\Gamma} [(W - f_i u_i) dx_2 - T_i \frac{\partial u_i}{\partial x_1} ds] \quad (1c)$$

USING J-INTEGRAL TO EVALUATE G_{IIC} AND K_{IIC} FOR A DEFORMATION BAND

We can take a step toward understanding propagation of a deformation band if we determine the J-integral in terms of the strength distribution along a mode II band. We will consider a band that is in a finite medium, as shown in Figure 6, but the finite medium is no concern just now. We will integrate along the surface of the deformation band. Consider an integration path that extends along the bottom of the band from P^* to the tip of the band, T, and back along the top of the band to point P^* (Figure 6).

Figure 6. Elastic body containing a deformation band and subjected to shearing as a result of lateral translation of rigid confining media at top and bottom.

Examination of eq. (1c),

$$J_P = \int_V [(W - f_i u_i) dx_2 - T_{i2} \frac{\partial u_i}{\partial x_1} ds]$$

indicates that the first two terms vanish if we can reasonably assume that dx_2 is negligible. In this sense, we assume that a deformation band is extremely thin, just like a fracture. Thus,

$$J_P = \int_V [-T_{12} \frac{\partial u_1}{\partial x_1} dx_1 - T_{22} \frac{\partial u_2}{\partial x_1} dx_1] \quad (2a)$$

The second term in the resulting expression, eq. (2a), integrates to zero.

If we assume that the deformation band changes thickness uniformly, then the displacement gradient, $\frac{\partial u_2}{\partial x_1}$, is continuous across the band. Also, the traction, T_2 , on the upper surface is equal and opposite to that on the lower surface so the integral is zero. The remaining term is

$$J_P = \int_V -T_{12} \frac{\partial u_1}{\partial x_1} dx_1 \quad (2b)$$

It contains a discontinuous quantity, u_1 , which changes abruptly at the band. Thus we break the integral into two parts,

$$J_P = \int_{P^-}^{T} T_1 \frac{\partial u_1^-}{\partial x_1} dx_1 - \int_T^{P^+} T_1 \frac{\partial u_1^+}{\partial x_1} dx_1 \quad (2c)$$

The traction is minus the shear stress, τ_{21} , on the upper part of the deformation band (second term), whereas it is plus on the lower part, so that,

$$J_P = - \int_{P^-}^T \tau_{21} \frac{\partial u_1^-}{\partial x_1} dx_1 + \int_T^{P^+} \tau_{21} \frac{\partial u_1^+}{\partial x_1} dx_1 \quad (2d)$$

Rearranging,

$$J_P = \int_T^{P^+} \tau_{21} \frac{\partial (u_1^+ - u_1^-)}{\partial x_1} dx_1 = \int_T^{P^+} \tau_{21} \frac{\partial \Delta u_1}{\partial x_1} dx_1 \quad (2e)$$

where Δu_1 is the shift across the deformation band and τ_{21} is the shear strength along the band. Note that u in Figure 6 is Δu_1 in eq. (2e)

For a deformation band, the shear strength, τ_{21} , is determined by the amount of slip, Δu_1 , as indicated in Figure 4. For this reason, we can write eq. (2e) in the form,

$$J_P = \int_0^{\Delta u_1} \tau_{21} d\Delta u_1 \quad (3a)$$

Thus we can determine the J -integral in terms of the strength along the deformation band.

Figure 7. Relations between shear stress and displacement for Berea Sandstone, at a confining pressures of 12 MPa and 37 MPa. Peak and residual strengths are noted. Areas used in the computation shown with different symbols.

This integral is the entire area beneath the strength-displacement curve, for example, shown in Figure 7A. (The small depression in the curve is probably unreal and is ignored.) The tip of the fracture is represented by 0 displacement. The displacement increases from 0 to 0.22 mm for this sample

Palmer and Rice express the energy in terms of the shear traction within the main body of the band, τ , and the shear traction in front of the band, τ_p , by subtracting the residual strength, τ_r , times the displacement, δ_p , at a point within the band. We obtain the result that

$$J_p - \sigma_r \sigma_p = \frac{1}{2} [\sigma - \sigma_r] \sigma_r \quad (3b)$$

This formulation is of special value because it leads directly to the so-called energy release rate, G_{II} , as well as to the critical stress-intensity factor, K_{II} , for mode II propagation. Thus, according to Palmer and Rice,

$$J_p - \sigma_r \sigma_p = \frac{1}{2} [\sigma - \sigma_r] \sigma_r = G_{IIc} = \frac{1}{2G} K_{IIc}^2 \quad (3c)$$

For Berea Sandstone, we proceed as follows. For the sample subjected to 12 MPa confining pressure (Figure 7A) the traction on the band, $\sigma(?)$, ranges from the residual value of about 19 MPa to a peak value of 46 MPa.

$$\sigma_r = 19 \text{ MPa}$$

$$\sigma_p = 46 \text{ MPa}$$

Integrating the difference between the traction on the band and the residual strength, we obtain a value of

$$\frac{1}{2} [\sigma - \sigma_r] \sigma_r = G_{IIc} = \frac{1}{2G} K_{IIc}^2 = 1.99 \text{ MPa mm (0.00199 MPa m)}. \quad (4a)$$

We can express this result in terms of a characteristic displacement $\bar{\delta}$,

$$(\sigma_p - \sigma_r) \bar{\delta} = \frac{1}{2} [\sigma - \sigma_r] \sigma_r = 0.00199 \text{ MPa m}$$

so

$$(\sigma_p - \sigma_r) \bar{\delta} = 0.00199 \text{ MPa m}$$

which gives a value of characteristic displacement across the deformation band of about half the grain size (0.15 to 0.20 mm),

$$\bar{\delta} = 0.07 \text{ mm}$$

For the sample subjected to 37 MPa confining pressure, the residual and peak strengths are

$$\sigma_r = 54 \text{ MPa}$$

$$\sigma_p = 78 \text{ MPa}$$

and the analogous integral according to Figure 7B is,

$$\oint [\tau_{xx} - \tau_r] d\tau = G_{IIc} = \frac{1-\nu}{2G} K_{IIc}^2 = 3.36 \text{ MPa mm} \quad (0.00336 \text{ MPa m}). \quad (4b)$$

Again, we can express this result in terms of a characteristic displacement $\bar{\tau}$,

$$(\tau_p - \tau_r) \bar{\tau} = \oint [\tau_{xx} - \tau_r] d\tau = 0.00336 \text{ MPa m}$$

so

$$(\tau_p - \tau_r) \bar{\tau} = 0.00336 \text{ MPa m}$$

which gives a characteristic displacement roughly equal to the grain size (0.15 to 0.20 mm),

$$\bar{\tau} = 0.14 \text{ mm}$$

The relative brittleness of Berea Sandstone, as compared to overconsolidated London clays is suggested by comparing the values we obtained, 0.07 mm and 0.14 mm, with that obtained by Palmer and Rice (1973). They report that shear tests of overconsolidated clays had characteristic displacement values of 2 to 10 mm, which are between one and two orders of magnitude larger than those for Berea Sandstone.

In order to proceed, we must consider the boundary conditions of the body of rock that is going to contain the deformation band.

FINITE DEFORMATION BAND IN INFINITE MEDIUM

The simplest case we can consider is a doubly-terminated, pupative deformation band of length $2a$ surrounded, essentially, by an infinite body of rock subjected to far-field shear stress, τ_{21} , parallel to the pupative deformation band. The mode II stress intensity factor for such cases is (e.g., Lawn and Wilshaw, 1975, p. 58)

$$K_{II} = (\tau_{21} - \tau_r) \sqrt{a} \quad (5a)$$

in which τ_r is the residual strength of material within the deformation band. The energy release rate is,

$$G_{II} = \frac{(1-\nu)}{2G} K_{II}^2 = \frac{(1-\nu) \tau_{21} (\tau_{21} - \tau_r)^2}{2G} \quad (5b)$$

so, according to eq. (3a), the critical half length, a_c , for propagation of the fault can be calculated from,

$$\frac{(1-\nu) a_c (\tau_{21} - \tau_r)^2}{2G} = \int_0^a (\tau_{21} - \tau_r) d\tau = G_{IIc}$$

$$a_c = \frac{2G}{(1-\nu) (\tau_{21} - \tau_r)^2} \int_0^a (\tau_{21} - \tau_r) d\tau \quad (5c)$$

We will evaluate this result for the Berea Sandstone at 12 and 37 MPa confining pressures. We pick these confining pressures because Bernabe and Brace (1990) indicate that single deformation bands formed in samples at confining pressures in this range.

For confining pressure of 12 MPa, the integral for a shift of at least 0.1 mm across the band is 1.99 kPa m. The shear modulus at this confining pressure is about 6 GPa, so

$$2G/(1-\nu) = 12 \text{ GPa} \quad (5d)$$

because we estimate Poisson's ratio to be quite small, e.g., $\nu = 0.1$ or smaller (e.g., Johnson, 1970).

Thus, if we define a factor, F ,

The factor F is

$$F = \frac{2G}{(1-\nu) (\tau_{21} - \tau_r)^2} \int_0^a (\tau_{21} - \tau_r) d\tau \quad (5e)$$

which, for this particular example, is

$$F = 7.59 \text{ G } 10^2 (\text{MPa})^2 \text{ m}$$

Then eq. (4c) becomes,

$$a_c = F/(\tau_{21} - \tau_r)^2 \quad (5f)$$

According to Figure 7, the applied shear stress, τ_{21} , can range up to 46 MPa and the stress difference, that is, the difference between the applied shear stress, τ_{21} , and the residual strength, τ_r , can range from 0 to 26.6 MPa for Berea Sandstone under these conditions. Table 1 presents some values for critical flaw lengths. The critical flaw length ranges from 7 m for a stress difference of 1 MPa (shear stress only slightly greater than the residual strength), to a flaw length of 1 cm for a stress difference of 27 MPa (far-field shear stress about same as peak strength).

Table 1. Critical flaw lengths for Berea Sandstone

Critical Flaw Length	Applied shear stress	Stress difference
(m)	τ_{21} (MPa)	$\tau_{21} - \tau_r$
Confining Pressure= 12 MPa		
6.8	20	1
0.3	24	5
0.04	33	13
0.012	45	25
0.011	46	27
Confining Pressure= 37 MPa		
2.93	57	2
0.77	59	5
0.11	67	13
0.05	73	18
0.03	78	24

The results of the calculations are consistent, it seems, with the stress-strain curve for Berea Sandstone under a confining pressure of 12 MPa (Figure 3 and Figure 16). Note that, for applied shear stresses less than the peak value of 46 MPa, the flaws in the sample would have to be relatively large compared to the size of the sample. The samples were 1.6 cm in diameter and 3.8 cm in length. The critical flaw length is 4 cm for an applied shear stress of 13 MPa, so the sample should not produce a deformation band at such a low load.

Indeed, one might wonder whether the critical flaw size of 1 cm, which corresponds to the peak strength, there could be a flaw so large in such a small sample. On the other hand, one must bear in mind that the solution is for a flaw in an infinitely large sample, so we cannot answer such a question with this analysis. We will reconsider this issue when we examine the solution for a flaw in a rock of finite size.

Figure 10. Relation between shear stress and displacement for Berea Sandstone, at a confining pressure of 12 MPa. Peak and residual strengths are noted. Areas used in the computation shown with different symbols.

For confining pressure of 37 MPa, the integral for a shift of at least 0.1 mm across the band is 3.29 kPa m. The term containing the shear modulus at this confining pressure is,

$$2G/(1-\nu) = 16 \text{ GPa} \quad (5g)$$

Thus, using the definition of factor, F , in eq. (4c),

the factor in this case is

$$F = 17.13 \times 10^2 (\text{MPa})^2 \text{ m}$$

The critical length of pupative deformation band is

$$a_c = F/(\nu_{21} - \nu_r)^2$$

values of which are given in Table 1.

ROCK BODY OF FINITE SIZE IN RIGID LOADING FRAME

The analysis of growth of a deformation band in an infinite medium results in the critical length of band that will grow unstably. Now we will consider an analysis of rock confined by rigid blocks separated a finite distance and ask how thick the rock body must be for the deformation band to propagate unstably. This problem, also, was analyzed by Palmer and Rice (1990).

The deformation band is contained within an elastic body, the upper and lower boundaries of which are rigid and therefore subjected to uniform displacement conditions (Figure 6). The

thickness of material containing the band is h and the displacement of the upper rigid boundary relative to the lower is ${}_o u$.

Shift Across the Band

We can express the shift across the deformation band in terms of the stresses applied to the rock in front of and over the band. We assume that the material out away from the tip region of the band, along the dashed path, $?$, is elastic. Then the shear strain, ${}_o ?_{12}$, in material in front of the fault, beyond the influence of the singularity, is

$${}_o ?_{12} = (1/2)({}_o u/h) \quad (6a)$$

and the shear stress is,

$${}_o ?_{12} = 2G {}_o ?_{12} = G({}_o u/h) \quad (6b)$$

in which G is the elasticity shear modulus.

The material above the band, far away from the tip, but also far from the left-hand boundary will be subjected to a strain,

$${}_r ?_{12} = (1/2)({}_o u/h) - (1/2)(?_p/h) = (1/2)({}_o u - ?_p)/h \quad (7a)$$

so the shear stress in the material over the fault is

$${}_r ?_{12} = G({}_o u - ?_p)/h \quad (7b)$$

It is smaller than the stress in front of the band, eq. (6b).

Subtracting eqs. (7a) and (6a) we obtain an expression for the shift on the band in terms of the stress difference in front of and over the band,

$$?_p = (h/G)({}_o ?_{12} - {}_r ?_{12}) \quad (8)$$

The deformation band initiates at a flaw at the left boundary of the rock (Figure 6), and has propagated part way through the length of the sample in response to displacement of the upper rigid block with respect to the lower. We can use the J-integral to determine whether the deformation band can continue propagating, provided the sample is much longer than the deformation band, the thickness is much smaller than the length of the block of rock, and the end of the deformation band is far from either end of the block. Under these conditions the length of the block can be considered to be infinite. As a result, h is the only significant dimension.

Critical Stress Intensity Factor

We will assume that eq. (3c) can be derived, with a high degree of approximation, for a narrow deformation band,

$$J_{P \rightarrow P} = \int_{P \rightarrow P} \sigma_r d\sigma = G_{IIc} = \frac{1-\nu}{2G} K_{IIc}^2 \quad (9)$$

For Berea Sandstone subjected to 12 MPa confining pressure (Figure 7A) the traction on the band, σ_r , ranges from the residual value of about 19 MPa to a peak value of 46 MPa. So the integral is

$$\int_{P \rightarrow P} \sigma_r d\sigma = G_{IIc} = \frac{1-\nu}{2G} K_{IIc}^2 = 1.99 \text{ MPa mm (0.00199 MPa m)}. \quad (10a)$$

For the sample subjected to 37 MPa confining pressure, the residual and peak strengths are $\sigma_r = 54$ MPa and $\sigma_p = 78$ MPa so the analogous integral is,

$$\int_{P \rightarrow P} \sigma_r d\sigma = G_{IIc} = \frac{1-\nu}{2G} K_{IIc}^2 = 3.36 \text{ MPa mm (0.00336 MPa m)}. \quad (10b)$$

The J integral is path independent, so we can compute another expression for the J integral by integrating along the path designated γ in Figure 6. The increment of distance along the path of integration is ds . As shown in Figure 6, the integration path begins at P^- , passes through the elastic material below the deformation band, along the lower rigid boundary, then across the elastic material in front of the band, along the upper rigid boundary and then back through elastic material to point P^+ on the opposite side of the band.

For this problem, the J integral along the path γ increases with increasing applied load. When it reaches the critical value given by eq. (21), the band becomes active and propagates if the load is increased further.

The J integral in general is eq. (1c),

$$J_P = \int_{\gamma} [(W - f_i u_i) dx_2 - T_i \frac{\partial u_i}{\partial x_1} ds] \quad (11)$$

The integrand in eq. (11) vanishes along both the rigid boundaries because dx_2 vanishes there. The rigid overburden and the elastic substratum are bonded together, so $\partial u_i / \partial x_1$ also vanishes at the rigid boundaries. Finally, far from the tips of the deformation band the derivative $\partial u_i / \partial x_2$ is nonzero but the derivatives, $\partial u_i / \partial x_1$, that is $\partial u_1 / \partial x_1$ and $\partial u_2 / \partial x_1$, vanish because the material is subjected to uniform simple shearing. For these reasons the expression for the J integral for the path shown in Figure 6 becomes,

$$J_P = \int_{\gamma} W dx_2 \quad (12a)$$

The shear strain is constant, ϵ_{21} , in front of the band and a different constant, ϵ'_{21} , behind the tip of the band. Therefore, integrating along the path gives the result,

$$J_P = W(\epsilon_{21})h - W(\epsilon'_{21})h \quad (12b)$$

The negative sign in eq. (12b) is a result of the normal being in one direction for the path over and under the fault, and in the opposite direction in front of the fault.

For any strain, γ_{pq} , experienced by the rock, the stress work integral is defined as

$$W(\gamma_{pq}) = \int_0^{\gamma_{pq}} \tau_{ij} d\gamma_{ij} \quad (13a)$$

Using this definition, we can determine the value of the J integral in eq. (12b). The value of eq. (13a) has the two values,

$$W(o\gamma_{12}) = 2 \int_0^{o\gamma_{21}} \tau_{21} d\gamma_{21} \quad (13b)$$

$$W(r\gamma_{12}) = 2 \int_0^{r\gamma_{21}} \tau_{21} d\gamma_{21} \quad (13c)$$

If the material is linear elastic for this range of strains, then

$$d\gamma_{21}/2G = d\tau_{21} \quad (13d)$$

and

$$W(o\gamma_{12}) = (1/G) \int_0^{o\gamma_{21}} \tau_{21} d\gamma_{21} = \frac{o^2 \gamma_{21}^2}{2G} \quad (14a)$$

$$W(r\gamma_{12}) = (1/G) \int_0^{r\gamma_{21}} \tau_{21} d\gamma_{21} = \frac{r^2 \gamma_{21}^2}{2G} \quad (14b)$$

so that the J integral in eq. (23c) becomes

$$J_p = \int_p W dx_2 = (h/2G) [o^2 \gamma_{21}^2 - r^2 \gamma_{21}^2] \quad (15)$$

The critical stress intensity factor, though, eq. (15), is written in terms of the J integral minus the residual strength times the displacement at the arbitrary point p,

$$J_{P-r} = \frac{1}{2} \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] G_{IIc} = \frac{1}{2} K_{IIc}^2$$

Therefore eq. (15) becomes,

$$J_{P-r} = (h/2G) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] - J_{P-r} \quad (16a)$$

However, according to eq. (27),

$$J_P = (h/G) \left(\frac{1}{2} - \frac{1}{2} \right) = (h/G) \left(\frac{1}{2} - \frac{1}{2} \right) \quad (16b)$$

so eq. (16a) becomes,

$$J_{P-r} = (h/2G) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] - (h/G) \left(\frac{1}{2} - \frac{1}{2} \right)$$

or,

$$J_{P-r} = (h/2G) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right]^2 \quad (16c)$$

Critical Thickness of Test Specimen

Equation (16c) is the driving-force term, whereas eq. (9)

$$J_{P-r} = \frac{1}{2} \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] G_{IIc} = \frac{1}{2} K_{IIc}^2$$

is the resisting-force term. The driving force must equal or exceed the resisting force for a deformation bands to propagate, so we write,

$$(1/2G) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] \geq (1/h) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] G_{IIc} = (1/h) \frac{1}{2} K_{IIc}^2 \quad (17a)$$

Note that we can rewrite eq. (17a) in the form,

$$h \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] \geq (2G) \left[\frac{1}{2} \left(\frac{1}{2} \right) \right] \quad (17b)$$

or, in terms of the factor, F , defined in eq. (5e),

$$F = \frac{2G}{(1-2)} \left[\frac{1}{2} \left(\frac{1}{2} \right) \right]$$

$$h \geq \frac{(1-2)F}{\frac{1}{2} \left(\frac{1}{2} \right)}, \quad (18)$$

It is interesting that the critical thickness for a body of rock containing a pupative deformation band is about three times as large as the critical band length for an infinite body of rock. Table 1 shows values of the critical size of deformation bands which, when multiplied by λ , become values of the critical thickness of sample confined by rigid boundaries.

The results indicate that, for a sample of Berea Sandstone under a confining stress of 12 MPa the thickness of the sample must be at least 3.5 cm thick in order for the deformation band to grow unstably. For a confining pressure of 37 MPa, the sample must be about 10 cm thick for the deformation band to grow unstably.

These results are consistent with the experimental observations that individual bands form in samples subjected to confining pressure of 12 MPa, but multiple deformation bands form if the sample is confined at a pressure of 37 MPa. For larger confining pressures, the samples would have to be larger. The samples tested are too small to show individual deformation bands that grow unstably at the higher confining pressures.

Correction for Plastic Flow

The analyses of unstable growth of deformation bands in very large bodies and in bodies of finite size assume that the rock far to the right of the tip of the deformation band (Figure 6) and the rock far to the left of the tip is linear elastic. We made this implicit assumption when we wrote eq. (13c), for example.

In fact, though, as discussed in Appendix A, the stress-strain curves for Berea Sandstone are nonlinear and represent both elastic and plastic deformation. The energy density that drives growth of the deformation band is W , as we assumed in the derivation, but it is only part of W . W is different for the loading and unloading parts of the stress-strain curves for the sandstone. During loading, part of the energy density is elastic and therefore recoverable and part is plastic and therefore unrecoverable. It is only the recoverable energy density that can drive the deformation band forward, consuming energy provided by the unloading rock.

Unfortunately, Bernabe and Brace (1990), and Scott and Nielsen (1991), like most experimentalists, do not report elasticity properties during unloading. Therefore we can only assume that shear moduli for loading and unloading are the same. That is what we have done in previous pages.

Therefore, the energy density terms in eq. (13)

$$W(o, \gamma_{12}) = 2 \int_0^{\gamma_{21}} \tau_{21} d\gamma_{21}$$

$$W(r, \gamma_{12}) = 2 \int_0^{\gamma_{21}} \tau_{21} d\gamma_{21}$$

are integrated for the areas under the appropriate unloading curve, as shown in Figure 11B.

Figure 11. Energy densities. A. Sample loaded to σ_o , γ_o . The energy density is the area beneath the curve, but above the residual strength, σ_r . Much of energy density, though, is nonrecoverable. B. Unloading curve for sample. The energy density shown beneath the curve is the elastic component and is due to unloading. It is this energy density that is available to make the deformation band propagate unstably.

APPENDIX A. RHEOLOGICAL MODEL FOR POROUS SANDSTONES

The stress-strain relations typically reported in rock-testing literature are not as useful as they could be because the tester generally varies more than two variables at a time and therefore the effects of hidden variables may mask the behavior one wishes to study. The test results reported by

Scott and Nielsen (1991) unfortunately, fall into this category. The variables that they do report provide valuable information, of course, because they illustrate various phenomena. We wish that they would identify even more variables.

I had wanted this paper to be based on sound experimental measurements that would provide a theoretical basis for analyzing the propagation of deformation much as faults or fractures. Most of our understanding of these structures comes from field observations and from microscopic analysis of samples near or along deformation bands. It is always difficult to deduce properties, though, from the dead geologic record of a process; this is a frustration of the historical aspect of geologic studies. I had heard about experiments that had produced deformation bands, so I was excited to learn from the experiments.

Well.....the experimental studies turned out to be somewhat disappointing. I spent several weeks sifting through the records in these papers, trying to make sense of the diagrams and descriptions. The main frustration is that the experiments—like almost all experiments in rock mechanics—are done under conditions that make generalization almost impossible. In the end I could not make sense of the experimental results. The results of my studies are in Appendix A, along with some rather pointed comments about what, fundamentally, is wrong with such experiments. Essentially the problems are twofold: First, in order to measure rock properties, one must change one pair of invariants at a time. There are three strain invariants and three stress invariants. In standard experiments, all three invariants are changed at the same time. No wonder so little has been learned from experiments on rocks! Second, in order to correlate measurements of rock properties with observations of rock deformation and fracturing, step-by-step, as the rock is deformed. This requires one to study many samples, of course.

My own attempt at experimental rock deformation suffers from the first problem, but not the second (Peng and Johnson, 1972). At that time I did not know about invariants. They had never been mentioned in the literature I read at that time (I did not read the right literature). We did, though, dissect the rock samples in order to follow the development of fracturing. Most of our basic contributions—the fact that axial rupture is expected in properly tested uniaxial samples; that fact that cracks form a diagonal pattern of small beams that buckle and fail, producing a kink-like fault; the fact that the growth of microcracks in samples is largely controlled by the end conditions of the samples—were made because we dissected our samples and described what we saw and because we theoretically analyzed the stress state in the samples subjected to various loading conditions. Modern experimentalists generally present these conclusions as though they were always self-evident, but they have not advanced to the level of doing proper experimentation by changing invariants systematically.

STRESS INVARIANTS

The results shown in Figure 3, in the text, are not particularly useful, because they are not cast in terms of strain and strain invariants. In order to understand properties of isotropic materials, one needs to work with invariants. The first two stress invariants reflect the mean stress and the maximum shear stress (The maximum shear stress is half the value plotted on the vertical axis). The first two strain invariants reflect the volumetric strain and the maximum shear strain.

In general in continuum mechanics, one expresses mechanical properties of homogeneous media in terms of stress and strain invariants (Rivlin, 1948; Malvern, 1969). In terms of principal stresses, three invariants can be written in the form,

$$I_1 = \sigma_1 + \sigma_2 + \sigma_3 \quad (A1a)$$

$$II_2 = -(\sigma_1 \sigma_2 + \sigma_2 \sigma_3 + \sigma_3 \sigma_1) = (1/2)(\sigma_{ij} \sigma_{ij} - I_2^2) \quad (A1b)$$

$$III_2 = \sigma_1 \sigma_2 \sigma_3 \quad (A1c)$$

In this case we could define the mean stress, $\bar{\sigma}$, in terms of the first invariant,

$$\bar{\sigma} = I_1/3 = \frac{(\sigma_1 + \sigma_2 + \sigma_3)}{3} \quad (A2)$$

Alternatively, we can define the invariants in terms of *deviatoric stresses*, which we will identify with a prime. For example,

$$\sigma_1' = \sigma_1 - \bar{\sigma}$$

or

$$\sigma_1 = \sigma_1' + \bar{\sigma} \quad \sigma_2 = \sigma_2' + \bar{\sigma} \quad \sigma_3 = \sigma_3' + \bar{\sigma}$$

Substituting equations of the form of (A3) into eqs. (A1) we obtain an alternative form of the invariants,

$$I_2' = \sigma_1' + \sigma_2' + \sigma_3' = 0 \quad (A4a)$$

$$II_2' = \frac{1}{2} \sigma_{ij}' \sigma_{ij}'; \text{ (sum on repeated indices)} \quad (A4b)$$

Or, in terms of principal stresses

$$II_2' = \frac{1}{2}(\sigma_1'^2 + \sigma_2'^2 + \sigma_3'^2) \quad (A4c)$$

that is,

$$II_2' = \frac{1}{2}[(\sigma_1 - \bar{\sigma})^2 + (\sigma_2 - \bar{\sigma})^2 + (\sigma_3 - \bar{\sigma})^2] \quad (A4d)$$

$$III_2' = \quad (A4e)$$

We can define two invariants of the strains similarly (Johnson, 1979),

$$I_2 = \epsilon_1 + \epsilon_2 + \epsilon_3 \quad (A5a)$$

$$\Pi' = 2\tau_{ij}'\tau_{ij}' \quad (A5b)$$

Written in these forms, it is clear in eq. (A2) that the first invariant, eq. (A1a), is proportional to the mean normal stress (e.g., to the negative pressure) and the deviatoric second invariant, (eq. (A4d)), is related to the shear stresses applied to the body

RUDNICKI AND RICE GENERALIZATION OF TERZAGHI-COULOMB LAW

We follow Rudnicki and Rice (1975) in expressing a Terzaghi-Coulomb relation for elastic-plastic materials. An increment of deformation has an elastic component and a plastic component. Let a preceding superscript of e indicate elastic strain and p indicate plastic. Then an increment of elastic shearing can be expressed as

$$\tau^e = \frac{\tau}{G} \quad (A6a)$$

in which τ is given in eq. (A5b), G is the elasticity shear modulus and τ is the elastic shear strain invariant. An increment of the mean strain is proportional to the change in mean stress,

$$\tau^e = \frac{\tau}{K} \quad (A6b)$$

where K is the elasticity bulk modulus. Thus we have defined the elasticity of the material.

The plastic strain increments are,

$$\tau^p = \frac{\tau \tan(\phi)}{h} \quad (A6c)$$

$$\tau^p = \frac{\tau}{K} + \tau \tan(\phi)$$

or,

$$\tau^p = \frac{\tau}{K} + \frac{\tau \tan(\phi)}{h} \tan(\phi) \quad (A6d)$$

in which K is the plasticity bulk modulus (generally taken to be infinity), ϕ is the angle of internal friction h is the hardening modulus and $\tan(\phi)$ is the coefficient of dilatancy.

The total strains include elastic and inelastic components, so we sum pairs of eqs. (A6)

$$\tau = \frac{\tau}{G} + \frac{\tau \bar{\tau} \tan(\phi)}{h} \quad (A7a)$$

$$\tau \bar{\tau} \left[\frac{\tau}{K} + \frac{\tau}{P K} \right] + \frac{\tau \bar{\tau} \tan(\phi)}{h} \tan(\phi) \quad (A7b)$$

We can generalize these one-dimensional expressions in eqs. (A6) and (A7) by replacing τ , $\bar{\tau}$ and $\bar{\tau}$ with their invariant forms (Johnson, 1979; Aydin and Johnson, 1983, p. 24)

$$\tau = 2\sqrt{\tau_{ij}\tau_{ij}} = 2\sqrt{[(\tau_{xy}^2 + \tau_{yz}^2 + \tau_{zx}^2) - (\tau_{xx}\tau_{yy} + \tau_{yy}\tau_{zz} + \tau_{zz}\tau_{xx})]} = 2\sqrt{-(\tau_1\tau_2 + \tau_2\tau_3 + \tau_3\tau_1)} \quad (A8a)$$

The square root of half the second deviatoric invariant is a shear stress which we will designate with τ ,

$$\tau = \sqrt{\frac{1}{2}\tau'_{ij}\tau'_{ij}} = \sqrt{\frac{1}{2}[(\tau_1 - \bar{\tau})^2 + (\tau_2 - \bar{\tau})^2 + (\tau_3 - \bar{\tau})^2]} \quad (A8b)$$

$$\tau = \tau_1 + \tau_2 + \tau_3 \quad (A8c)$$

$$\bar{\tau} = \frac{(\tau_1 + \tau_2 + \tau_3)}{3} \quad (A8d)$$

Thus, eqs. (A7) and (A8) show how increments of invariants of stresses and are related to increments of invariants of strains for the generalized elastic/Coulomb model of Rudnicki and Rice (1975).

Rudnicki and Rice (1975) derive from these results general expressions for increments of deviatoric stress and strain and increments of deviatoric mean stress and volumetric strain,

$$\tau = \frac{\tau \tau'_{ij}}{G} + \frac{\tau \tau'_{ij}}{\tau} \quad (A9a)$$

$$\tau \tau_{kk} = \frac{\tau \bar{\tau}}{K_0} + \tan(\phi) \tau \quad (A9b)$$

in which

$$\tau = \frac{\tau'_{kl}\tau'_{kl}}{2\tau} \tan(\phi) \frac{\tau \bar{\tau}}{h} \quad (A9c)$$

$$K_0 = \frac{e K^P K}{(e K + P K)} \quad (A9d)$$

$$\sigma'_{ij} = \sigma_{ij} - \sigma_{ij}(1/3)\sigma_{kk} \quad (A9e)$$

$$\sigma'_{ij} = \sigma_{ij} - \sigma_{ij}(1/3)\sigma_{kk} \quad (A9e)$$

$$\sigma'_{ij} = \frac{1}{2} \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \quad (A9f)$$

Plane deformation

In solving a wide range of problems in geology we assume plane deformation, whether the rheological properties are elastic, viscous or plastic. Therefore, it will be convenient to show that eqs. (A11) become the familiar two-dimensional expressions. Note that, for plane deformation, we set one of the principal strains equal to zero, say ϵ_3 , and the corresponding principal stress equal to the mean stress. Thus, for plane deformation the stress invariants become, for mean stress (eq. (A8d)),

$$\bar{\sigma} = \frac{(\sigma_1 + \sigma_2 + \sigma_3)}{3}$$

$$\bar{\sigma} = \frac{(\sigma_1 + \sigma_2)}{2} = \frac{(\sigma_{xx} + \sigma_{yy})}{2} \quad (A10a)$$

For deviatoric stresses,

$$\sigma' = \sqrt{\frac{1}{2} \sigma'_{ij} \sigma'_{ij}} = \sqrt{\frac{\sigma'_{xx} - \sigma'_{yy}}{2}^2 + \sigma'^2_{xy}}$$

In terms of principal stresses,

$$\sigma' = \sqrt{\frac{1}{2} [(\sigma_1 - \bar{\sigma})^2 + (\sigma_2 - \bar{\sigma})^2]} = \frac{|\sigma_1 - \sigma_2|}{2} \quad (A10b)$$

Similarly, the strain invariants become

$$\epsilon = \epsilon_1 + \epsilon_2 \quad (A10c)$$

$$\epsilon' = 2\sqrt{-(\epsilon_1 \epsilon_2 + \epsilon_2 \epsilon_3 + \epsilon_3 \epsilon_1)}$$

$$\epsilon' = 2\sqrt{-\epsilon_1 \epsilon_2} \quad (A10d)$$

Therefore, the incremental stress-strain equations become,

$$\tau = \frac{\tau_1 + \tau_2}{2} \sqrt{\frac{1 - \tau_1 \tau_2}{G}} + \frac{\tau_1 - \tau_2}{2} \frac{(\tau_1 + \tau_2)}{h} \tan(\phi) \quad (A11a)$$

$$\tau(\tau_1 + \tau_2) = \frac{\tau_1 + \tau_2}{2} \left[\frac{\tau}{\epsilon_K + p_K} \right] + \frac{\tau_1 - \tau_2}{2} \frac{(\tau_1 + \tau_2)}{h} \tan(\phi) \quad (A11b)$$

Radial symmetry

The rheological equations that we will apply to experiments of radial symmetry (so-called "triaxial" tests) such as those performed by Bernabe and Brace (op.cit.) are eqs. (A7)

$$\tau = \frac{\tau}{G} + \frac{\tau \tau \tan(\phi)}{h} \quad (A12a)$$

$$\tau \tau \tau \tau \left[\frac{\tau}{\epsilon_K} + \frac{\tau}{p_K} \right] + \tau \tau \tan(\phi) \quad (A12b)$$

The invariants for triaxial tests are,

$$\tau = 2\sqrt{-(3\tau_a \tau_c + \tau_c^2)} \quad (A13a)$$

$$\tau = \sqrt{\frac{1}{3} |\tau_a - \tau_c|} \quad (A13b) \text{ aok}$$

$$\tau = \tau_a + 2\tau_c \quad (A13c)$$

$$\tau = \frac{(\tau_a + 2\tau_c)}{3} \quad (A13d)$$

We determine the constants by isolating the elasticity and plasticity moduli. An increment of shear is proportional to the shear stress,

$$\tau^e \tau = \frac{\tau \tau}{G}$$

so that,

$$G \tau = \frac{\tau \tau}{\tau^e \tau} \quad (A14a)$$

An increment of the volumetric strain is proportional to the change in mean stress,

$$\frac{\partial \epsilon}{\partial K}$$

so that the bulk modulus is,

$$\epsilon K \frac{\partial \epsilon}{\partial \epsilon} \quad (A14b)$$

where ϵK is the elasticity bulk modulus. Thus we have defined the elasticity of the material.

The plastic strain increments are given in eqs. (A6). Now that we know the elasticity moduli, we can use eqs. (A6) to compute the hardening modulus,

$$\frac{\partial \epsilon}{\partial \epsilon} = \frac{\partial \epsilon \tan(\epsilon)}{h} \quad (A14c)$$

If we assume that the bulk modulus for plastic flow is very large, the plastic volumetric strain is related to the coefficient of dilatancy,

$$\epsilon - \epsilon^p = (\epsilon \tan(\epsilon))$$

or,

$$\tan(\epsilon) = \frac{\epsilon \epsilon}{\epsilon - \epsilon^p} \quad (A14d)$$

This completes determination of the moduli. We will apply the results of these derivations to the deformation of Berea Sandstone.

DERIVATION OF COULOMB AND ELASTIC PROPERTIES FOR A POROUS SANDSTONE FROM LABORATORY TESTS

In order to determine the properties of the Berea Sandstone we need to relate the invariants as specified in the Coulomb-elastic model that we have tentatively adopted. Note that Figure 3 provides values for ϵ_a (axial strain), ϵ (volumetric strain), $\epsilon_a - \epsilon_c$, and ϵ_c . We convert these values to invariants in eqs. (A13). Since the volumetric strain and axial strain are specified, we compute the radial strain with the relation,

$$\epsilon_c = (\epsilon - \epsilon_a)/2 \quad (A15a)$$

Then, besides ϵ , we have data required to specify the invariants,

$$\tau = 2\sqrt{-(3\tau_a^2 + \tau_c^2)} \quad (A15b)$$

$$\tau = \sqrt{\frac{1}{3}|\tau_a - \tau_c|} \quad (A15c)$$

$$\tau = \frac{(\tau_a + 2\tau_c)}{3} \quad (A15d)$$

Figure A1 shows the stress-strain curves presented in terms of the shear strain and shear stress invariants. These curves are comparable to those in Figure 3A, except they show variations of basic variables. (Compare to Figure 3A, in main body of text). Figure A1 shows the stress-strain curves in terms of mean strain and mean stress invariants. We will see, though, that these curves, also, do not tell the whole story.

Examination of Figure A1 indicates that the volumetric deformation of the Berea Sandstone appears to be largely elastic, especially at higher confining pressures of 52 and 67 MPa. At the higher pressures the relations closely approach a straight line. There is significant deviation for the samples confined at pressures of 12 and 37 MPa. Both deviate when the mean stress minus the confining stress is about -30 MPa. The deviation is at a mean stress of about -42 MPa

Figure A1. Relations between a shear stress, τ , (the square root of the second deviatoric invariant of the stresses) and a shear strain, γ , (a similar measure of shear strain) for Berea Sandstone. Confining pressures range from 12 to 67 MPa. The slope of the relation between the first few points is the shear modulus, G , which ranges from about 5 to 10 GPa. (Data from Bernabe and Brace, 1990).

ELASTIC MODULI

The initial slopes of the stress-strain curves shown in Figure A1 should be proportional to the shear modulus according to eq. (A14a)

$$\gamma = G \epsilon \quad (A16a)$$

The values range from about 5 to 10 GPa. In terms of Young's modulus, E , and Poisson's ratio, ν , the shear modulus is

$$G = \frac{E}{2(1+\nu)} \quad (A16b)$$

The elasticity shear modulus appears to be relatively variable for the Berea Sandstone. It is derived from Figure A1A, and the relation,

$$\nu_s = 2G \epsilon_s \quad (A16c)$$

or

$$\frac{(\epsilon_s - \epsilon_c)}{2} = 2G \frac{(\epsilon_s - \epsilon_c)}{2} \quad (A16d)$$

The shear modulus, as a function of the maximum shear stress, is shown in Figure A2 for the various confining pressures. The diagram indicates that the maximum shear modulus increases with confining pressure from about 6 GPa at a pressure of 12 MPa, to about 11 GPa at a pressure of 52 MPa. The decrease of shear modulus with maximum shear stress presumably is a result of strain softening, an inelastic effect. Thus the maximum is probably the best estimate of the shear modulus that would be measured in unloading tests. Bernabe and Brace (1990) report values of 15 to 19 GPa for the same range of pressures; presumably they measured the shear modulus in unloading, which is quite proper.

Figure A2. Relations between shear modulus and shear stress for Berea Sandstone at various confining pressures. Relations between maximum shear stress and elasticity shear modulus at various confining pressures. The secant modulus is the slope of a line between a point on the stress-strain curve and the origin of coordinates. The peak values range from 5.5 GPa, with low confinement, to about 11 GPa for high confinement. The bulk modulus is 2 to 4 times as large as the shear modulus. The downward bending of the curves for higher shear stresses presumably is a result of plastic flow, so the shear moduli reported would be minimum values.

The slope of the stress-strain curves shown in Figure A3 is proportional to the bulk modulus as shown in eq. (A14b)

$$^eK = \frac{E}{3(1-2\nu)}$$

where eK is the elasticity bulk modulus. The values range from 15 to 30 GPa. The bulk modulus is variable, but the shear modulus appears to vary with confining pressure. In terms of Young's modulus and Poisson's ratio, the bulk modulus is,

$$^eK = \frac{E}{3(1-2\nu)}$$

Thus we have defined the elasticity of the material.

ANGLES OF INTERNAL FRICTION

We can estimate angles of internal friction for the peak stress and for the residual stress, defined as the stress corresponding to the flat stress-strain curve a large shearing (Figure A1). According to the Rudnicki and Rice model, eq. (A6c), the shear stress must satisfy the inequality,

$$\tau \geq \tau_p + \tau_r \tan(\phi) \quad (A17a)$$

Note that the mean stress is always negative in the experiments we are studying. For the peak and residual strengths, we assume that the shear stress is

$$\tau_p = \tau_p \tan(\phi_p) \quad (A17b)$$

$$\tau_r = \tau_r \tan(\phi_r) \quad (A17c)$$

Values of peak and residual friction angles determined with these relations are in a table in Figure A1. We see that generally the peak friction, ranges from 47 to 55° and the residual friction ranges from 46 to 50°. The peak friction is slightly higher than the residual friction. The difference, though, is quite small, ranging from $\phi_p - \phi_r = 1$ to 5°.

Perhaps we should assume that the peak and residual friction angles are the same. In order to investigate the strength properties further, the peak and residual shear and mean stresses in eqs. (A17b) and (A17c) are plotted in Figure A4. The slope of the curve is the tangent of the friction

angle. We see that the slopes are quite similar. Also, if the curves are projected to the origin, it appears that they pass through the origin of coordinates, indicating that the cohesive strength is negligible. Thus the strength is adequately described by either the peak or residual friction angle.

The conclusion that the peak and residual friction angles are nearly the same is confirmed by experiments and field analyses of landslides in overconsolidated and normally consolidated soils by Skempton (1964, 1984).

Figure A4. Mohr diagrams for peak and residual strengths in terms of mean stress and maximum shear stress. The slopes of the lines drawn through the maximum shear stress are nearly the same; about 32 to 33°. Thus the peak and residual friction angles are nearly the same. The shear strength intercepts, though, are higher for peak than for residual strengths. The slope of the line tangent to the Mohr circles decreases with increasing mean pressure, indicating that friction angle reduces with mean pressure. An approximation to a strain-line is about 34°, and the intercept is about 20 MPa.

A.

B.

Figure A5. Properties of Berea Sandstone. Data from Bernabe and Brace, 1990. A. Relations between mean stress and peak and residual strengths for Berea Sandstone at confining pressures

ranging from 12 to 67 MPa. The slope of the relations is related to the angle of internal friction. Note that the slope is roughly the same for peak and residual strength values, but that the difference between the peak and residual strengths is a result of a constant, which we consider to be like a cohesive strength. B. Relation between porosity and mean stress. The relation deviates markedly from a straight line (probably elastic) at certain values of mean stress and that the deviation is one of dilatancy. The porosity is essentially constant during "failure," marked by plotting together of several data points at the ends of the curves.

SIMULTANEOUS CHANGES OF INVARIANTS

A complication, however, with the experiments reported here is that the mean stress tracks the shear stress invariant, as shown in Figure A4. The magnitude of the mean stress is higher for the peak than for the residual strengths. The ratio of peak to residual strength is 2.5 and the ratio of mean stress is 1.9. Thus we would expect the friction values to be slightly higher for peak strength than for residual strength measurements as indicated in the table in Figure A1.

Figure A6. Relations between mean stress and volumetric strain invariants.

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Figure A7A Plot of relations between shear strength, as measured by the shear stress invariant, and the mean stress for Berea Sandstone. (Data from Bernabe and Brace, 1990). The upper curve is for peak strengths and the lower for residual strengths.

Figure A7B. Plots of mean stress (first) and shear stress (second) invariants relative to shear strain invariant, showing that the first and second stress invariants change together. This can confuse interpretations of the rheology of the rock. For example, if the shear strength is frictional, and proportional to the mean stress, the shear strength can change simply because the mean stress changes.

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Figure A8A Plot of relation between plastic shearing invariant and plastic volume change. The slope of the relations is proportional to the dilatancy of the material.

Figure A8B. Relations between the available shear stress, that is, the difference between the shear stress and the frictional strength, and the plastic shear strain invariants. The curves are invalid where they drop below zero because the rock is unloading. At zero, the shear stress invariant is equal to the frictional strength. The general slopes of these curves provide estimates of the hardening moduli which are listed in the table. They range from about -0.8 to -2.2 GPa. The negative values indicating that the rock is generally strain softening. These are to be compared to elasticity shear moduli of 5 to 10 GPa.

In fact, though, we cannot know what controls the variation. If other invariants, in particular the mean stress, had been held constant, the curves would have provided unambiguous relations. The confining pressure, which is *not* an invariant, was held constant, however, so the mean stress increased as the shear stress increased in the experiments. The upper curve in Figure A3 shows one of the shear stress-strain invariant plots (Figure A1; 12 MPa) and the lower curve shows the first invariant, the mean stress, as a function of the shear strain invariant.

This problem characterizes experimental rock mechanics. The investigators do not work with invariants. As a result, it is nearly impossible to generalize their results.

HARDENING (SOFTENING) MODULUS

Now that we know the elasticity moduli, we can use eq. (A14c) to compute the hardening modulus,

Figure A9. Relations between the available shear stress, that is, the difference between the shear

stress and the frictional strength, and the plastic shear strain invariants.

(A9 shows relations between the available shear stress, $\tau + \sigma_n \tan(\phi)$, and the plastic shear strain. According to eq. (A14), the slope of the relation is the hardening modulus,

$$h = \frac{d(\tau + \sigma_n \tan(\phi))}{dP_\gamma}$$

The curves in Figure A6 are invalid where they drop below zero because the rock is unloading. At zero, the shear stress invariant is equal to the frictional strength. The general slopes of these curves are estimates of the hardening moduli, which are listed in the table. They range from about -0.8 to -2.2 GPa. These are to be compared to elasticity shear moduli of 5 to 10 GPa. The negative values indicate that the rock is generally *strain softening*.

ANGLE OF DILATANCY

Using Rudnicki and Rice Definition

If we assume that the bulk modulus for plastic flow is very large, the plastic volumetric strain is related to the coefficient of dilatancy, which, according to eq. (A14d), is

$$\tan(\psi) = \frac{d\epsilon_v}{d\epsilon_p}$$

Figure A10 shows the relation between angle of dilatancy and total shear strain. The dilatancy angle ranges up to 30 to 40° for a total shear strain of about 0.25 at the lowest confining pressure

of 12 MPa. These high angles correspond to the peak of the shear stress-strain curve for a confining pressure of 12 MPa (Figure A3). The peak angle of dilatancy is about 10° and corresponds to a shear strain of about 0.275 at a confining pressure of 37 MPa. Figure A1 shows that, at this confining pressure, the angle of dilatancy is maximal at strains of 0.25 to 0.275 where the stress is maximal. At a confining pressure of 52 MPa, the angle of dilatancy is small, variable and is probably essentially zero.

Figure A10. Relation between angle of dilatancy and total shear strain. Values for small strains and highest confining pressure were excluded. At low confining pressure (12 MPa) the dilatancy angle ranges up to 30 to 40° for shear strain of about 0.25. At a confining pressure of 37 MPa, the peak angle of dilatancy is about 10° and corresponds to a shear strain of about 0.275.

Using Typical Ad-hoc Definition

It seems that, as geologists, of both the experimental and field varieties, we make every effort to take a perfectly good and well-defined concept and turn it into a half-truth or into a pure fairy tale.

The way that Barnebe and Brace (1980) and I (choke!) (Johnson, 1995) have defined dilatancy is clearly ad-hoc. In my own case, I know why I did it. It was my own stupidity. I cannot defend Barnebe and Brace. Surely they must have some other problem.

My focus (Johnson, 1995) was on the formation of shear zones, so I assumed that it was the dilation associated with the shearing related the change in thickness of the shear zone to the amount of shearing. This definition is fine, perhaps, but it is not the same as Rudnicki and Rice's (1975) definition. The only way my definition would apply would be if, somehow, the orientation of the shear zone were known, and shearing were concentrated on that zone. Well.....this is the kind of thinking that gets one into trouble in science. By looking at the dilatancy this way, there is no way that the dilatancy could cause the shear zone to be oriented in such a way. The orientation was already determined, by *assumption* (ugh).

If one wants to use the notion of dilatancy to predict the orientation of a shear zone, then one must use the definition, in terms of invariants, of the type that Rudnicki and Rice (1975) define. That way, the orientation is not forced by one's world view!

Bernabe and Brace (1990) calculated dilatancy in a similar, even more peculiar way. Rather than working with the rock as a homogenous body, they assumed that the deformation occurred along a band whose orientation was pre-determined. They assume that the band is inclined 30° to the axial stress. At another time and place, their assumption would be stated as an assumption that the angle of internal friction is 30° because they assume that the friction angle controls the orientation of the failure band.

They indicate that, in samples that failed along a single shear band, at low confining pressures, there was initially elastic deformation followed by strong dilation prior to failure. After failure, the deformation was accommodated by frictional sliding on the fault without further change of volume. In the transitional regime of confining pressures, the dilation decreased with increasing confining pressure. They report that the dilation totally vanished at a confining pressure of about 70 MPa⁴¹. At somewhat higher confining pressure, in the so-called ductile regime, only compaction occurred (op cit., p. 94).

ESTIMATION OF LOCALIZED DEFORMATION IN BERE A SANDSTONE

Figure 16A shows the shear stress-strain relations and Figure 16B shows the relation between angle of dilatancy, ϕ , and shear strain according to their calculations of stress and strain on the assumed band inclined 30° to the compression. The dilatancy in this case is different: it is a measure of permanent volume change, $d_p \phi_v$, with change in (absolute value of) permanent shear strain, $d_p |\phi_s|$ parallel to the band developing with an inclination of 30° to the direction of maximum compression. In any case, we assume a relation of the form

$$d_p \phi_v = 2 d_p |\phi_s| \tan \phi \quad (\text{A15a})$$

so that,

$$\tan \phi = d_p \phi_v / 2 d_p \phi_s \quad (\text{A15b})$$

Figure 16B shows that the angle of dilatancy is about 65° for a confining pressure of 12 MPa, decreases to about 40° for a confining pressure of 37, and 15° for a confining pressure of 52 MPa. At 67 MPa, and higher pressures, the angle is zero.

The relatively high angle of 60° for the angle of dilatancy agrees rather well with that deduced (60°) for conjugate, strike-slip deformation bands in Entrada Sandstone at Arches National Park and that deduced (40 to 60°) for oblique-slip, essentially normal faults, in the Goblin Valley area of San Rafael Desert (Johnson, 1995, p.196). One reason might be that the angle of dilatancy estimated by Johnson was defined the same way, eq. (A1), as that assumed by Bernabe and Brace (1990).

⁴¹ Equivalent to pressure at depth of about 3.5 km.

APPENDIX B. WONG'S MEASUREMENTS OF G_{IIc}

Wong's measurements and those of some other investigators reported in his paper are presented in Table III.

In calculating the Mode II fracture toughness, K_{IIc} , the Modulus of elasticity is taken as the middle of the range shown in Table V for a rock type with the same name and the Poisson ratio is taken to be the average. According to Table IV,

$$E \approx 50 \text{ GPa} \quad \nu \approx 0.1 \quad \text{granite}$$

$$E \approx 70 \text{ GPa} \quad \nu \approx 0.12 \quad \text{quartzite}$$

We know that, for *plane strain* the relation between G and K is

$$G_I = K_I^2(1-\nu^2)/E$$

$$G_{II} = K_{II}^2(1-\nu^2)/E$$

$$G_{III} = K_{III}^2(1+\nu)/E$$

and for *plane stress*,

$$G_I = K_I^2/E$$

$$G_{II} = K_{II}^2/E$$

$$G_{III} = K_{III}^2(1+\nu)/E$$

Since Poisson's ratio is so small, critical stress intensities are nearly the same for plane stress and plane strain. Thus,

$$K_{IIc} = \sqrt{\frac{G_{IIc} E}{(1-\nu^2)}}$$

The values of critical stress intensity, for Mode II failure, are shown in Table III. In general, the fracture toughness for Mode II is two or more times the fracture toughness for Mode I, as indicated in Tables III and V. In following pages we will estimate Mode II fracture toughnesses for the porous, Entrada Sandstone. For an individual band, which forms in rock with a porosity of 25% we estimate the fracture toughness to be $K_{IIc} = 2$ to $6 \text{ MPa m}^{1/2}$. For a narrow zone of bands, the zone forms within rock with about 5% porosity. We estimate the fracture toughness to be $K_{IIc} = 13$ to $53 \text{ MPa m}^{1/2}$.

Table III. Laboratory Measurements of Shear Fracture Energy, G_{IIc} and Shear Fracture Toughness, K_{IIc} .

Rock Type	G_{IIc} (J m ⁻²)	* K_{IIc} MPa m ^(1/2)	Comments
Fichtelgebirge granite	4.7×10^4	48	55 MPa confining pressure, Room temp
"	7.3×10^4	60	300 MPa confining pressure. Room temp.
Witwatersrand quartzite	0.8×10^4	24	14 MPa confining pressure. Room temp
"	1.3×10^4	30	28 MPa confining pressure. Room temp.
Westerly granite	0.7 to 0.8×10^4	19 to 20	Room conditions
"	0.3 to 0.8×10^4	12 to 20	20 to 80 MPa confining pressure, room temp
"	1.3 to 5.1×10^4	25 to 50	250 MPa confining pressure, temperatures from 160 to 668°C.
Entrada sandstone		2 to 6	About 25% porosity. Individual bands
"		13 to 53	About 5% porosity. Narrow zones of deformation bands.

(All but Entrada sandstone after Wong (1982), p. 992)

The values are reported in Table III. We have no measurements of Mode I fracture toughness for Entrada Sandstone, but the data presented by Atkinson and Meredith (1987) indicate that Mode I fracture toughnesses for a coarse-grained porous sandstone range from $K_{Ic} = 1.5$ MPa m^{1/2} for porosity of about 7% to $K_{Ic} = 0.81$ MPa m^{1/2} for porosity of 13%. Thus the Mode II fracture toughness may be about ten times the Mode I fracture toughness for the sandstone.

Table IV. Short version of Table I

Mode I Fracture Toughness of Some Rocks (Data from Atkinson, 1989).

Material	Fracture Toughness MPa m ^(1/2)	Comments
Berea ss	1.46	6.8% porosity
"	0.57	7.9% porosity

"	0.65	8% porosity
"	0.81	13% porosity
Nugget (Navajo) SS	0.22 to 0.34	
Chelmsford granite	0.50 to 0.7	Values in Table I
Bohus granite	0.66 to 2.03	western Sweden
Sierra white granite	0.79	
pink granite	1.66 to 2.03	
Westerly granite	0.6 to 2.17	
Oil shale	0.37 to 1.08	
shale	0.61 to 0.95	porosity 3.2 to 0.2%
clay shale	0.8 to 1.6	
siltstone	1.04 to 1.37	porosity 9.5 to 4.4%
Dry siltstone	0.73	

Table V Mechanical Properties of Some Earth Materials

Rock Type	Compressive Strength ^a (MPa)					Tensile Strength ^b (MPa)				Young's Modulus ^c (GP)		
	Max.	Min.	50% of data within		Max.	MMin.	Max.	MMin	Max.	Min.	Ave.	
BBasalt	360	81	?	?	?	112	49	0.42	-0.04	0.17		
DDiabase	360	160	280-350	?	?	107	72	0.28	0.10	0.26		
DDiorite	360	160	170-240	?	?	87	55	0.32	0.17	0.26		
DDolomite	360	62	?	?	?	93	49	0.18	-0.09	0.04		
GGneiss	250	150	170-240	?	?	70	30	0.15	0.04	0.10		
GGabbro						100	60	0.25	0.1			
GGranite	290	160	170-240	56	28	76	17	0.55	-0.30	0.04		
LLimestone	260	37	140-210	40	14	82	29	0.50	-0.37	0.17		
LLondon Clay (d)	?	?	?	?	?	?	0.15	0.01	?	?	?	?
MMarble	240	46	210-240	?	?	87	28	0.31	0.10	0.25		
MMarblestone	190	72	69-140	?	?	33	4.1	0.94	-0.30	0.26		

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QQuartzite	630	140	?	?	?	100	42	0.29	0.07	0.12
SSandstone	230	33	?	19	7	50	9.7	0.61	-0.51	0.02
SShale	230	76	69-140	?	?	52	12	0.92	-0.51	0.10
SSiltstone	320	34	?	?	?	44	13	?	?	0.12

^aAfter Obert and Duvall (6).

^bAfter U.S. Bur. mines Repts. Invests. 4459, 4727, 5130, and 5244.

^cAfter Obert and Duvall (6) and Clark (A2).

^dAfter Bishop et al. (1).

^e Some of data from Turcotte and Schubert, 1982, p. 432.

HOW KINK BANDS WORK AND HOW TO WORK WITH KINK BANDS (LONG EXPLANATION)

by

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Abstract

The sense of layer-parallel shear caused by regional shortening and emplacement of detachment sheets can be determined from the geometry of kink folds. In this paper we present the butterfly diagram of locking angles, which is a new method that uses locking angles of kink bands to determine the direction of maximum compression responsible for the formation of both monoclinical and conjugate kink bands, as well as the contact friction angle for conjugate kink bands. We also review the aspects of the theory of kink band formation that are essential to deduce information from kink folds observed experimentally or in the field. We analyze kink folds in two tectonic settings, using the butterfly diagram. The kink folds at these locations indicate the displacement direction on a regional décollement in eastern Alaska, and provide insight into the local stress state during Laramide deformation in central Utah.

INTRODUCTION

Analysis of kink folds can be instrumental for determining the displacement direction of faults or the local tectonic stress state (e.g. Kleist, 1972; Reches, 1978; Naruk et al., 1986; Gottschalk et al., 1998). Much experimental and theoretical work has been invested in attempts to understand how kink folds form (eg. Dewey, 1965; Weiss, 1968; Ghosh, 1968; Reches and Johnson, 1976; Johnson, 1977 and 1995; Pfaff, 1986; Pfaff and Johnson, 1989). The theory of kink band formation, however, has never been presented in a form useful to geologists who wish to interpret kink bands in the field. A method of analyzing kink folds with field measurements is presented in Johnson and Manuszak (in prep.), which uses a newly created *butterfly diagram* to determine the direction of maximum compression responsible for the formation of kink folds as well as determine the contact friction angle. In this paper we summarize the parts of the theory of kink-band formation essential to understanding the butterfly diagram and deducing information from field measurements. In addition we describe how conjugate and monoclinical kink folds form (Fig. 1). To illustrate the butterfly diagram, we provide two field examples from eastern Alaska and central Utah.

Fig. 1. Three variants of kink folds. All three types form under layer-parallel compression. The magnitude of the layer-parallel shear stress determines which variant will form. A. Symmetric conjugate kink fold. Contains identical kink bands facing left and right. Produced by zero layer-parallel shear. B. Asymmetric, conjugate kink fold. Left-lateral layer-parallel shear stress causes the asymmetry shown in the figure. C. Asymmetric, monoclinal kink fold. Larger left-lateral shear produces right-facing kink bands.

ASYMMETRY OF DRAG FOLDS AND KINK FOLDS

The asymmetry of folds is commonly used in structural geology to determine the sense of layer-parallel shear. For example, Pumpelly's rule (Wilson, 1960) states that the sense of asymmetry of parasitic folds can be used to infer the location of structural features of large folds. This rule is based on the assumption that right-facing asymmetric folds are produced by right-lateral layer-parallel shear and left-facing asymmetric folds are produced by left-lateral shear. As pointed out by Pfaff and Johnson (1989), Pumpelly's rule can be misleading because the asymmetry of monoclinal kink folds does not follow the rule of thumb for facing direction and sense of shear.

The central problem of interpreting the sense of shear indicated by asymmetric folds is the determination of the kind of fold one is studying. One must be able to recognize three kinds of asymmetric folds— drag (or parasitic) folds (Fig. 2A), asymmetric conjugate kink folds (Fig. 2B), and monoclinal kink folds (Fig. 2C). Although all these folds form under layer-parallel compression and layer-parallel shear stress, they form in different ways.

Kink Folds

A *kink band* is the fundamental building block of kink folds. It is a tabular shear zone that cuts across layered media such as bedding, foliation, crystal lattices or cleavage

planes⁴². It is characterized by a monocline with an inner limb in which the layering has been reoriented with respect to that in the two outer limbs (Fig. 3; Dewey, 1965, p. 460; Spencer, 1977, p. 51; Suppe, 1985, p. 336).

Two or more single kink bands may form under the same stress state to create a *kink fold*, of which there are three variants: symmetric conjugate, asymmetric conjugate, and monoclinical (Fig. 1). The type of kink fold that forms is dependent primarily on the amount of layer-parallel shear stress (Johnson, 1977). For example, symmetric conjugate kink folds (Fig. 1A) form under negligible layer-parallel shear stress while asymmetric conjugate kink folds (Fig. 1B) and monoclinical kink folds (Fig. 1C) form under increasingly greater amounts of shear stress respectively (Johnson, 1977). Both kink folds shown in Fig. 1 form under left-lateral shear stress, but if the sense of shear is reversed, the asymmetries of the fold forms are reversed as well (Fig. 2B,C). The essential facts about the sense of layer-parallel shear that assisted the layer-parallel compression in the formation of the idealized folds shown in Fig. 1 have been thoroughly established in the field and in experimental and theoretical investigations (e.g., Dewey, 1965; Reches and Johnson, 1976; Ramberg and Johnson, 1976; Pfaff and Johnson, 1989).

Drag Folds

Drag folds consist of a series of asymmetric plications with alternating short and long limbs and with axial planes that dip in the same direction as the long limb (Fig. 2A) (DeSitter, 1964, p. 279; Ramberg, 1963; Pfaff, 1986). Drag folds thus can be considered to be folds that formed by mechanical buckling of layers and then by kinematic simple shearing so that the asymmetry of the folds reflects the sense of shear (Ramberg and Johnson, 1976; Pfaff, 1986). Therefore the sense of layer-parallel shear responsible for drag (or parasitic) folds is easy to interpret: right-facing drag folds (Fig. 2A) are produced by right-lateral shear and left-facing drag folds are produced by left-lateral shear.

Interpreting Loading Conditions

Pfaff and Johnson (1989) addressed the problems of interpreting the loading conditions responsible for asymmetric folding. The problem is that one cannot apply the rule of thumb to all asymmetric folds that the facing direction of a fold indicates the direction of simple layer-parallel shear. The drag fold (Fig. 2A) and the asymmetric conjugate kink fold (Fig. 2B) have that sense of asymmetry. The monoclinical kink fold, however, has the opposite sense of asymmetry (Fig. 2C) from drag folds and asymmetric, conjugate kink folds.

To understand this opposite sense of asymmetry it is necessary to understand how kink bands form. We will review the theory of kink folding that is essential to understanding this process.

⁴² Kink bands have been formed experimentally by layer-parallel compression of foliated materials with a wide range of rheological properties (Donath, 1961; Turner and Weiss, 1965; Paterson and Weiss, 1966; Weiss, 1968; Ghosh, 1968; Johnson, 1970; Cobbold et al., 1971; Ellen, 1971; Gay and Weiss, 1974; Honea and Johnson, 1976; Reches and Johnson, 1976; Ramberg and Johnson, 1976).

Fig. 2. Asymmetric folds formed under conditions of layer-parallel shortening and right-lateral layer-parallel shear. A. Right-facing drag (or parasitic) fold. Steeper limb faces right. B. Asymmetric conjugate kink fold. Steeper limb faces right. C. Left-facing monoclinal kink fold. Steeper limb faces left.

HOW KINK BANDS INITIATE AND LOCK

The key difference between kink folding and drag folding lies in the strength of contacts or interbeds (Honea and Johnson, 1976; Reches and Johnson, 1976; Pfaff, 1986). Kink bands have been demonstrated to occur in multilayers in which strength of contacts (Honea and Johnson, 1976) or strength of interbeds (Ellen, 1971) must be overcome in order for the stiff layers to slip over one another (Honea and Johnson, 1976). Kink bands form when the strength is overcome locally and slippage occurs along inner limbs. Pfaff and Johnson (1989) have shown that even moderate nonlinear resistance to slip within interbeds is sufficient to cause kink folds to form. Drag folds form in layers where the strength at contacts or interbeds is relatively low and slippage occurs easily. Because of this relatively low friction between layers, drag folds typically form a wave train, while high friction between layers allows isolated kink bands to form.

Scenario of Formation of a Kink Band

To illustrate conditions that lead to the development of a kink band, we will relate the initiation, propagation and stabilization of a kink band to the loading conditions of a specific multilayer. The angle of friction between layers is assumed to be 30° for this example.

Initially, before loading, there is no kink band and the layering is planar, except for a small local perturbation (Fig. 3A). Then the multilayer is loaded, subjected to layer-parallel maximum compression, σ_1 , and constant layer-normal minimum compression, σ_3 . When the difference, $\sigma_1 - \sigma_3$, becomes sufficiently large, the kink band propagates through the multilayered material much as shown in Figs. 3B through 3D.

Fig. 3. Loading conditions during initiation, growth and locking of a kink band.

The propagation of the kink band can best be related to the principal stresses in terms of the stress ratio, σ_1/σ_3 . Figure 3E shows the stress ratio required for propagation of the kink band as a function of the orientation of the layering within the inner limb, θ , as first illustrated by Reches and Johnson (1976).

Suppose that the angle of the initial perturbation, θ_i , shown in Fig. 3A is 2° . As a result of the interaction of the layer-parallel compression and the local perturbation, θ_i , a shear stress is generated parallel to layering within the perturbation and may become large enough to cause localized slippage. If the layer-parallel compression increases (and thus the stress ratio, σ_1/σ_3 , increases since σ_3 is constant), the shear stress within the perturbation increases until the strength between layers is overcome. Kink band formation begins when the stress ratio reaches $\sigma_1/\sigma_3 = 22.5$ (point A, Fig. 3E) and the layers slip locally within the perturbation. The layering of the inner limb reorients unstably to a higher angle, θ (Point B, Fig. 3E). This process is unstable because the critical stress ratio (the ratio required for slippage between layers) at higher limb angles is actually

lower than the initial critical stress ratio, $\tau_1/\tau_3 = 22.5$. For example, the critical stress ratio drops almost 50 %, to about 12, as the angle of layering increases one degree, from 2° to 3° . As the limb angle increases to 30° the critical stress ratio decreases to its minimum value, $\tau_1/\tau_3 = 3$ (Point C, Fig. 3E). At this stress ratio the orientation of layering in the inner limb becomes stable because an increase in the limb angle requires an increase in the critical stress ratio.

The angle corresponding to point C is an example of a *locking angle* for a kink band (Reches and Johnson, 1976). If the stress ratio does not increase above the value for Point C (Fig. 3E), the layering within the kink band will lock at an angle of 30° (Fig. 3C). If the stress ratio increases, the kink band will lock at a higher angle. If the final stress ratio is equal to the initiation stress ratio, $\tau_1/\tau_3 = 22.5$, then the reoriented layering in the inner limb of our example locks at 58° (point D, Fig. 3E)

Conditions of Kink-band Formation

Now that we have explained the general features of the way a kink band initiates, becomes unstable and then stabilizes and locks, we will present the conditions and relationships that describe these phenomena. It has been demonstrated by others⁴³ that the necessary conditions for kink band formation are:

1. a multilayered material (bedded stiff layers or interbedded stiff and soft layers),
2. a nonlinear relation, such as cohesive or frictional strength, between shear stress and flexural slip at contacts of layers or within interbeds,
3. maximum principal compression inclined less than $(45-\theta/2)$ degrees to the layering in the outer limbs,
4. and an initial perturbation of layer orientation.

Some authors have used the term normal kink band or contractional kink band for structures formed under layer-normal compression (e.g. Ramsey, 1987; Kidan and Cosgrove, 1996). These structures are planar zones of localized shear deformation (Means and Williams, 1972; Kidan and Cosgrove, 1996) and should not be confused with the buckle folds we are discussing in this paper. We choose to exclude these structures as types of kink bands because, while they may resemble kink bands in shape, the mechanism of formation is different.

In order to define the quantities that control the formation of kink bands we show part of an idealized kink band in Fig. 4. The layering in the inner limb of the kink band is oriented at a counterclockwise (positive) angle, θ , to the layering in the outer limb. The maximum compression, σ_1 , is oriented at a counterclockwise angle ϕ to the layering in the outer limb, and the minimum compression, σ_3 , is oriented normal to σ_1 . The tractions acting on the contacts between layers within the inner limb of the kink band are σ_s (parallel to bedding) and σ_n (normal to layering). A fundamental assumption of the analysis of instability in the following paragraphs is that the stress state is the same within the inner and outer limbs (Johnson, 1995).

⁴³ A more nearly complete theory of kink-band formation is provided by Peng and Johnson, 1972; Honea and Johnson, 1976; Reches and Johnson, 1976; Johnson, 1977, 1995; Johnson and Pfaff, 1989; and Johnson and Fletcher, 1994.

Fig. 4. Geometry of idealized kink band and stress state.

If the strength of the contact between layers is specified by Terzaghi-Coulomb's law, then for localized slippage to occur, the shear traction, τ_s , must overcome the cohesive strength, C , and the frictional strength, $(\sigma_n - p_w)\tan(\phi)$. This is expressed mathematically as

$$|\tau_s| \geq C + (\sigma_n - p_w) \tan(\phi), \quad (1)$$

in which ϕ is the angle of internal friction (or contact friction), σ_n the normal traction across layer contacts, and p_w is pore-water pressure.

From the Terzaghi-Coulomb strength law in eq. (1), an expression for the ratio of principal stresses required to initiate slippage on the inner limb of a kink band or at an initial perturbation in layering can be derived from equations for equilibrium of forces on the element shown in Fig. 4:

$$\frac{\sigma_1}{\sigma_3} \geq \frac{1}{\sin(\theta - \beta) \cos(\theta + \beta + \phi)} \left[C \cos \phi + p_w \sin \phi \right] \cos(\theta - \beta) \sin(\theta + \beta + \phi) \quad (2a)$$

Note that the stress ratio in eq. (2a) is less than the term on the right-hand side if there is no slippage and is equal to the term on the right if there is slippage. Eq. (2a) shows that, besides the stress ratio, the variables that control the slippage within inner limbs of kink bands are the orientation of layering, $(\theta - \beta)$, within the inner limb relative to the direction of maximum compression and variables expressing the strength of contacts.

Many aspects of kink-band formation can be understood by separately considering cohesive or frictional strengths. By assuming frictional strength, that is the cohesion and pore-water pressure are zero, a simpler expression for the stress ratio is obtained from eq. (2a) (Reches and Johnson, 1976):

$$\frac{\sigma_1}{\sigma_3} \geq \frac{\tan(|\theta - \beta| + \phi)}{\tan|\theta - \beta|}, \quad C = 0; \quad p_w = 0. \quad (2b)$$

The stress ratio depends on the contact friction angle, ϕ , and the difference between the inner limb angle, β , and the inclination angle, θ , of the maximum compression to layering in the outer limbs (inset in Fig. 5A). Figure 5A is a plot of eq. (2b) for friction angles of $\phi = 10, 20, 30$ and 40° . The plot shows the stress ratio, σ_1/σ_3 , required for slip between layers to occur, as a function of the angle, $\theta - \beta$, between the maximum compression and the inner limb. For angles, $|\theta - \beta|$ equal to 0 or 90° , the critical stress

ratio required for slip between layers approaches infinity. The minimum critical stress ratio occurs at the bottom of the curve at inner limb angles of 45-???

Fig. 5. Relations between ratio of principal stresses that can be supported by strength of contacts or strength of interbeds (critical stress ratio) and the orientation, ???, of the maximum compression to the layering within the kink band. A. Frictional strength for angles of contact friction, ϕ , of 10 to 40°. Cohesive strength assumed to be zero. Minima shift toward lower angles as friction angle increases, indicating that the orientation of the kink limb will tend to be lower for high friction. B. Zero friction. Only cohesive strength. Minimum is always at ??? 45°.

As another extreme, we assume that the strength is purely cohesive where either the contact friction angle, ϕ , is zero or the pore-water pressure is equal to the normal traction acting on the potential slip surface. In the latter case we say that the *effective friction angle* is zero, and the strength of contacts or interbeds is purely cohesive (Lambe and Whitman, 1969).

For purely cohesive strength, eq. (2a) simplifies to the expression:

$$\frac{\tau_1}{\tau_3} = 1 + \frac{2C}{\tau_3 \sin(2|\theta - \phi|)}, \quad p_w = \tau_n, \quad (2c)$$

which depends on the difference between the inner limb angle, θ , and the inclination angle, ϕ , of the maximum compression and on the cohesive strength ratio, C/τ_3 . Figure 5B is a plot of eq. (2c) for cohesive strength ratios, C/τ_3 , of 1, 3, 5, and 10. The plot gives the stress ratio, τ_1/τ_3 , required for slip between layers, as a function of the angle, $|\theta - \phi|$. For angles of $|\theta - \phi| = 0$ or 90° , the stress ratio required for slip between layers approaches a maximum, which is infinity. The minimum critical stress ratio occurs at $|\theta - \phi| = 45^\circ$.

Ultimate Locking Angles of Kink Bands

Previously we noted that the locking angle of a kink band, θ_L , is the inner limb angle corresponding to the final stress ratio. In the example shown in Fig. 3E, the locking angle has a limit of 60° . That is, as the stress ratio approaches infinity, the locking angle approaches 60° . We refer to the locking angle at infinite stress ratio as the *ultimate locking angle*, θ_{UL} . The stress ratio in eq. (2a) is infinity if the denominator, of the coefficient is zero, or when $\cos(|\theta - \phi|) = 0$. The relevant solution is a simple expression for the (positive) ultimate locking angle of layers within the inner limb (Reches and Johnson, 1976). For left-facing kink bands,

$$\theta_{UL} = 90^\circ - \theta + \phi; \quad \theta < 0. \quad (3a)$$

For right-facing kink bands,

$$\theta_{UL} = -(90^\circ - \theta + \phi); \quad \theta > 0. \quad (3b)$$

We see that the ultimate locking angle is determined by only two variables, the contact friction angle, ϕ , and the angle of inclination, θ , of the maximum compression. We should note here that eq. (3) is valid for conditions of both purely cohesive strength ($C = 0$) and purely frictional strength ($C = 0$).

The ultimate locking angle is a useful concept for interpreting a kink band because the locking angle is nearly equal to the ultimate locking angle for a wide range of stress ratios. For the example shown in Fig. 3E, with $\theta = 30^\circ$, the ultimate locking angle, corresponding to a very large stress ratio, is $\theta_{UL} = 60^\circ$ according to eq. (3a). Even at a stress ratio of $\tau_1/\tau_3 = 7.5$ the locking angle is $\theta_L = 55^\circ$. Thus we can see from our example that the locking angle of a kink band will be reasonably close to the ultimate locking angle, regardless of the final stress ratio.

CONDITIONS NECESSARY TO FORM THE THREE TYPES OF KINK FOLDS

The inclination, θ , of maximum compression determines whether kink folds form as a set of monoclinical bands or as a set of conjugate kink bands (Fig. 6). Gay and Weiss (1974) first recognized this relationship from experiments. We can understand this relationship with the theoretical work of Reches and Johnson (1976). When the orientation of the maximum compression (θ_1 in Fig. 6) is parallel to layering ($\theta = 0$), symmetric conjugate kink bands form with locking angles of the left-facing limb, θ_{Ll} , and the right-facing limb, θ_{Lr} , equal in magnitude, $|\theta_{Ll}| = |\theta_{Lr}|$ (Fig. 6A). When the maximum

compression is slightly inclined to the layering in a clockwise or counterclockwise sense, kink bands may form as asymmetric conjugate kink bands where $|\phi_{L1}| \neq |\phi_{L2}|$ and $\phi \neq 0$ (Fig. 6B). When the maximum compression is more highly inclined to the layering, kink bands may form as monoclinic kink folds (Fig. 6C), with either left-facing or right-facing limbs ($\phi < 0$ for left-facing kink bands). Finally, when ϕ reaches a critical angle ($\phi \approx 45^\circ/2$, where ϕ = the contact friction angle; Reches and Johnson, 1976) kink bands will no longer form and shortening will be accommodated by general layer-parallel slip along all of the beds.

Fig. 6: The three different types of kink bands and the inclination angle, φ , of the maximum compression responsible for each type. The locking angle, φ_L , is negative for right-facing kink bands and positive for left-facing kink bands.

Fig. 7. Stress-state domains of initiation of monoclinal and conjugate kink folds. The domain is determined by angle, θ , between maximum compression and layering in outer limbs, and by the angles of initial left-facing perturbations, θ_L , and right-facing perturbations, θ_R . The inner domain includes conjugate kink folds for $\theta_L \leq \theta \leq \theta_R$. The folding is symmetric for $\theta = 0^\circ$ and asymmetric otherwise. In the example shown the maximum compression is inclined in a counterclockwise direction, $\theta = 10^\circ$, and the perturbations are $\theta_R = -8^\circ$ and $\theta_L = 13^\circ$. As indicated in the figure, the orientation of the maximum compression falls within the range, θ_R to θ_L so that the perturbations are expected to grow into asymmetric, conjugate kink folds.

The relations between type of kink fold and orientation of the direction of maximum compression in comparison to the orientations of perturbations are summarized in Fig. 7. The general orientation of layering outside the limb of the kink band is indicated with the heavy horizontal line. The maximum compression, θ_1 , is inclined at an angle (positive counterclockwise), θ , to the layering. There are three shaded pie-shaped areas that indicate ranges of orientations of maximum compression corresponding to different types of kink folds. The middle pie-shaped area, "asymmetric and symmetric conjugate kink folds," extends on either side of the orientation of the general layering. The boundaries of this pie-shaped area are defined by the orientations of right-facing, θ_R , and left-facing, θ_L , initial perturbations that might develop into kink bands.

The other two pie-shaped areas are between the lines representing the left- or right-facing perturbations and the limits of θ , $\theta = (45^\circ - \theta/2)$. These pie-shaped areas indicate the limits of the orientation of the maximum compression if monoclinal kink bands are to form. When θ equals $\theta = (45^\circ - \theta/2)$, there is general shearing of the layers in the outer limbs and the shearing instability which is fundamental to kink folding disappears (Johnson and Reches, 1976).

Symmetric conjugate kink folds

To illustrate the conditions necessary for conjugate kink bands to form, we consider the case of a multilayer with frictional contacts ($C = 0$ and $\theta = 30^\circ$) subjected to horizontal compression with an initial left-facing perturbation, $\theta_L = 3^\circ$, and an initial right-facing perturbation, $\theta_R = -3^\circ$. Then from Fig. 8A, with $\theta = 0$, it can be seen that a stress ratio of about $\sigma_1/\sigma_3 = 15$ can initiate a kink band at both the left- and right-facing perturbations. If the stress ratio reaches 15, and is maintained until the perturbations grow into kink bands and lock, the absolute values of the locking angles for the right and left facing limbs will both be 57° . This is how the symmetric conjugate kink band shown in Fig. 8B forms. If the stress ratio becomes very large, the locking angle approaches the ultimate locking angle, $\theta_{UL} = 60^\circ$.

The stress state during initiation shown in Fig. 8C indicates that the perturbations are expected to grow into symmetric conjugate kink bands.

Asymmetric conjugate kink folds

As a second example, assume similar conditions $C = 0$, $\theta = 30^\circ$, $\theta_L = 3^\circ$, and $\theta_R = -3^\circ$ (Fig. 9A). Asymmetric conjugate kink bands will form if the maximum compression is (positively) inclined with an angle $\theta = 2^\circ$. The stress ratio of 39 required to initiate slippage on this perturbation, when $\theta_L - \theta = 1^\circ$, is indicated by point f , shown in the right-hand side of Fig. 9A.

Fig. 8. Conditions responsible for symmetric conjugate kink folds. Layers subjected to layer-parallel compression. A. Stress-orientation paths for right- and left-facing perturbations that are growing into kink bands. Right- and left-facing kink bands initiate at a stress ratio of about 15. The stress ratio drops and then rises again as the kink band locks. B. Symmetric conjugate kink fold formed under conditions assumed in A. C. Direction of maximum compression is parallel to layering in outer limb. A slightly larger inclination, θ , would cause the conjugate kink fold to be asymmetric.

Fig. 9. Conditions responsible for asymmetric conjugate kink folds. Layers subjected to layer-parallel compression and left-lateral, layer-parallel shear. A. Stress-orientation paths for right- and left-facing perturbations that are growing into kink bands. Right-facing kink band initiates first, at a stress ratio of about 7. The stress ratio drops and then rises again as the kink band locks. If the stress ratio rises to a high enough level, the left-facing perturbation grows into a kink band. B. Asymmetric conjugate kink fold formed under conditions assumed in A. C. Direction of maximum compression lies between limits of initial left-facing, θ_L , and right-facing, θ_R , perturbations. A slightly larger inclination, θ , would cause the kink fold to be monoclinial.

The magnitude of the inclination angle of the right-facing perturbation is much larger, $\theta_{ri} - \theta = -5^\circ$, and the stress ratio required to initiate slippage is much smaller, about 7.

Thus, the stress ratio on the layers with perturbations increases to a value of about 7 (point a, Fig. 9A). If that stress ratio is maintained, the right-facing limb locks at $\theta_{Lr} - \theta = -55^\circ$. As the stress ratio increases along the path b-c-d-e to about 40, the right-facing limb changes stably to larger angles, reaching about $\theta_{Lr} - \theta = -60^\circ$. At that stress ratio the left-facing perturbation becomes unstable, and the left-facing kink band forms. The right-

facing kink band unloads, but its orientation remains because the deformation is permanent (path *f-g*, Fig. 9A).

For a large final stress ratio the locking angles will be reasonably close to the ultimate locking angles. The left-facing kink band will therefore lock at $\theta_L = 90 - \theta + \theta = 63^\circ$ (Eq. 3). The right-facing kink band will lock at $\theta_{Lr} = -(90 - \theta + \theta) = -57^\circ$. This is how the asymmetric conjugate kink band shown in Fig. 9B forms.

The stress state during initiation shown in Fig. 9C indicates that the perturbations are expected to grow into asymmetric conjugate kink bands.

Monoclinal Kink Folds

We have found based on our field observations that monoclinal kink folds occur more commonly than conjugate kink folds. We will provide an example of conditions that would create a monoclinal kink band, and in the process, see why they would form preferentially.

For this third example, assume similar conditions as for the symmetric folding $C = 0$, $\theta = 30^\circ$, $\theta_i = 2^\circ$ and $\theta_{ri} = -2^\circ$. The compression is inclined with a larger angle of $\theta = +4^\circ$. Figure 10C indicates that such perturbations will tend to produce right-facing, monoclinal kink folds (Fig. 10B). The inclination angle of the left-facing perturbation, $\theta_L = 2^\circ$, is smaller than the angle of the compression, $\theta = 4^\circ$, so, the maximum compression is inclined at a higher (4°) to the layering outside of the perturbation than to the perturbation (2°). The shear stress acting on the left facing perturbation is therefore smaller than the shear stress acting on the layering outside the perturbation, and so localized slip will not occur on the left facing perturbation and a left facing kink band will not form. The right-facing perturbation can form at the critical stress ratio, $\theta_1/\theta_3 = 7.5$ (Fig. 10A). Thus, suppose that the stress ratio, θ_1/θ_3 , is increased to 7.5. The right-facing (R) perturbation grows into a band and tightens as the stress ratio increases back to its original value of 7.5 (R_{lock} , Fig. 10A). The layering within the kink band has an orientation, $(\theta_L - \theta)$, of -54° so that the locking angle, θ_L , between the layering in the outer and inner limbs is $\theta_L - \theta = 54^\circ + \theta = -50^\circ$.

We are thus able to understand why monoclinal kink bands might form preferentially in multilayers with the appropriate properties and subjected to layer-parallel compression and layer-parallel shear. In order for both facing directions to form, the size of one of the perturbations must be much larger than the other perturbation. Both left-facing and right-facing kink bands form and the fold is an asymmetric conjugate kink fold if the perturbations are large enough. Otherwise the kink fold will be monoclinal. In right-lateral, layer-parallel shearing the right-facing perturbation would have to be much larger than the left-facing perturbation. In left-lateral layer-

Fig. 10. Conditions responsible for monoclinal kink folds. Layers subjected to layer-parallel compression and left-lateral, layer-parallel shear. A. Stress-orientation paths for right-facing perturbations that are growing into kink bands. Right-facing kink band initiates at a stress ratio of about 7. The stress ratio drops and then rises again as the kink band locks. B. Monoclinal kink fold formed under conditions assumed in A. C. Direction of maximum compression lies outside limits of initial left-facing, θ_L , and right-facing, θ_R , perturbations, in the range of right-facing monoclinal kink folds.

parallel shearing the left-facing perturbation would have to be much larger than the right-facing perturbation.

HOW TO WORK WITH KINK BANDS AND KINK FOLDS IN THE FIELD

The Butterfly Diagram of Ultimate Locking Angles

When working with kink bands in the field, the single observation we can make is the locking angle. With the locking angle of a monoclinal kink fold and an estimate of the friction angle, we can obtain θ . We will also show that in the case of conjugate kink bands, the friction angle does not have to be estimated because we can calculate it.

We have developed the *butterfly diagram*, shown in Fig. 11, which is used to determine the inclination, θ , of maximum compression and the contact friction angle, ϕ , with measurements of locking angles (Johnson and Manusak, in prep.). The butterfly diagram is a plot of eqs. (3), but with necessary limiting conditions. It plots θ as a function of the ultimate locking angle, θ_{UL} , which we can replace with the locking angle,

θ_L , for $\theta = 10, 20, 30$ and 40° . The range of possible solutions (θ_L, θ) for a given θ value falls within the shaded "wings" of the diagram. The boundaries of the wings are the result of limitations on θ which for any given friction angle ϕ are $(45-\phi/2)$ and $-(45-\phi/2)$. This is so because, if $|\theta|$ exceeds $(45-\phi/2)$, the layers in the outer limbs outside the kink band will slip and the kink band will not develop (Reches and Johnson, 1976). The darker shaded triangles of the butterfly diagram are for use with monoclinical kink folds, while both the light and dark shaded triangles are for use with conjugate kink folds.

The effect of the inclination, θ , of the maximum compression on locking angles is easily seen with the butterfly diagram. For any friction angle, ϕ , rotation of the maximum compression direction counterclockwise (positive θ) relative to the layering increases the locking angle for left-facing bands and decreases the locking angle for right-facing bands. Rotation of maximum compression clockwise (negative θ) decreases the locking angle for left-facing bands and increases the locking angle for right-facing bands. For example, suppose $\phi = 20^\circ$. Then if the inclination of the compression is $\theta = +10^\circ$, the left-facing band locks at $+80^\circ$ whereas the right-facing band locks at -60° . If the inclination of the compression is opposite, $\theta = -10^\circ$, the left-facing band locks at $+60^\circ$ and the right-facing band at -80° .

In the following analyses using the butterfly diagram, we assume that the orientation of the principal stresses does not change as the kink band forms. The effect of a changing stress state on kink fold formation is dealt with in later in the paper and briefly in Reches and Johnson (1976).

Using the Butterfly Diagram with Conjugate Kink Bands

The butterfly diagram in Fig. 11 can be used to determine both the friction angle, ϕ , and the inclination of maximum compression, θ , if the orientation of both limbs of a conjugate kink fold (θ_L and θ_R) have been measured. The quantities ϕ and θ are found by 1) plotting vertical lines through both θ_L and θ_R on the butterfly diagram, 2) plotting the horizontal line which intersects the two vertical lines at common values of ϕ and θ , 3) recording the values of ϕ and θ as the inclination of maximum compression and friction angle respectively. Examples from the San Rafael Swell in central Utah and the Mentasta Mountains in the eastern Alaska Range are given later to illustrate the use of the butterfly diagram for conjugate kink folds.

Using the Butterfly Diagram with Monoclinical Kink Folds

Unlike conjugate kink bands, where a single value for ϕ and θ is obtained, the butterfly diagram is used for monoclinical kink folds to

Fig. 11. Butterfly diagram relating (ultimate) locking angles, θ_L , of right- and left-facing kink bands to orientation, ϕ , of the maximum compression and the contact friction angle, β . Inset diagrams indicate facing of kink bands. For example, for a friction angle of $\beta = 10^\circ$, and maximum compression parallel to bedding, $\phi = 0$, the ultimate locking angle is $+80^\circ$ for a left-facing kink band and -80° for a right-facing kink band. Dark shaded portion is for use with monoclinal kink bands. Both light and dark shaded portions used for conjugate kink bands.

determine a range of values for ϕ . The ranges are obtained by 1) plotting a vertical line through θ_L on one side of the diagram, 2) estimating a range of ϕ that falls within the shaded "wing" (discussed below), 3) plotting the range of horizontal lines that intersect the line through θ_L within the estimated ϕ range, and 4) recording the values of ϕ that the horizontal lines intersect as the range of the inclination of maximum compression. An example from the Mentasta Mountains, Eastern Alaska Range, is used later to illustrate the use of the butterfly diagram for monoclinal kink folds (see Fig. 14).

Estimating the value of ϕ

The locking angles of kink bands are determined by the strength properties of interbeds or interfaces between layers (eqs. 3). In the case of monoclinal kink folds, an estimate of the contact friction angle must be determined to obtain ϕ . Fortunately it has been demonstrated that there is little variation in friction angles for a wide variety of rock types such as granites, sandstones, and limestones (Byerlee, 1978). For such materials we would expect friction angles between 20° and 40° , and therefore relatively low (positive) ultimate locking angles given by

ϕ

ϕ

$$\theta_{UL} - \phi = +(90^\circ - \phi) \quad \phi \text{ 50 to } 70^\circ.$$

On the other hand, for slip between interbedded stiff and soft materials, such as sandstone and mudstone or chert and mudstone, where slip occurs in the soft materials (mudstone), the contact friction angle may be lower. For example, the friction angle for clays and clay shales ranges from 0 to 20° (Attewell and Farmer, 1976; Lambe and Whitman, 1969). In such materials we expect relatively high (positive) ultimate locking angles given by

?

?

$$\theta_{UL} - \theta = 90^\circ - \theta \approx 70 \text{ to } 90^\circ.$$

The estimated range of θ may be further restricted with the butterfly diagram depending on the locking angle. For example, consider a kink fold that forms in a shale with a locking angle of 80° . While in general it is reasonable to estimate that θ may range from 0 to 20° , in Fig. 3B it can be seen that only θ values between 0 and 10° lie within the shaded "wing" at a locking angle of 80° .

Large Kink Folds in the McCarthy Formation, Eastern Alaska Range

There are spectacular examples of large kink folds in the McCarthy Formation in the eastern Alaska Range near the Yukon-Alaskan border. The kink bands form monoclinial kink folds and asymmetric conjugate kink folds in the footwall of the Lost Creek detachment⁴⁴, which places a Mesozoic sedimentary sequence on top of the Paleozoic-Mesozoic Wrangellia composite terrane. The kink folds are near the detachment surface so they may well have formed at the time of thrusting. If so, the sense of asymmetry should reflect the transport direction of the detachment block.

The Lost Creek detachment fault dips to the north at approximately 30° and its trace is from northwest to south-southeast (Fig. 12). Cross-cutting relationships of dikes and plutons constrain the age of the Lost Creek detachment to be Cretaceous (Richter et al, 1977; Richter, 1976; and Richter and Schmoll, 1973) during which regional compressional deformation events were recorded throughout Wrangellia and the western margin of North America (Refs).

The hanging-wall of the Lost Creek detachment consist of the Jurassic-Cretaceous Nutzotin Mountains sequence (Fig. 12). The rocks consist primarily of fine grained, siliciclastic turbidity deposits with interbedded sandstones and channelized conglomerates. Drag folds formed in the hanging wall above a highly deformed 10 m thick fault zone.

The footwall of the detachment consist of the Late Triassic McCarthy formation, one of the youngest formations within the Wrangellia composite terrane. The McCarthy consists of about 100 m of dark-gray, fine-grained limestone, with bedding ranging in thickness from a few centimeters to a few meters, interbedded with black chert, siliceous argillite, and carbonaceous shale (Fig. 13B). It is within the McCarthy formation that the large-scale kink bands have formed.

The asymmetric folds in the footwall of the Lost Creek detachment are identified as kink bands because 1) they contain an isolated tabular shear zone that cuts across bedding, 2) they consist of two outer limbs which are subparallel to each other and an inner limb which is inclined with respect to the outer limbs, and 3) they have formed within multilayered material with interbedded stiff (limestone) and soft (shale) layers (Fig. 13).

⁴⁴ Newly described. Named herein.

The inner limb of the kink bands are on the order of 4 m wide. The kink bands are isolated and continuous through more than 40 m of section (Fig ???). There are at least eight kink bands facing northeast, toward the source area of the overthrust block, and there is one kink band in the same area facing southwest.

The locking angle was measured of one southwest-facing and one northeast-facing kink band. The southwest facing kink band has a locking angle, $\theta_{UL} = 83^\circ$ and the kink band facing northeast has a locking angle, $\theta_{UL} = 83^\circ$.

There are two ways one can work with the limb orientations of the kink bands within the McCarthy Formation:

Assume that the set of eight kink bands facing southwest are a monoclinial fold. Determine the compression direction that is consistent with the data.

Assume that the southwest facing and the northeast facing kink bands form an asymmetric conjugate kink fold. Determine both the contact friction angle and the compression direction.

In both cases, we use the butterfly diagram.

Fig. 14. Simplified monoclinial butterfly diagram used for analysis of McCarthy kink folds. $\theta_L = 75^\circ$, θ ranges from 0 to 15° , and β ranges from 0 to -15° .

Starting with the first case, most of the kink bands face northeast, and only in one place is there an opposite-facing kink band, thus we assume that the northeast-facing kink bands form a monoclinial fold. We identify the view in the butterfly diagram as toward the southeast (Fig. 14) and define left- and right-facing kink limbs accordingly. The northeast-dipping limb has a positive angle, $\theta_L = 75^\circ$. We can determine the sense of layer-parallel shear and a reasonable range of values for θ and β with the butterfly diagram. First, a vertical line is plotted through $\theta_L = 75^\circ$ on the right side of the simplified butterfly diagram in Fig. 14. Next, the range of contact friction angles, β , for the interbedded shale and limestone of the McCarthy is estimated. For $\theta_L = 75^\circ$, it can be seen with the butterfly diagram that β is restricted to the range between 0 and 15° (shaded in Fig. 7). This range is consistent with the range of β between 0 and 20° for friction within shales and clays. Finally, a range of horizontal lines is plotted on the diagram such that they intersect the vertical θ_L line within the range 0 to -15° (shaded dashed rectangle in Fig. 7). This produces a range of solutions: -15° to 0° .

The second approach is to assume that the northeast-facing and southwest-facing kink bands formed under the same stress state as an asymmetric conjugate kink fold. We can use the butterfly diagram to determine simultaneously the friction angle and the orientation of maximum compression. We again identify the view in the butterfly diagram as toward the southeast (Fig. 15) and define left- and right-facing kink limbs accordingly. The southwest-dipping limb has a negative angle, θ_L $\approx$ -83° . The northeast-dipping limb has a positive angle, θ_L $\approx$ 75° . The two vertical dashed lines in Fig. 15 represent these angles. The solution is obtained by drawing a horizontal line from one vertical line to the other, making sure that each end of the line is on the same friction angle, β . Thus, we obtain an estimate of the friction angle, β $\approx$ 10° , and an inclination of the maximum compression, θ $\approx$ -5° , to the bedding.

The results of the two methods are in agreement. Furthermore, the values obtained with both approaches are reasonable. The slight inclination of the maximum compression, so that θ ranges from 0 to -15° if the kink folds are treated as monoclinical, or θ $\approx$ -5° if the folds are conjugate, indicates that the layer-parallel shear was right-lateral in the view of Fig. 14. The range of low friction angles between 0 and 15° is consistent with the properties of clays and shales.

Fig. 15. Simplified butterfly diagram used to determine inclination of maximum compression and contact friction angle.

The most important interpretation that can be made from the kink folds which have formed in the McCarthy Formation is the displacement direction along the detachment between the Mesozoic Nutzotin Mountains sequence and the accreted Island arc. Both approaches discussed above result in an interpretation of a thrusting relationship with the upper block moving relatively southwest.

The interpretation of sense of layer-parallel shear using the butterfly diagram is consistent with other lines of evidence suggesting southwest displacement of the hanging wall, which include stepping directions of slickensided fault surfaces within the décollement and asymmetry of drag folds immediately above the décollement.

Fig. 16. Geology of Spotted Wolf section of the San Rafael Swell.

Small Kink Folds in the Navajo Sandstone, San Rafael Swell, Utah

There are many examples of small-scale conjugate kink bands in the Navajo Sandstone at the eastern edge of the San Rafael Swell, Utah. The kink bands and numerous small faults formed within the monoclinical fold at the edge of the Swell. The folds in the Navajo are on the order of 10 cm wide and many meters long, so they are about two orders of magnitude smaller than those in the McCarthy Formation. Also, the low angles, on the order of 30° , between limbs of the kink bands in the Navajo Sandstone are in stark contrast with the high angles, on the order of 70 to 85° , between limbs of kink bands in the McCarthy Formation.

Analysis of the kink bands in the Navajo Sandstone can provide us with important information about the deformation of the strata during formation of the San Rafael Swell, for which the mechanism of uplift is unknown. With our understanding of kink-band formation, we can use the orientation of the kink bands to determine the local stress state at the time of uplift.

The San Rafael Swell is an asymmetric, domical uplift on the Colorado Plateau that stretches ~ 120 km north to south and ~ 50 km east to west (Fig. 16). The structural relief on the east side of the Swell is on the order of 1 km. Uplift initiated in latest Campanian time (late Cretaceous; Lawton, 1983) and is believed to be related tectonically to the Laramide basement uplifts of the Rocky Mountain foreland (e. g. Cross, 1986). The Swell is structurally similar to other major uplifts on the Colorado Plateau (Kelley and Clinton, 1960). The beds of late Paleozoic through Mesozoic rocks comprising the Swell dip gently westward over the center and west edge of the Swell and a monoclinical fold forms the eastern flank of the Swell where the beds dip steeply eastward and are locally overturned. It is in the steeply-dipping inner limb of the monocline that the kink bands in the Navajo Sandstone formed.

The location of the kink bands described here is marked with a star in Fig. 16. Small faults formed parallel to the cross-bedding at this location (Mark Fischer, personal

communication). The cross-bed surfaces apparently have relatively low strength, and as discussed elsewhere (Johnson and Johnson, in prep), slippage occurred along the cross-beds in response to the bending of the strata during formation of the San Rafael Swell. In places this slip along cross-beds was highly localized into narrow shear zones in the form of kink bands (Fig. 17).

The faulting in the Navajo Sandstone, related to kink-band formation, is in the form of deformation bands⁴⁵, which are very thin, tabular bodies of highly localized cataclasis in porous sandstones. Deformation bands have been studied in detail by several investigators (Aydin, 1977, 1978; Aydin and Johnson, 1978, 1983; Jamison and Stearns, 1982; Underhill and Woodcock, 1987; Cruikshank et al., 1991; Zhao and Johnson, 1991; Antonelli et al., 1994).

The kink bands are narrow tabular zones of folded cross-beds about a decimeter wide and several meters long. The isolated fold shown in Fig. 17 is characteristic of a kink band with a short, reoriented inner limb and two straight outer limbs. Furthermore, the slip is clearly localized within the inner limb, as one can see the high concentration of deformation bands parallel to cross-bedding within the inner limb compared to the relatively sparse distribution in the outer limbs.

Fig. 17

The cross-bed parallel faults within the inner limb are offset by deformation bands⁴⁶ oriented both parallel and at a low angle to the walls of the kink band. The offsetting relationship clearly indicates that the faults parallel to cross-beds formed before the other faults. We may be able to explain the sequence of faulting in this shear zone with our understanding of kink-band formation. As the cross-beds folded locally into a kink band, slippage between layers, in the form of deformation bands, was localized within the inner limb. Shear stresses within the zone were accommodated by folding and slippage between layers until the kink band locked. Presumably the faults sub-parallel to the walls of the kink band subsequently formed after the kink band locked, during continued shearing.

Figure 18 shows asymmetric conjugate kink bands in the Navajo Sandstone. The view is roughly parallel to the strike of the cross-bed surfaces and the trend of the San Rafael monocline. The left-facing kink band is oriented at a higher angle to the layering than the right-facing kink band. By analyzing the geometry of these kink bands

⁴⁵ Also known as Atilla faults, micro faults, seal faults, granulation seams. See description by Aydin (1978).

⁴⁶ The faults at a low angle to the walls of the kink band are in an orientation consistent with faults that have been termed Riedel shears. Riedel shears are "minor" faults that form at a relatively low angle to a major fault or shear zone (e.g. Suppe, 1985). Jamison and Stearns (1982) documented these same structures in the Wingate Sandstone in Colorado National Monument. They in fact called these tabular shear zones Riedel shear zones rather than kink bands.

A.

B.

Fig. 18. Conjugate kink fold in Navajo Sandstone. Left-facing steep limb and right-facing shallow limb. A. Interpreted. The locking angles are largest limb angle of the right-facing kink band and the corresponding limb angle of the left-facing kink band. B. Uninterpreted.

simultaneously, using the butterfly diagram, we can determine roughly the contact friction angle in the deformation bands and the orientation of the maximum compression.

The inclination, β , of the maximum compression is determined by the orientation of the locking angles of the two kink bands, assuming both kink bands formed under the same stress state. While the orientation of the inner limb varies with position along the kink band, the largest angle of inclination of the inner limb is the locking angle, θ_L . The largest angle in the left-facing limb in Fig. 18 is $\theta_L = +66^\circ$ (range 34 to 66°) and the corresponding angle in the right-facing limb is $\theta_L = -34^\circ$ (range -20 to -34°).

We can use the butterfly diagram in Fig. 19 to determine β and θ_L . The point (θ_L, β) which is the solution for both the right and left facing limbs must lie on the heavy, vertical dashed lines through 66° and -34° . The heavy, horizontal dashed line shows the solution: $\beta = 17^\circ$ and $\theta_L = 40^\circ$.

The high contact friction angle is consistent with the range of 20 to 40° measured for many clastic and crystalline rocks. Furthermore, the inclination of maximum compression to layering of 17° is consistent with an independent measurement based on stress inversion of small faults (Johnson, in prep.). The cross-bedding surfaces shown in Fig. 4 are dipping 15 to 20° to the east (right) and the beds (not visible in the photo) are dipping 50° to the east, so if the kink bands formed while the beds were in their present position, then the maximum compression was roughly horizontal. The stress inversion of small faults in Johnson (in prep.) also calculated horizontal maximum compression at this location.

Fig. 19. Simplified butterfly diagram used with Navajo kink fold. $\beta = 40^\circ$ and $\theta_L = 17^\circ$.

CHANGING STRESS STATE

Notice in Fig. 11 that θ_L is limited to the range from 0 to 90° for monoclinical kink folds (dark shaded region). This limitation is a result of the assumption that the orientation of maximum compression relative to the layering does not change throughout formation of the kink fold. It is however possible for the inner limb of monoclinical kink folds to exceed 90° if the orientation of the compression relative to the layering does change during formation. The locking angle of a monoclinical kink fold will increase if the compression direction rotates in the positive (counterclockwise) direction for left-facing kink folds or the negative (clockwise) direction for right-facing kink folds. For example, consider

monoclinal kink folds in shale that formed under the conditions of effective friction $\phi = 0^\circ$, and $\theta = -5^\circ$. With the butterfly diagram we can see that the locking angle is $\theta_L = 85^\circ$. Now suppose that the layering rotates or the stress state changes so that $\theta = 0^\circ$. If we follow along the $\theta = 0^\circ$ line on the butterfly diagram from the point $(\theta_L, \theta) = (85, -5)$ to $\theta = 0^\circ$, we see that the new locking angle is $\theta_L = 100^\circ$. In this way an overturned monoclinal kink fold can form. Notice that the left-lateral layer-parallel shear in the final stress state ($\theta > 0$) is opposite in sense to the right-lateral layer-parallel shear that initiated the kink fold ($\theta < 0$). For overturned monoclinal kink folds to form, therefore, it is necessary that the sense of layer-parallel shear reverse.

The butterfly diagram in general should only be used in structurally "simple" settings, however. If it is clear that the deformation reflects multiple, superimposed stress states, then the butterfly diagram will not be useful. While we can determine that overturned monoclinal kink folds form under a changing stress state, it is not possible in general to deduce a changing stress state with the butterfly diagram. For example consider an asymmetric conjugate kink fold that completely forms and locks with $\theta_{Lr} = -70^\circ$, $\theta_{Li} = 50^\circ$, $\theta = 30^\circ$, and $\phi = -10^\circ$. Now suppose the stress state changes so that $\phi = 5^\circ$. With the butterfly diagram, we can see that the left-facing kink band will reorient and lock at $\theta_{Li} = 65^\circ$. The right-facing kink band however will not reorient because the locking angle for the initial stress state is greater than the locking angle for the final stress state, and clearly, under sub-layer-parallel compression, the kink band will not relax to a lower locking angle. The butterfly diagram cannot deduce the final stress state with the geometry of the altered kink fold. In this example one would determine with the butterfly diagram that the final inclination of maximum compression is $\theta = -2.5^\circ$. Thus we would interpret the incorrect sense of the final layer parallel shear stress.

CONCLUSIONS

A summary of the theory of kink band formation shows that the important factors controlling kink-band formation are the strength between layers, maximum compression inclined between $\theta(45\text{--}90/2)$ degrees to the layering in a multilayered material, and an initial perturbation. Although the existing theory of kink-band formation has been developed over the years in many separate publications (e. g. Peng and Johnson, 1972; Honea and Johnson, 1976; Reches and Johnson, 1976; Johnson, 1977; 1995; Johnson and Pfaff, 1989; Johnson and Fletcher, 1994), we have provided a concise compilation of the theory that culminates with the butterfly diagram (Fig. 11), a new way to present the relationships among θ , θ_L , and ϕ .

By measuring only the locking angles of monoclinal kink folds in the field, one can determine, with the butterfly diagram, the orientation of maximum compression and therefore sense of layer-parallel shear responsible for formation of the folds. We illustrate this feature by analyzing monoclinal kink folds in the McCarthy Formation, in the footwall of the Lost Creek thrust sheet, and using the result to determine the transport direction on a décollement. The butterfly diagram predicts that the maximum compression was inclined 0 to 15° to the southwest, implying southwest transport direction on the décollement. This interpretation is supported by independent measurements of slickensided surfaces and orientation of drag folds. The butterfly diagram also predicts that the angle of contact friction falls in the range from 0 to 15° , which is consistent with the range determined experimentally for shales.

By measuring locking angles of conjugate kink bands, the butterfly diagram can be used to determine both the orientation of maximum compression and the angle of contact friction. We use the butterfly diagram to analyze conjugate kink folds in cross-bedded Navajo Sandstone at San Rafael Swell, Utah. Analysis of an asymmetric conjugate kink fold indicates horizontal compression, which is consistent with an independent result

from analysis of faults (Johnson, in prep.). The contact friction angle is determined to be 40°, which is consistent with friction angles measured for many clastic and crystalline rocks.

ACKNOWLEDGEMENTS

Partial support of this research was provided by the Donald Levandowski memorial scholarship, Cedric J. Newby scholarship, Linda Horn memorial scholarship, and the John T. Dillon Alaskan Research Grant. The remainder of the support for this research was provide by the Department of Energy grant DE-FG02-98ER14886 and the donors of the Petroleum Research Fund, administered by the American Chemical Society. Special thanks is given to field assistants Shane Smith, Mike Depersia, Brian Lareau, and Cyndi Johnson, and Joe Durdella for his helpful comments.

HOW TO ANALYZE KINK FOLDS (SHORT EXPLANATION)

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ABSTRACT

The ability to deduce the sense of layer-parallel shear during regional shortening and emplacement of detachment sheets and blocks is central to our ability to determine relative movements of crustal masses. The sense of asymmetry of drag (parasitic) folds is a well known indicator of relative transport directions but the asymmetry of asymmetric kink folds and monoclinial kink folds are more difficult to interpret. The problem is that monoclinial kink folds that display the *same* sense of asymmetry as drag (parasitic) folds are produced by the *opposite* sense of layer-parallel shear. Here we examine the sense of asymmetry of kink folds and explain how to measure the geometry and correctly deduce the sense of shear. The orientation of maximum compression and therefore the sense of layer-parallel shear responsible for formation of both drag folds and kink folds can be determined by making appropriate field measurements.

Key words: Kink band, kink fold, drag fold, asymmetric kink fold, conjugate kink fold, geometry of kink fold.

INTRODUCTION

This paper addresses questions that arose during field studies of folds in the eastern Alaska Range and in the San Rafael Swell, Utah. The questions are, how does one recognize a kink fold and how does one obtain mechanical information from measurements of fold geometry? In the process of answering these questions, and interpreting the field examples, we have developed a useful method for analyzing kink bands and correctly determining the sense of shear indicated by the sense of asymmetry of kink folds.

ASYMMETRY OF FOLDS

A *kink band* is the fundamental building block of kink folds. It is a tabular shear zone that cuts across layered media such as bedding, foliation, crystal lattices or cleavage planes. It is characterized by a monocline with an inner limb in which the layering has been reoriented with respect to that in the two outer limbs (Fig. 1; Dewey, 1965, p. 460; Suppe, 1985, p. 336).

There are three variants of kink folds: symmetric conjugate, asymmetric conjugate, and monoclinal kink folds (Fig. 1). The symmetric conjugate kink fold (Figure 1A) indicates negligible layer-parallel shear stress. The asymmetric conjugate kink fold shown in Fig. 1B indicates significant left-lateral, layer-parallel shear stress. The monoclinal kink fold shown in Fig. 1C indicates greater left-lateral layer-parallel shear stress (Johnson, 1977). If the layer-parallel shear stress is reversed, the asymmetries of the fold forms are reversed.

Figure 1. Kink and drag folds. All four types form under layer-parallel compression. A. Symmetric conjugate kink fold. Contains identical kink bands facing left and right. Produced by layer-parallel compression (zero layer-parallel shear). B. Asymmetric, conjugate kink fold. Right-lateral layer-parallel shear stress causes the asymmetry shown in the figure. C. Asymmetric, monoclinal kink fold. Larger right-lateral shear causes only left-facing limbs to form. D. Drag (or parasitic) fold. Shorter limb faces to the right.

Drag folds might easily be confused with asymmetric conjugate kink folds or monoclinal kink folds because of their asymmetry. Drag folds consist of a series of asymmetric plications with alternating short and long limbs and with axial planes that dip in the same direction as the long limb (Fig. 1D) (DeSitter, 1964, p. 279). Drag folds can be considered to be folds that formed by mechanical buckling of layers and then by kinematic simple shearing (Ramberg, 1963). Layer-parallel simple shearing accounts for most of the asymmetry of drag folds (Ramberg and Johnson, 1976; Pfaff, 1986) and for this reason, the sense of layer-parallel shearing responsible for drag (or parasitic) folds is easy to interpret.

We see in Fig. 1B that the asymmetric conjugate kink fold has the same sense of asymmetry as the drag fold in Fig. 1D. In both cases, right-lateral shearing is applied and the short limb is facing right. The facing direction of the steeper limb is reversed if the sense of shear is reversed. The monoclinal kink fold, however, has the opposite sense of asymmetry (Fig. 1C) from drag folds and asymmetric, conjugate kink folds. Thus the problem is that one cannot apply the intuitive notion to

some kinds of kink folds that the facing direction of the steeper limb of an asymmetric fold indicates the direction of layer-parallel shear.

The central problem of interpreting the sense of shear of asymmetric folds is therefore the determination of the kind of fold one is studying. One must be able to recognize three kinds of asymmetric folds—*drag (or parasitic) folds* (Fig. 1D), *asymmetric, conjugate kink folds* (Fig. 1B) and *monoclinal kink folds* (Fig. 1C).

EFFECT OF INCLINATION OF MAXIMUM COMPRESSION ON KINK FOLD TYPE

Figure 2 shows the three types of kink folds and the maximum compression directions responsible for the formation of each type. As we show in more detail elsewhere (Johnson et al., in prep.), the inclination of the direction of maximum compression, β , determines whether the kink folds form as a set of monoclinal or conjugate kink bands. When the orientation of maximum compression is parallel to layering ($\beta = 0$), *symmetric conjugate kink bands* form (Fig. 2A). The locking angles of the left-facing limb, θ_{Lr} , and the right-facing limb, θ_{Lr} , are equal in magnitude, $|\theta_{Lr}| = |\theta_{Lr}|$.

When the maximum compression is slightly inclined ($\beta \neq 0$) to the layering in a clockwise or counterclockwise sense, kink bands may form as *asymmetric conjugate kink bands* $\theta_{Lr} \neq \theta_{Lr}$ (Fig. 2B). Finally, when the maximum compression is more highly inclined to the layering, kink bands may form as *monoclinal kink folds* with all left-facing or right-facing limbs (Fig. 2C).

A. Symmetric Conjugate Kink Bands

B. Asymmetric Conjugate Kink Bands

C. Monoclinal Kink Band

Figure 2. The three different types of kink bands and the orientation of the maximum compression direction, β , responsible for each type. The locking angle, θ_L , is negative for right-facing and positive for left-facing kink bands.

The locking angles of kink bands can be expressed in terms of only two variables, the friction angle, ϕ , at layer surfaces and the inclination of maximum compression, θ . For positively sloping layers in the inner limb (left-facing limb in Fig. 2B)

$$\theta_L = 90^\circ - \phi - \theta; \quad \theta \geq 0, \quad (1a)$$

whereas for negatively sloping layers (right-facing limb in Fig. 2B),

$$\theta_L = 90^\circ - \phi + \theta; \quad \theta \geq 0. \quad (1b)$$

STRENGTH OF INTERFACES BETWEEN LAYERS

When working with kink bands in the field, the single observation we can make is the locking angle. If we are able to measure only one locking angle, eq. (1) gives us a relation between locking angle, θ_L , friction angle, ϕ , and inclination of the maximum compression, θ . In this case, if we can reasonably estimate the friction angle, we can estimate the inclination of compression.

The locking angles of kink bands are determined by the frictional strength properties of interfaces between layers or of interbeds. So, in order to interpret a locking angle of a kink band we need to be able to estimate the effective angle of internal friction, ϕ .

Some rules of thumb from soil mechanics will help us select appropriate friction angles. For flexural slip between relatively permeable beds, such as sandstones or interbedded sandstones and siltstones, we would assume purely frictional behaviors (Corresponding to residual strength of soil mechanics). For such materials we would expect friction angles between 20 and 40°, so we should expect relatively low (positive) locking angles given by

$$\theta_L = +(90^\circ - \phi) + \theta \quad (50 \text{ to } 70^\circ) + \theta$$

On the other hand, flexural slip between interbedded stiff and soft materials, such as sandstone and mudstone or chert and shale where the soft materials—mudstone and shale—have low permeability, should be described in terms of a material whose strength is purely cohesive; zero friction angle, $\phi = 0$ (Corresponding to undrained strength of soil mechanics). We expect magnitudes of locking angles in such materials to be 90° or even larger,

$$\theta_L = 90^\circ - \phi + \theta \quad 90^\circ + \theta$$

depending on the inclination, β , of the maximum compression to layering.

THE BUTTERFLY DIAGRAM OF LOCKING ANGLES

We introduce the butterfly diagram (Fig. 3), which is a plot of eqs. (1), but with appropriate limits, in order to facilitate the determination of friction angle and inclination of compression in the case of measurements of a pair of conjugate kink bands or of multiple measurements of kink bands. As we will show below, if we have measurements of the orientation of the limbs of a conjugate fold, we can use the butterfly diagram to determine both the friction angle and the inclination of the compression.

Figure 3. Butterfly diagram relating locking angles, θ_L , of right- and left-facing kink bands to orientation, β , of the maximum compression and the angle of internal friction, ϕ . Inset diagrams indicate facing of kink bands. Heavy, dashed lines are for use with the Navajo Sandstone example. The heavy, solid line is for use with the McCarthy example.

The butterfly diagram (Fig. 3) helps one to determine these quantities. It is a plot of eqs. (1), but with necessary limiting conditions. (Note: the heavy lines are for use with the field examples as discussed below.) It plots β as a function of locking angle θ_L for $\phi = 10, 20, 30$ and 40° . Two of the boundary lines are a result of limiting angles of maximum compression, β . Given a value for friction angle, ϕ , the angle of inclination of maximum compression, β , can range between $45^\circ - \phi/2$ and $45^\circ + \phi/2$ (the range falls within the shaded area of the butterfly diagram).

This is so because, if $|\beta|$ exceeds $45^\circ - \phi/2$, the layers in the outer limbs outside the pupative kink band will slip and a kink band will not develop. As a result of the range of possible angles, β , there is a range of possible locking angles for a given friction angle, ϕ .

LARGE KINK FOLDS IN THE MCCARTHY FORMATION, EASTERN ALASKA RANGE

We will illustrate how to use the butterfly diagram with field examples.

There are excellent examples of large kink bands in the McCarthy Formation in the eastern Alaska Range near the Yukon-Alaskan border. The kink bands form monoclinial kink folds (Fig. 4) in footwall rocks of the Lost Creek detachment, which places a Mesozoic sedimentary sequence on top of one of the largest Paleozoic-Mesozoic accreted terranes in North America, the Wrangellia composite terrane. The kink folds are situated closely beneath the detachment surface so they may well have formed at the time of thrusting. If so, the sense of asymmetry should reflect the transport direction of the detachment block.

The Lost Creek detachment fault dips to the north at approximately 30° and its trace is from northwest to south-southeast. The footwall rocks of the detachment, which contain the large kink bands, consist of the Late Triassic McCarthy Formation, about 100 m of dark-gray, fine-grained limestone, with bedding ranging in thickness from a few centimeters to a few meters, interbedded with black chert, siliceous argillite, and carbonaceous shale (Fig. 4 shows clearly the nature of the bedding).

The asymmetric folds in the footwall of the Lost Creek detachment display the characteristics of kink bands:

- ? They contain an isolated tabular shear zone that cuts across bedding. The inner limb is up to 4 m wide and 40 m long.
- ? They consist of two outer limbs, which are subparallel to each other, and an inner limb, which is inclined with respect to the outer limbs.
- ? Eight of the kink bands face NNE. One faces SW.
- ? A locking angle of $??_L$ was measured in the field for one of these kink bands?
- ? They have formed within multilayered material with interbedded stiff (limestone) and soft (shale) layers (Fig. 4).

We identify the view in the butterfly diagram as toward the southeast (Fig. 3) and define facing direction of the kink limb accordingly. Thus the left side of the butterfly diagram is to the NE and the right side is to the SW for this example.

The field measurements indicate that the northeast-dipping limb has a positive angle, $??_L$. The vertical solid line in Figure 5 represents this angle. Because thin shale layers are interbedded with thicker limestone beds, we assume low frictional strength (μ between 0 and 10%) between layers. This results in an inclination of the maximum compression, μ between -5° and 15° , to the layering outside the kink bands.

The direction of inclination of the compression (and, therefore, the sense of layer-parallel shear) is reasonable. The slight inclination of the maximum compression so that μ -5° to -15° indicates that the layer-parallel shear was right-lateral in the view of Figure 4. This is consistent with transport of rocks above the detachment from northeast to southwest in the area.

The conclusion that the hanging-wall block of the Lost Creek detachment fault moved generally southwestward relative to footwall rocks is supported by two other indicators: stepping directions of slickensided faults within the detachment fault zone and asymmetry of drag (parasitic folds) in rocks immediately above the detachment.

SMALL KINK FOLDS IN THE NAVAJO SANDSTONE, SAN RAFAEL SWELL, UTAH

There are many examples of small monoclinical and conjugate kink folds in the Navajo Sandstone at the eastern edge of the San Rafael Swell, Utah. The kink bands and numerous small faults formed along the monoclinical fold defining the eastern flank of the Swell.

The kink folds in the Navajo form a nice contrast with the kink folds in the McCarthy. Those in the Navajo are on the order of 10 cm wide and many meters long, so they are about two orders of magnitude smaller than those in the McCarthy. Also, the low angles, on the order of 30° , between limbs of the kink bands in the Navajo Sandstone are in stark contrast with the high angles, on the order of 75° , between limbs of kink bands in the McCarthy Formation. The low angle in the Navajo indicates a high frictional strength between slipping layers of sandstone whereas the high angle in the McCarthy indicates a low frictional strength in shale interbeds.

The kink bands in Navajo Sandstone formed in cross-beds within steeply dipping sandstone beds on the eastern, monoclinical flank of the San Rafael Swell. As discussed elsewhere (Johnson and Johnson, in prep), slippage occurred along the cross-beds in response to the bending of the strata during formation of the structure. In places this slip along cross-beds was highly localized into narrow shear zones in the form of kink bands (Fig. 6).

The folds in the Navajo Sandstone have all the characteristics of kink bands and kink folds. The kink bands are narrow tabular zones of folded cross-beds about a 10 cm wide and several meters long. Figure 6 shows asymmetric conjugate kink bands in the Navajo Sandstone. The view is roughly parallel to the strike of the cross-bed surfaces and the trend of the San Rafael monocline. The left-facing kink band is oriented at a higher angle to the layering than the right-facing kink band. By analyzing the geometry of these kink bands simultaneously using the butterfly diagram we can determine roughly the internal friction angle in the deformation bands and the orientation of the maximum compression, assuming both kink bands formed under the same stress state.

While the orientation of the inner limb varies with position along the kink band, the largest angle of inclination of the inner limb is the locking angle, θ_L . The largest angle in the left-facing limb in Fig. 17 is $\theta_{Ls} = +66^\circ$ (range 34 to 66°) and the corresponding angle in the right-facing limb is $\theta_{Lr} = -34^\circ$ (range -20 to -34°). We can use the butterfly diagram in Fig. 3 to determine β and ϕ . The point (ϕ, β) which is the solution for both the right and left facing limbs must lie on the heavy, vertical dashed lines through 66° and -34° . The heavy, horizontal dashed line shows the solution: $\beta = 17^\circ$ and $\phi = 40^\circ$.

The orientation of the maximum compression, β is particularly interesting. The cross-bedding surfaces shown in Fig. 6 are dipping 15 to 20° to the east (right) and the beds (not visible in the photo) are dipping 50° to the east. In Johnson and Johnson (in prep.), we show that the faults parallel to cross-bedding formed late in the uplift of the San Rafael Swell, when the beds were at (or near) their present orientation. Therefore the kink bands must have formed with the beds in their present position. With $\beta = 17^\circ$, the maximum compression is thus roughly horizontal. This result is consistent with the stress state obtained by analysis of faulting at the same location by Johnson (in prep.).

Figure 6. Asymmetric conjugate kink bands in cross-beds of Navajo Sandstone. Locking angle is 66° for the left facing limb -34° for the right facing limb.

CONCLUSIONS

By measuring only the locking angles of kink bands in the field one can determine, with aid of the butterfly diagram, the orientation of maximum compression and therefore the sense of layer-parallel shear responsible for formation of kink folds. In the case of conjugate kink folds, both the internal angle of friction, ϕ , and angle of inclination, λ , of the maximum compression direction can be determined directly using the butterfly diagram. In the case of monoclinical kink folds, however, the internal friction angle must be estimated.

Additionally, we illustrated the value of analyzing kink bands with examples from southern Alaska and Utah. The Alaskan kink bands reflect the transport direction on a detachment fault between an extensive Mesozoic sedimentary basin and one of the largest accreted terranes in North America. Independent measurements of slickensided surfaces and orientation of drag folds affirm the transport direction. The kink bands at San Rafael Swell reflect the orientation of maximum compression during Laramide deformation of the Navajo Sandstone. The direction of maximum compression is affirmed by independent measurements of orientations and slip directions of multiple faults.

ACKNOWLEDGEMENTS

This research was made possible by financial support for field work by the Donald Levandowski memorial award the Cedric J. Newby award, Department of Earth and Atmospheric Sciences,

Purdue. The remainder of the field expenses were paid by Department of Energy grant DE-FG02-98ER14886.

SOME COMMENTS ON HIGH-AMPLITUDE KINK FOLDING

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1999

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INTRODUCTION

In a companion paper (Johnson and Johnson, 2000), which discusses the division of single thick layers into multiple thin layers by the formation of layer-parallel faults (delamination⁴⁷) in several monoclinial flexures, is controlled by two parameters: the contact strength, τ_{cont} , at layer contacts and the change in curvature, $\frac{d\kappa}{ds}$, that is, the *coil*, of the layers. This result is based on our assumptions that the shear stress is given approximately by high-amplitude beam/plate theory and that delamination, which occurs at midheight of a layer, is driven by the layer-parallel shear stress, τ_{mid} , at midheight of the layer:

$$\tau_{mid} \geq \tau_{cont} \geq \frac{BT^2}{8} \frac{d\kappa}{ds}, \quad (1)$$

⁴⁷ Delamination is the local or general subdivision of a single mechanical layer into two or more mechanical layers. A single mechanical layer may contain many or no sedimentary layers. For example, a slab of granite bounded by joints is a mechanical layer. A varved clay could be a single mechanical layer if the strength of contacts is high or many mechanical layers if the strength of contacts between varves is low.

where s is the distance along the layer, κ is curvature, $\frac{d\kappa}{ds}$ is coil, T is the thickness, b is the width (unity for a layer), B is an elasticity modulus, and $I = \frac{bT^3}{12}$ is moment of inertia of the layer. We expect that if a layer delaminates it will delaminate at the location where τ_{mid} is maximal in the layer. In the companion paper we cited several field examples where the delamination is as expected according to eq (1).

The result in eq. (1) is unusually simple and depends entirely on variables that can be measured or estimated. The elasticity modulus B can be estimated in terms of the rock type (e.g., Johnson, 1970, p. 62 and 202). The sign of the contact strength, τ_{cont} , can be determined from the sense of slip between layers (right-lateral is positive). The coil and the layer thickness can be measured in the field. The sign of the coil is illustrated in Fig. 1 in terms of an Archimedes spiral. The coil, $\frac{d\kappa}{ds}$, of a layer is negative if the fold tightens as it is traversed in a clockwise direction as does the negative Archimedes spiral in Fig. 1B. The coil, $\frac{d\kappa}{ds}$, of a layer is positive if the fold tightens as it is traversed in a counter-clockwise direction (Fig. 1A).

Figure 1. Coil, the change of curvature along a curve, illustrated with Archimede's spiral. A. Negative spiral, showing negative coil. B. Positive spiral, showing positive coil.

FIELD EXAMPLES OF DELAMINATION

Watts Fold

Pfaff (1986a) studied monoclinial folds in interbedded mudstone and siltstone in central Pennsylvania. The folds are exposed in roadcuts along highway 22/322 along the Juniata River, about 35 km northwest of Harrisburg, Pennsylvania. The rocks are part of the Trimmers Rock and Brallier formations, of Devonian age. The Watts fold, shown in Fig. 2, is a large monoclinial fold, with a limb length of about 10 m. At the upper end of the fold the inner limb is in mudstones containing a few siltstone beds. Below that, the inner limb cuts through a package of four rock units, each about 20 m thick and alternating from mudstone to siltstone.

The Watts fold is shown in Fig. 2 via a sketch of traces of fractures. Bedding surfaces were not mapped, thus the spacing of the lines indicates the spacing of fractures. Although the

entire monoclinial fold could not be mapped in detail, the map of fractures shown in Fig. 2 allows one to visualize the distribution of fractures through the inner and outer limbs.

The inner limb is clearly delaminated. The bedding-parallel fractures are widely spaced in the outer limbs and closely spaced in the inner limb. The concentration of bedding-parallel fractures in the inner limb is about ten fractures per meter, whereas the concentration in the outer limbs is about one fracture per meter.

Figure 2 (modified from Pfaff, 1986a). Sketch of bedding in Watts monoclinial kink fold, along highway 22/322 in central Pennsylvania. Detailed map shows only fault traces in a large exposure of the limb of the kink band along a roadcut bench.

Amity Hall Fold

The Amity Hall fold is exposed in rock units nearby those in Fig. 2 in a roadcut near Amity Hall, along highway 22/322. The sketch of traces of fractures in Fig. 3A shows that the inner limb of the fold, which is about 2 m long, is more thoroughly delaminated than the two outer limbs (Fig. 3A). The spacing of fractures is reduced at least by one half in the inner limb.

The extra fractures in the inner limb extend only as far as the fold hinges, as shown in a detailed sketch (Fig. 3B) of the synclinal hinge of a layer marked with a rectangle in Figure 3A. The sketch shows several bedding-parallel faults that accommodated 1 to 2 cm of right-lateral slip and terminated near the hinges of the fold. The (mechanical) hinges separate right-lateral from left-lateral shearing on the upper and lower surfaces of the layer.

A.

B.

Figure 3 (from Pfaff, 1986a) A. Map of traces of fractures in monoclinial fold. Localization of bedding-parallel fractures in inner limb (Mapped by A.M. Johnson). B. Detail map of hinge marked with a box in A. (Mapped by V. Pfaff).

SPOTTED WOLF SECTION OF SAN RAFAEL SWELL

A final example of a delaminated inner limb of a monocline is in the Navajo Sandstone at San Rafael Swell, Utah. The San Rafael Swell is an asymmetric, domical uplift on the Colorado Plateau located in central Utah, stretching about 120 km north and south and 50 km east and west (Fig. 4) with stratigraphic offset on the order of 1 km. Uplift has been determined to have initiated in latest Campanian time (late Cretaceous) (Lawton, 1983) and is believed to be related tectonically to the Laramide basement uplifts of the Rocky Mountain foreland (eg. Cross, 1986). The fold is structurally similar to other major uplifts on the Colorado Plateau (Kelley and Clinton, 1960). The beds of the late Paleozoic through Mesozoic rocks comprising the Swell dip gently westward over the center and west edge of the uplift and a monoclinal fold forms the eastern flank of the Swell where the beds dip steeply eastward and are locally overturned. The limb is characterized morphologically by dramatic flat-irons of Navajo Sandstone (Fig. 5).

Fig. 4 Geologic setting of Spotted Wolf section of San Rafael Swell

The Navajo Sandstone beds within the inner limb of the monocline have a wide range of dip angles and radii of curvature along the Spotted Wolf Section shown in Fig. 4. The Spotted Wolf section of the monocline extends 10 km from Spotted Wolf Canyon (at Interstate Highway 70) in the north to Iron Wash in the south. The radius of curvature of the synclinal hinge ranges from 760 m, near I 70, to 40 m near midlength of the section. At both ends of the Spotted Wolf section, the fold hinges are wide open, the monocline is broad, and dips are relatively shallow, 30° or less (Fig. 5A). In the middle part of the section, the fold hinges are much tighter, the monocline is much narrower and the beds within the inner limb dip much more steeply and are overturned in some places (Fig. 5B).

The Navajo Sandstone is a massive, cross-bedded sandstone. Bedding planes are difficult to locate, in part because beds can be as thick as several tens of meters, but large low-angle cross-beds give the Navajo a layered appearance. Small faults formed parallel to cross-bedding in places in the inner limb of the monocline (Mark Fisher, pers. comm., 1998; Johnson and Johnson, in prep; Johnson, in prep). The frequency and occurrence of these faults vary from place to place along the Spotted Wolf section and appear to correlate with tightness of the synclinal hinge. For example, small cross-bed-parallel faults are scarce at the northern end at I 70 where roadcuts provide nearly perfect exposures, and they are non-existent at the southern end of the section where the beds dip about 10°. The radii of curvature of the synclinal hinges are on the order of 1000 m in each area. In

contrast they are abundant in steeply dipping beds where the synclinal hinge is tight, with a radius of curvature of 50 m.

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Fig. 5 A. Aerial view of southern end of Spotted Wolf section of San Rafael monocline. Beds dip steeply in foreground and shallowly in background. B. Aerial View of middle part of Spotted Wolf section. Prominent white flat-iron rock is Navajo Sandstone.

Fig. 6 is an interpretive cross section of the monocline based on the observable geology and radii of curvature measured in the field. The cross section (marked A-A' in Fig. 4) is located the Navajo beds dip 60° and the radius of curvature of the synclinal hinge is 40 m. The location of observed cross-bed-parallel faults in the Navajo sandstone is indicated in the figure.

Fig. 6 Interpreted cross section (A-A' in Fig. 4) of San Rafael Swell based on observable geology and measured radii of curvature in the field.

ANALYSIS OF HIGH-AMPLITUDE KINK FOLDING

Two of the field examples presented above, the Amity Hall fold and the Watts fold, are characteristic of monoclinial kink bands. They are isolated monoclinial folds with an inner limb that has been reoriented relative to the layering in the two outer limbs. The fold hinges are quite sharp and the three limbs are quite straight. In both of these examples, the inner limbs were thoroughly delaminated.

We will show that the delaminations in the field examples can indeed be explained theoretically in terms of location of maximum layer-parallel shear stresses. We will consider the general case that includes high-amplitude folding, so the analysis presented here supercedes the low-amplitude solution for the kink form (Honea and Johnson, 1976; Johnson, 1977, Chapt. 6). Previous analyses of the kink form have been based on low-amplitude folding theory and the analyses become seriously in error if the fold limb dips exceed 10° (Johnson and Fletcher, 1994, Chapt. 5). In the Amity Hall and Watts folds, the inner limbs are clearly dipping more than 10° relative to the outer limbs, so the low-amplitude solution is inappropriate for analyses of these folds.

We will therefore present a new, high amplitude analysis that is not restricted to low dip angles. The drawback is that there is no analytical solution to the resulting differential equation so it must be solved numerically. Our method of solution is suggested by Euler's *Elastica* for a free elastic beam (e.g., Johnson, 1970) but it is novel in the treatment of contact strength.

Differential Equation Used for the Kink Fold Form

To obtain the differential equations used for the kink-fold form we consider force and moment equilibria of the element of a layer shown in Figure 8. The element is bounded by circular segments and radii so that the analysis is valid for high-amplitude folding. Acting on the element are bending moments, M , and shear stress, τ_{avg} , on each side, and shear stress equal to shear strength, τ_{cont} , at top and bottom contacts, representing top and bottom of the layer.

Figure 8 A. Single layer and surrounding medium. B. Small element of a single layer subjected to horizontal compression, P , bending moments M , and contact strength τ_{cont} .

We can derive an expression for the average shear stress, τ_{avg} , using notation of high-amplitude beam/plate theory (e.g., Timoshenko and Gere, 1961; Johnson, 1970) with the standard equilibrium equations for the infinitesimal circular element shown in Fig. 8B. Summation of moments about the center of the element (Fig. 8B), provides the equation,

$$-\frac{dM}{ds} \tau_s + \{ \tau_{cont} (\tau_s + (\tau/2) \tau) + \tau_{cont} + \frac{\tau \tau_{cont}}{\tau_n} T \} (\tau_s - (\tau/2) \tau) \{ (T/2) b - \tau_{avg} \tau_s T b = 0$$

or

$$-\frac{dM}{ds} / T b + (\tau_{cont} - \tau_{avg}) + \frac{\tau \tau_{cont}}{\tau_n} \{ (T/2) - (\tau/2)^2 (\tau \tau_s) \} = 0. \quad (2a)$$

We will assume that the contact strength, τ_{cont} , is the same on the top and bottom of the layer so that $\frac{\tau \tau_{cont}}{\tau_n} \neq 0$. Then after rearranging, eq. (2a) reduces to

$$\frac{1}{bT} \frac{dM}{ds} + \tau_{avg} - \tau_{cont} = 0. \quad (2b)$$

This result is the relation between change in bending moment with position, s , along a layer, the average shear force, $bT\tau_{avg}$, on a cross section and a shear force applied to the surfaces of the layer, $bT\tau_{cont}$, due to contact strength (e.g., Koch et al., 1981).

In order to complete the analysis, we sum forces in the n -direction on the element in Fig. 8 and set the result equal to zero,

$$\frac{\tau_{avg}}{s} bT - P \cos \theta + \tau_{nn} \{ (s + (T/2)) - (s - (T/2)) \} b - \tau_{nn} T (s - (T/2)) b = 0$$

or

$$\frac{\tau_{avg}}{s} \frac{F}{bT} \cos(\theta) \frac{d\theta}{ds} + \frac{\sigma_{nn}}{n} \frac{T}{2} \frac{d\theta}{ds} - \sigma_{nn} \frac{d\theta}{ds}.$$

Integrating with respect to s ,

$$\tau_{avg} = \sigma_a \sin \theta + q, \quad (3a)$$

where $\sigma_a = \frac{P}{bT}$ is the axial stress and $q = \frac{\sigma_{nn}}{n} \frac{T}{2} \frac{d\theta}{ds} - \sigma_{nn} \frac{d\theta}{ds}$ is the reaction of the medium.

Equation (3a) shows that the layer-parallel shear stress increases with the inclination, θ , of layering, the axial stress, σ_a , and the reaction of the medium, q . We will neglect the reaction of the medium in this problem, thus we assume that the kink band is long (as in Honea and Johnson, 1976), so that

$$\tau_{avg} = \sigma_a \sin \theta \quad (3b)$$

Thus the relevant differential equation combines eqs. (2b) and (3b),

$$\frac{BI}{bT} \frac{d^2\theta}{ds^2} + \sigma_a \sin \theta - \tau_{cont} = 0. \quad (4)$$

Solution for High-amplitude Kink Folds

We begin the solution by multiplying both sides of eq. (4) by $2d\theta/ds$,

$$2 \frac{d\theta}{ds} \frac{d^2\theta}{ds^2} \frac{BI}{bT} + 2\sigma_a \sin \theta \frac{d\theta}{ds} - 2\tau_{cont} \frac{d\theta}{ds} = 0, \quad (5a)$$

and then integrating with respect to s ,

$$\frac{1}{2} \frac{d\theta}{ds} \frac{BI}{bT} + 2\sigma_a \cos\theta + 2\tau_{cont} = c. \quad (5b)$$

We can replace the arbitrary constant, c , with a definite constant, α , the slope angle at the inflection point. The curvature, $\frac{d\theta}{ds}$, is necessarily zero at the inflection point where $\theta = \alpha$ (Fig. 9), so eq. (5b) provides the relation

$$c = 2\sigma_a \cos\alpha + 2\tau_{cont}$$

so that eq. (5b) becomes

$$\frac{BI}{bT} \frac{1}{2} \frac{d\theta}{ds} + 2\sigma_a (\cos\theta - \cos\alpha) + 2\tau_{cont} (\theta - \alpha) = 0. \quad (5c)$$

In order for the fold to be a kink form, we must require the curvature and slope angle be zero at the edges of the kink form. Letting $d\theta/ds$ and θ be zero in eq. (5c), we find that the contact strength, axial stress and slope angle at the inflection point are related by

$$\tau_{cont} = \sigma_a \frac{1 - \cos\alpha}{\alpha} \quad (5d)$$

so that eq. (5c) can be written in nondimensional form as

$$\frac{1}{2} \frac{dT}{ds} - 6 \frac{1}{B} [\cos\theta + \frac{1}{\alpha}(1 - \cos\alpha)] = -6 \frac{1}{B} (1 - \cos\theta). \quad (5e)$$

We can solve eq. (5e) numerically along with

$$\frac{dy}{ds} = \sin\theta; \quad \frac{dx}{ds} = \cos\theta$$

to obtain the kink-fold forms shown in Fig. 9.

Figure 9. Kink-band fold form for various axial loads. $\sigma_a^* = \sigma_a / B$ where σ_a is the axial stress given in eq. (5d).

This completes the analysis of the high-amplitude kink fold.

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