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Techniques for Identifying Faults and Determining Their Origins

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ABSTRACT

Regulatory criteria for siting nuclear power plants require that faults be characterized as to their potential for generating earthquakes and causing surface deformation, or that the absence of the potential for these occurrences be demonstrated. Satisfying these criteria requires the ability to distinguish between tectonically induced faulting, shallow faulting induced by strong ground motions, and faulting caused by nontectonic phenomena. Nontectonic faults can produce ground deformation but are not capable of producing significant earthquakes and vibratory ground motion (i.e., they are nonseismogenic). These faults often have physical characteristics similar to those of tectonic faults, but they differ in terms of their causative forces and potential hazard. Nontectonic faults, which are driven predominantly by gravitational forces, include those produced by slope failure processes (e.g., landslides), dissolution phenomena (e.g., karst collapse), evaporite migration (e.g., salt domes and salt flowage), volcanism (e.g., dike-emplacement and caldera collapse), sediment compaction (e.g., growth faults, subsidence), and unloading phenomenon (e.g., pop-ups). Tectonic faults, which may or may not be seismogenic, include primary structures capable of producing earthquakes (i.e., seismogenic faults) and secondary structures that are produced by earthquakes but are not themselves capable of generating an earthquake (i.e., nonseismogenic faults). Nonseismogenic tectonic faults include secondary deformation in the hanging-wall above a blind thrust fault and strong ground motion phenomena (e.g., ridge-crest shattering, basin-margin fracturing).

An understanding of the geologic, geomorphic, and tectonic processes that result in surface deformation is essential for developing criteria to identify and evaluate the seismogenic potential of faults. It is critical to document the characteristics and relationships of faults in a systematic and explicit manner such that a defensible conclusion may be reached regarding the causative process. In this report, we (1) summarize the characteristics of faults resulting from tectonic and nontectonic mechanisms, focusing on diagnostic characteristics that can be used to assess their origin; and (2) develop criteria to identify and differentiate tectonic and nontectonic faults. We find very few exclusively diagnostic criteria to differentiate tectonic from nontectonic faults. Determining the geologic context of a fault by integrating a variety of observations and data provides the best method for differentiating and documenting the origin of a fault. Observations and measurements of scale, geometry (plan view and cross sectional), and timing (recurrence, rate of deformation) are the most important attributes to understand in order to confidently assess the origin and, thus, potential hazard of a fault.

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1 INTRODUCTION

Regulatory criteria for siting nuclear power plants require that seismic sources be characterized as to their potential for generating earthquakes and causing surface deformation, or that the absence of the potential for these occurrences be demonstrated (U. S. Nuclear Regulatory Commission [NRC], 1994, 1996). Satisfying these criteria requires the ability to distinguish between tectonically induced primary and secondary faulting, faulting induced by strong ground motions, and faulting caused by nontectonic phenomena. Nontectonic faults may have similar physical characteristics to tectonic faults, but differ greatly in terms of their origin and potential hazard. Therefore, differentiating between tectonic and nontectonic faults is critical to the assessment of seismic hazards for nuclear facilities.

During the past few decades, research in the fields of paleoseismology, seismology, geodesy, and geophysics has greatly increased our understanding of the earthquake process and our ability to characterize earthquake sources for seismic hazard analyses. Paleoseismology involves the use of geologic and geomorphic techniques to determine the age, frequency, magnitude, and slip of prehistoric earthquakes, in addition to mapping coseismic surface faulting and secondary deformation from historical earthquakes (e.g., Wallace, 1977; Allen, 1986; Crone and Omdahl, 1987; Schwartz, 1988; Reiter, 1990). A perspective on the range of recent research being conducted worldwide in the field of paleoseismology is provided in Yeats and Prentice (1996). Recently published textbooks by McCalpin (1996) and Yeats et al. (1997) describe paleoseismologic methods and investigative approaches.

Standard paleoseismic studies have been applied successfully at many sites in the western United States (see Schwartz, 1987, and Weldon, 1991, for summaries of many of these studies) to evaluate fault capability as defined by the NRC. A more difficult challenge is posed for assessments of fault activity in stable continental regions (SCRs), such as the central and eastern United States, where the tectonic signature of a fault may be masked by surficial processes (erosion, chemical weathering, deposition) in areas of low rates of tectonic activity or when faults do not reach the Earth's surface. Evidence suggests that the low rate of tectonic activity in SCRs may be reflected by "temporal clustering" of surface-rupturing events preceded or followed by long periods of inactivity (Kelson and Swan, 1990; Adams et al., 1991; Crone and Machette, 1994). In such tectonic settings it may be difficult to demonstrate the mechanism by which a fault is formed and the timing of the most recent event, both of which are critical to the evaluation of fault capability as defined by the NRC.

Faults that do not reach the surface (commonly referred to as buried or blind faults) also have been the focus of increasing concern and research in the past few years. The occurrence of several moderate- to large-magnitude earthquakes in California caused by reverse motion along buried or blind faults (including the 1982 New Idria, M_w 6.4; 1983 Coalinga, M_w 6.5; 1985 Kettleman Hills, M_w 6.1; 1987 Whittier Narrows, M_w 5.9; and 1994 Northridge, M_w 6.7 events) emphasizes the need to identify and characterize these earthquake sources for seismic hazard analyses. Although the primary fault may remain blind (i.e., not reach the surface), these faults typically produce surface deformation (either folding or secondary faulting) in plate margin areas (Lettis et al., 1997). Recent research in plate margin areas has focused on methods to characterize the fault source at depth based on the nature and distribution of surface and near-surface deformation (e.g., Bullard and Lettis, 1993; Ekström et al., 1992; Angell et al., 1994;

Burgmann et al., 1994; Lettis et al., 1997; Shaw and Suppe, 1996, Kelson et al., 1998).

Blind faults also are of concern in SCRs such as the central and eastern United States. Although there is abundant geologic and seismologic evidence for earthquake activity, the 1989 Ungava, Quebec, earthquake is the only case for which primary surface rupture has been documented in North America (Adams et al., 1991). This may be because most SCR events in North America are small ($M < 5.0$) and lack the energy or are too deep to produce surface rupture. A review of worldwide reverse-fault earthquakes in SCRs (Lettis et al., 1997) shows that blind reverse faults *commonly* do not produce and are not associated with recognizable primary or secondary surface deformation. The largest known historical earthquakes in the central and eastern United States, including events within the 1811-1812 earthquake sequence in New Madrid, Missouri, and the 1886 Charleston, South Carolina, earthquake, occurred on faults that may not reach the surface. In the New Madrid area, near-surface secondary deformation (tilting and faulting) has been documented for the Reelfoot fault, a buried reverse fault that ruptured during the 1811-1812 sequence (Russ, 1982; Kelson et al., 1992; VanArsdale et al., 1995; Kelson et al., 1996a). Surface uplift and deformation in the Reelfoot Lake area of the New Madrid seismic zone appear to be associated with this structure. Regional Quaternary uplift in coastal South Carolina may be evidence for a buried causative fault for the 1886 Charleston earthquake (Marple and Talwani, 1993). However, there is no evidence that the primary rupture associated with the Charleston earthquake intersects the ground surface. Buried or blind faults also are postulated in other seismically active areas in the eastern United States. In the Eastern Tennessee and Giles County, Virginia, seismic zones, seismicity is thought to originate on faults in the basement beneath the Appalachian detachment at depths of greater than 5 km (e.g., Bollinger and Wheeler, 1983, 1988; Powell et al., 1994). The possibility of near-surface deformation or expression of these faults in the topography in these areas is a continued topic of research (e.g., Law et al., 1993, 1994; Mills, 1994; Granger et al., 1997). The challenge for seismic hazard assessments in such tectonic environments is to identify the location of tectonic blind faults, and to assess whether these faults are seismogenic.

Partly in response to these and other fault capability issues that have arisen during the past two decades, the Nuclear Regulatory Commission revised their criteria and methodologies for identifying and characterizing seismic sources. The new criteria and guidelines (10 CFR Part 100 Section 100.23 and Regulatory Guide 1.165) specifically require that all tectonic sources, including blind faults, that are capable of generating surface deformation and strong vibratory ground motions be identified and characterized (*see* Figure 1.1). As defined by the NRC, a tectonic structure is a large-scale dislocation or distortion, usually within the Earth's crust, the extent of which may be on the order of tens of meters (yards) to kilometers (miles). In applying the NRC criteria and guidelines, it is important to differentiate between faults produced by tectonic processes and faults produced by nontectonic processes. Characterizing a fault's seismogenic capability (the size of earthquake that may occur on the fault) also is critical.

Differentiating tectonic from nontectonic faults requires a thorough understanding of (1) the tectonic, geologic, and geomorphic processes that can lead to formation of a fault, (2) physical characteristics of the fault itself, and (3) an understanding of the geologic and tectonic setting (context) within which the fault occurs. As defined in this study, a tectonic fault is produced by deep-seated crustal-scale processes acting at or

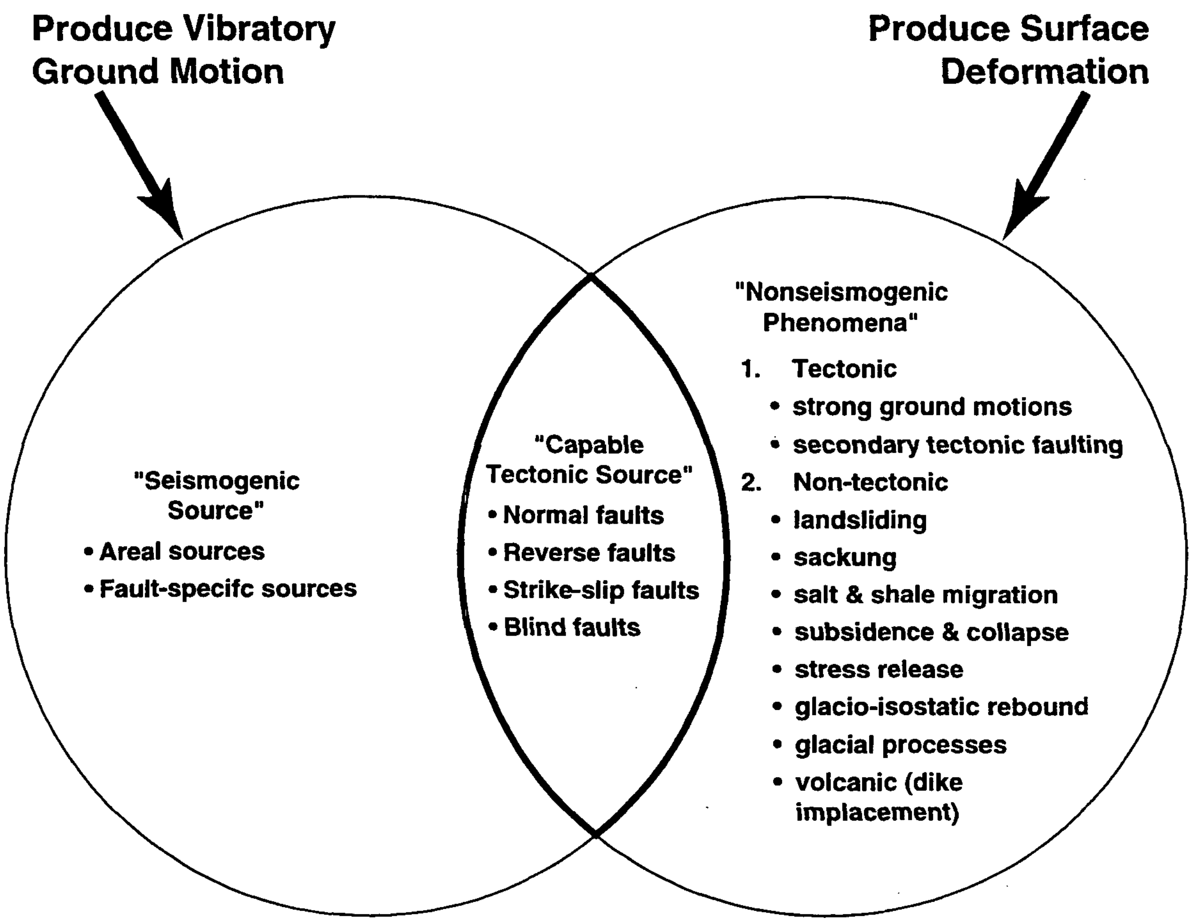


Figure 1.1 Schematic diagram showing the relationship among seismic sources, "seismogenic source," surface deformation, "nonseismogenic phenomena," and "capable tectonic source."

below seismogenic depths. Because the terms "tectonic" and "nontectonic" are mutually exclusive, a nontectonic feature is thus defined as a feature produced by shallow crustal or surficial processes acting above seismogenic depths. A seismogenic fault is defined as being capable of producing a moderate to large earthquake (M_W > 5), and a nonseismogenic fault is not capable of producing a moderate to large earthquake. These distinctions are important because of the presence of tectonic features, such as secondary ground cracks and margins of earthquake-induced landslides, which result from deep-seated earth movements but are not themselves capable of producing damaging earthquakes. As shown in Table 1.1, we identify seismogenic-tectonic features, nonseismogenic-tectonic features, and nonseismogenic-nontectonic features. Of primary concern to seismic hazard assessments are seismogenic-tectonic features, which can produce surface deformation as well as strong vibratory ground motions. Because nonseismogenic features may sometimes be interpreted as seismogenic features (and vice versa), a primary objective of this report is to provide criteria to help differentiate seismogenic-tectonic features from nonseismogenic features. Characteristics of seismogenic and nonseismogenic features are summarized in Sections 2.4, 2.5, 2.6 and 2.7 of this report. Faults produced by tectonic processes may or may not be seismogenic. Tectonic faults include both primary faults capable of producing earthquakes and secondary faults that are produced by earthquakes but are not themselves capable of generating an earthquake. Primary tectonic faults typically are classified into one of four categories based on sense of slip and fault geometry: strike-slip (transcurrent or transform) faults, normal faults, reverse faults, and blind or buried thrust faults, although the transitions between each category are gradational, and individual earthquakes on a specific fault may exhibit attributes of more than one category. Examples of secondary tectonic faults include hanging-wall deformations above a blind thrust fault and various types of strong ground motion phenomena (e.g., ridge-crest shattering, basin-margin fracturing). Examples of nontectonic faults include those produced by gravitational processes (e.g., landslide features, sackungen), dissolution phenomena (e.g., karst collapse features), evaporite migration (e.g., salt domes and salt flowage structures), sediment compaction (e.g., growth faults, subsidence structures), and isostatic adjustments (e.g., glacial rebound structures).

Table 1.1 General types of seismogenic and nonseismogenic features

	Seismogenic	**Nonseismogenic**
Tectonic	Seismogenic-tectonic faults Example: Active normal fault (Section 2.4)	Nonseismogenic-tectonic faults Example: Secondary tectonic faulting in hanging wall of blind reverse fault (Section 2.5)
Nontectonic	Seismogenic-nontectonic faults Example: none identified	Nonseismogenic-nontectonic faults Example: Growth faults related to salt diapirism and gravity glides (Section 2.6)

This report summarizes the results of a two-year study to develop a methodology for differentiating tectonic from nontectonic faults. The objectives of the study were to:

- summarize characteristics of and methods to identify seismogenic-tectonic faults, and
- outline procedures, develop criteria, and provide guidance to differentiate faults generated by tectonic processes (including primary and secondary faulting and other seismogenic deformation) from faults produced by nontectonic phenomena.

In the following sections, we briefly summarize the regulatory context and definitions developed by the Nuclear Regulatory Commission that pertain to this study, then discuss the approach and overall scope of the study.

1.1 Regulatory Criteria for Fault Capability

The recently revised 10 CFR part 100, Section 100.23, "Geologic and Seismic Siting Factors" (U.S. NRC, 1996), replaces Appendix A to 10 CFR Part 100 as criteria for siting future nuclear power plants. This regulation requires that seismic sources be characterized as to their potential for generating earthquakes and causing surface deformation, or that the absence of these potential occurrences be demonstrated. Section 100.23 requires that the geologic and seismic siting factors considered for design include a determination of the Safe Shutdown Earthquake (SSE) ground motion for the site, the potential for surface tectonic and nontectonic deformation, the design bases for seismically induced floods and water waves, and other design conditions. The proposed regulations also would require that uncertainty inherent in estimates of the SSE be addressed explicitly through an appropriate analysis, such as a probabilistic seismic hazard analysis or a suitable sensitivity analysis.

The definitions and criteria used by the NRC to identify seismic sources involve identification of both the origin and timing for most recent movement of faults within specified regions. A "capable tectonic source," as defined by the Nuclear Regulatory Commission (1996) (Regulatory Guide 1.165, Appendix A), is described by at least one of the following characteristics:

(a) presence of surface or near-surface deformation of landforms or geologic deposits of a recurring nature within the last approximately 500,000 years or at least once in the last approximately 50,000 years

(b) a reasonable association with one or more large earthquakes or sustained earthquake activity that usually is accompanied by significant surface deformation

(c) structural association with a capable tectonic source having characteristics of section **a** above, such that movement on one could be reasonably expected to be accompanied by movement on the other

Distinguishing between tectonic and nontectonic surface deformation is a key aspect of fault capability assessments. Satisfying the above criteria requires differentiating tectonic faults that are capable of generating earthquakes from nontectonic faults and tectonically induced secondary ground deformation features that are not capable of generating significant earthquakes. In past licensing studies, surface displacements caused by phenomena other than tectonically induced faulting (e.g., loading features related to glaciation or deglaciation, collapse structures related to karst terrain, or shallow, listric growth faults) have not been easily differentiated from tectonic deformation. Although nontectonic deformation, like tectonic deformation, can pose a ground deformation hazard to nuclear power plants, the methods and approaches to mitigate the two types of phenomena will differ. The more recent regulatory guidelines (Regulatory Guide 1.165, Appendix D) explicitly state that "if questionable features cannot be demonstrated to be of non-tectonic origin they should be treated as tectonic deformation."

1.2 Approach and Scope

An understanding of the tectonic and nontectonic processes that result in surface deformation is essential for developing criteria to identify and evaluate the seismogenic potential of faults produced by those processes. The objectives of this study, therefore, are to (1) provide an overview of the current, state-of-the-science knowledge on tectonic and nontectonic deformation; (2) summarize the characteristics of faults resulting from tectonic and nontectonic mechanisms, focusing on the identification of key diagnostic characteristics that can be used to assess their origin; and (3) develop criteria to identify and differentiate tectonic and nontectonic faults.

Our study consisted of the following activities:

(1) compile and review existing literature and data on tectonic and nontectonic faults

(2) convene a panel of individuals having expertise in neotectonic studies and investigating and characterizing specific nontectonic features (Table 1.2) to provide detailed information on the identification and characterization of tectonic and nontectonic faulting throughout the United States

(3) convene a technical workshop of interested professionals and researchers to elicit information, perspective, and direction from the scientific community regarding the issues of concern, and methods and criteria that can be applied to make these assessments of the character and seismic potential of faults (Appendix B)

(4) organize a session at the 1997 American Geophysical Union Spring Meeting to provide an opportunity for researchers, particularly from the central and eastern United States, to discuss case studies and research pertinent to this study (Appendix C)

(5) describe field relationships and/or characteristics and applicable techniques to differentiate tectonic and nontectonic surface deformation and to identify active blind faults

(6) develop recommendations as to how the analysis of tectonic versus nontectonic deformation should be applied in licensing nuclear power plant sites

The results of these studies are summarized in this report as outlined below:

- **Section 2.0** provides a generalized description of techniques commonly used to characterize seismogenic tectonic faults; descriptions of the various types of surface faults (primary tectonic, secondary tectonic, faults induced by strong ground motion, and nontectonic) and blind faults; and characteristics that can be used to identify them.

- **Section 3.0** provides a discussion of the criteria used to differentiate tectonic and nontectonic faults and the applicable techniques and methodology that should be employed to investigate these criteria at a site.

- **Section 4.0** discusses the application of these criteria to the assessment of seismogenic and/or ground rupture capability as defined by the U. S. Nuclear Regulatory Commission.

Table 1.2 Panel of technical experts

Expert	Affiliation	Areas of Expertise
Dr. David Amick	Science Applications International Corporation (SAIC)	Paleoliquefaction Secondary deformation Landslides
Mr. William R. Cotton	William Cotton and Associates, Inc.	Landslides Seismic hazards
Dr. Clark H. Fenton	Woodward-Clyde Federal Services	Glacio-isostatic (postglacial) faulting Neotectonics
Dr. Peter W. Huntoon	Department of Geology and Geophysics University of Wyoming	Gravity tectonics Salt tectonics
Dr. James P. McCalpin	GEO-HAZ Consulting, Inc.	Paleoseismology Sackungen
Dr. Frank J. Pazzaglia	Department of Earth and Planetary Sciences, University of New Mexico	Eastern U.S. neotectonics Tectonic geomorphology
Dr. Richard P. Smith	Idaho Nuclear Engineering Laboratory (INEL)	Paleoseismology and seismic hazards Evaluations in volcanic terrains
Dr. Bruno C. Vendeville	Bureau of Economic Geology University of Texas at Austin	Growth faults, salt tectonics, structural modeling

2 CHARACTERISTICS OF SEISMOGENIC AND NONSEISMOGENIC FAULTS

2.1 General Approach and Organization

A **fault** is a planar or subplanar discontinuity within a rock mass along which one side has moved relative to the other in a direction parallel to the discontinuity. A **fault zone** is a region consisting of many closely spaced or anastomosing faults. Because faults are produced by differential displacement, they commonly produce a shear fabric within the rock mass. Faults and fault displacement are most commonly described in terms of **slip** relative to either the ground surface, with movement described in terms of vertical and horizontal displacement (i.e., throw and heave) or the plane of the fault. **Dip Slip**, displacement in the dip direction and **strike-slip**, displacement in the strike direction, are the principal components of **net slip**, which is used when both the orientation and magnitude of the displacement vector are known. This terminology also forms the basis for a common fault classification that results in three simple types of faults (Figure 2.1): (1) **dip-slip** faults, along which the slip is approximately parallel to the dip of the fault surface; (2) **strike-slip** faults, along which the slip is approximately parallel to the strike of the fault surface; and (3) **oblique-slip** faults, along which the slip is inclined on the fault surface. Dip-slip faults can be further divided into normal and reverse slip faults according to the relative displacement of the blocks on either side of the fault. Normal faults are those in which the hangingwall block moves down relative to the footwall block; reverse faults are those in which the hangingwall block moves up relative to the footwall block. Strike-slip faults are right-lateral, or dextral, if the fault block across the fault from the observer moved to the right; they are left-lateral, or sinistral, if that block moved to the left (Figure 2.1). Oblique-slip faults are described according to the relative components of strike-slip and dip slip. In addition, many dip-slip faults are not exposed at the earth's surface and are called buried or "**blind**" faults.

This categorization of faults is valid regardless of the process producing the fault displacement. Each basic type of fault may be formed by tectonic as well as nontectonic processes. For the purposes of this research, we define **tectonic faults as those which** formed as a result of deep-seated, crustal-scale tectonic processes, and **nontectonic faults as those** which formed as a result of shallow crustal or surficial processes. We further subdivide tectonic faults into **seismogenic tectonic faults**, which are capable of producing significant ($M_W \geq$ 5.0) earthquakes, and **nonseismogenic tectonic faults**, which are not capable of producing significant earthquakes.

This chapter is organized according to this categorization of faults. We first describe the characteristics of faults, without regard to whether the faults are produced by deep-seated tectonic processes or by other, nontectonic processes. This is followed by a general description of investigative techniques commonly used to characterize capable faults. We then provide an overview of each of the primary types of seismogenic tectonic faults (i.e., normal, reverse, strike-slip, blind), nonseismogenic tectonic faults, and nontectonic faults. Nonseismogenic faults discussed below include those related to tectonic processes (e.g., secondary faults, and features related to strong ground motions and triggered fault slip), as well as those related to nontectonic processes (e.g., landslides, sackungen, salt-related phenomena).

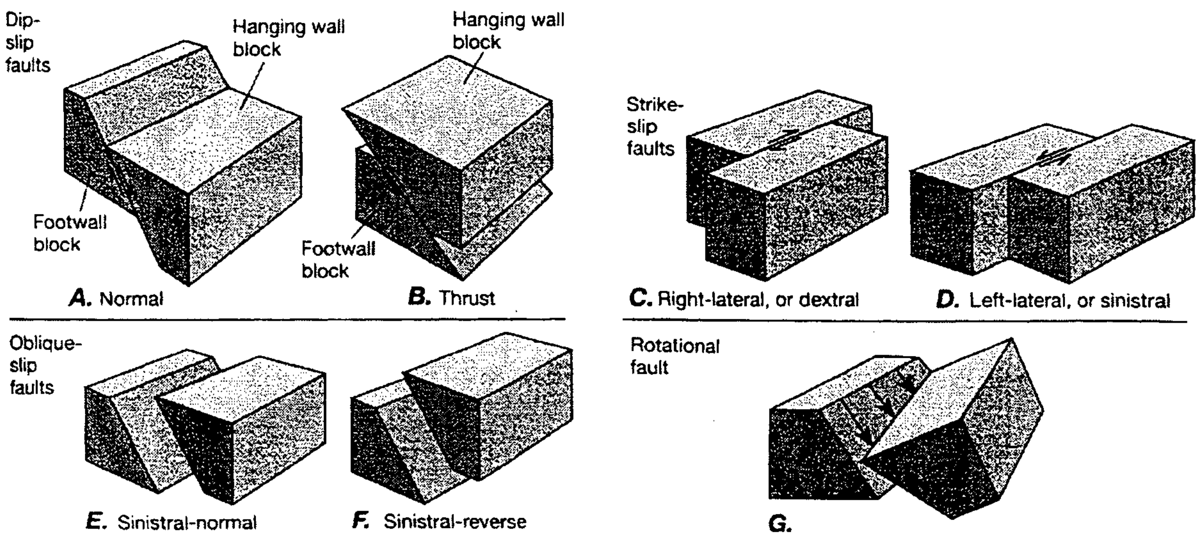


Figure 2.1 Faulted blocks showing the characteristic displacement for the different classes of faults From: STRUCTURAL GEOLOGY by Twiss and Moores © 1992 by W.H. Freeman and Company. Used with permission.

2.2 Recognition of Faults

This section describes the characteristics of normal, reverse, and strike-slip faults, without regard to whether the faults are produced by deep-seated tectonic processes or by nontectonic processes. Dip-slip, strike-slip, and oblique-slip faults each have several similarities and differences in basic properties that allow their identification and characterization. Twiss and Moores (1992), for example, divide the criteria for recognizing faults into three broad categories: (1) features intrinsic to faults themselves, (2) effects on geologic or stratigraphic units, and (3) effects on physiographic features. To this list, we add (4) contemporary seismicity, and (5) geodetic evidence of active deformation. The following section addresses these five categories.

2.2.1 Features Intrinsic to Faults

Faults commonly are associated with characteristic textures and structures that develop as a result of shearing (e.g., Sibson, 1977; Wise et al., 1984; Twiss and Moores, 1992). The textures and structures formed along a fault vary with the amount and rate of shear strain, with the physical properties of the rocks involved, and with the environmental conditions under which the faulting occurred. Analysis of fault-rock lithology and texture, therefore, can provide important information on the conditions of faulting (e.g., depth, temperature, fluid pressure). For example, faults formed at depths of less than about 10 to 15 km typically produce **cataclastic rocks** in the fault zone, whereas faults formed at greater than about 10 to 15 km or at temperatures more than about 250° to 350°C typically produce **mylonitic rocks** in the fault zone. **Cataclasites** result from brittle deformation by elasto-frictional fracturing and comminution of the rock into clasts or powder, with clasts typically being angular and internally fractured. These fault rocks include coarse-grained breccia and fine-grained gouge, and can range in thickness from millimeters to kilometers thick. **Mylonites**, in contrast, form primarily as a result of ductile deformation by crystal-plastic mechanisms and recrystallization of mineral grains, and often is accompanied by cataclasis. Crystal plasticity is a temperature-dependent process, and thus mylonites are produced deeper in the crust than cataclastic rocks. Thus, mylonitic rocks are formed as a result of deep-seated tectonic processes, whereas cataclastic rocks may from as a result of either deep-seated tectonic processes or other, near-surface processes.

Features within a cataclastic fault zone include slickensides, slickenside lineations or striations, and secondary mineral deposits. **Slickensides** are the smooth surfaces of fault planes often associated with shear-induced polishing of clay gorge or secondary mineralization. **Striations** are linear features present along fault planes that represent drag or tool marks due to irregularities in the fault surface (e.g., Hancock, 1985). These features commonly are associated with faults produced by both tectonic and nontectonic processes (Petit and Laville, 1987). The presence of **fault slices** or volumes of rock bordered on all sides by faults, clearly indicates fault displacement. Cataclastic faults and fault zones also often exhibit systematic arrays of fractures and internal secondary faults that have asymmetric and predictable geometries relative to the fault walls or fault-zone boundaries (e.g., Tchalenko, 1970; Chester and Logan, 1987; Petit, 1987). The predictable asymmetric nature of these secondary fault features provides a means to determine the relative direction of movement of the adjacent bounding blocks and are referred to as **shear criteria** or **kinematic indicators**.

Faults that develop at relatively shallow depths may develop open spaces within the fractured rock, which in turn may provide pathways for the flow of groundwater and hydrothermal fluids. This dilatancy and associated fracture porosity due to cataclastic faulting may undergo cyclic variations during dynamic conditions of earthquake rupture producing enhanced fluid flow along the fault (e.g., Sibson 1994; Muir-Wood and King, 1993). As a result, many faults may be associated with **secondary mineral deposits** within veins or as cement in preexisting fault breccia or gouge. The type and age of secondary fault fillings may provide information on the age and process that produced the brittle deformation. Careful description and documentation of these intrinsic structural features may be used to infer the existence of a fault even without direct evidence for displacement of stratigraphic or lithologic markers.

2.2.2 Effects of Faulting on Geologic or Stratigraphic Units

The presence of a fault is commonly identified from the juxtaposition of rock types that do not belong together in ordinary geologic sequences. **Discontinuity in geologic strata** may be a result of fault displacement, stratigraphic discontinuity (e.g., unconformity or facies change) or magmatic/diapiric intrusion. **Repetition of strata** or **omission of strata** in a known stratigraphic sequence is another indication of fault displacement (Twiss and Moores, 1992). **Drag folds** are structures formed in the region immediately adjacent to a fault surface due to frictional forces across the fault. For reverse faults, anticlinal drag folds form in the hangingwall and synclinal drag folds form in the footwall. The opposite relationships exist for normal faults. Two mechanisms are known to be responsible for the development of drag folds: (1) they may form solely due to frictional forces along a fault surface (e.g., Hatcher, 1994); and (2) they may form initially due to tip strains (flexures) near the leading edge of a fault during its growth, such as a fault-propagation fold, and are subsequently enhanced by continued fault development (e.g., Hancock and Barka, 1987; Hyett, 1990; Mitra, 1993). **Rollover folds** are fault-bend folds that form in the hanging wall of concave (listric) normal faults. Movement on listric normal faults results in non-uniform vertical displacement of the hanging wall strata due to decrease in the fault ramp angle with depth. For a given amount of dip slip, regions that overlay steeper sections of the fault will experience larger vertical displacement than those regions overlying shallow-dipping sections of the fault. This relationship between rollover and fault geometry can be used to infer the presence of listric fault geometries at depth (e.g., Gibbs, 1983; Xioa and Suppe, 1992). Other folds closely associated with faults are fault-bend folds, fault-propagation folds, and detachment folds. These structures are described later in this report under the section on reverse faults (Section 2.4.2).

2.2.3 Geomorphic Evidence of Faulting

Geomorphic evidence of fault activity includes features preserved on the landscape as the result of surface deformation (e.g., fault rupture, folding) and subsequent erosional or depositional processes that result from this deformation. Each major type of faulting — strike-slip, normal, and reverse — produces a characteristic assemblage of landforms. Geomorphic features indicative of late Quaternary fault activity include, but are not limited to, fault scarps, triangular facets, fault rifts, pressure ridges, shutter ridges, offset or deflected drainages, enclosed depressions or sag ponds, sidehill benches, aligned saddles, spring lines, vegetation lineaments, and linear drainages or troughs (Figure 2.2). These landforms range from centimeters to kilometers in scale from small scarps and fissures that develop at the time of surface faulting to large-scale geomorphic landforms such as escarpments and faceted ridge spurs along active mountain fronts, that result from repeated activity over tens to hundreds of thousands of years. Usually, combinations of these features are present if a fault has experienced repeated late Quaternary activity. The identification, delineation, and evaluation of geomorphic features associated with faults is an effective method for recognizing and characterizing active faults. Seismic source characteristics that can be obtained or estimated from quantification of geomorphic relationships include fault length, earthquake recurrence, displacement per event, and sense of fault slip. Geomorphic features also provide information on fault slip rate and the timing of most recent activity, if fault-related features are preserved on deposits or landforms of known age. Commonly, subsequent modification of the fault-related landforms due to high rates of erosion and/or sedimentation may obscure geomorphic evidence of recent displacement on active faults. Conversely, erosional patterns along inactive faults may produce landforms that mimic evidence of recent fault activity (e.g., linear drainages, fault-line scarps). In the early stages of identifying and mapping active faults for seismic hazard analyses, one should consider all forms of geomorphic expression in the search for potentially active faults.

2.2.4 Seismologic Evidence of Faulting

If a fault can be shown to have generated an earthquake during historic time, it is considered an active tectonic fault. However, the historical and instrumental records of earthquakes are short relative to tectonic processes and vary from region to region. Lacking instrumental coverage, confidence in associating an historical earthquake to a specific fault is highly dependent on anecdotal observations and written records shortly following an earthquake. The historical record of large earthquakes in the United States generally ranges from about 150 to 250 years. The historical record includes manuscripts, news or book accounts, personal diaries and, less commonly, verbal communications and legends. These records generally describe the effects of large earthquakes related to strong ground motions, but also may include reports of surface faulting, fissuring, or other phenomena that directly relate an earthquake to a specific fault. On a worldwide basis, the historical record varies from two to three thousand years in some regions (e.g., China, Japan and the eastern Mediterranean), to less than 100 years in remote, less populated areas. Importantly, the absence of historical activity cannot be used as evidence of fault inactivity. This is because the historical record generally is too short and incomplete relative to the lengths of time between large earthquakes in most parts of the world.

The record of earthquakes recorded by modern instruments (i.e., seismometers) dates back to about 1900. The magnitude and location accuracy of the instrumental record is largely dependent on the number of seismometers, their sensitivity and spatial coverage (distribution density). An increasingly sensitive and standardized global network of seismographic stations, largely to monitor nuclear testing, had by the mid-1960s resulted in global maps of seismicity that illuminated tectonic plate boundaries worldwide, including individual faults. Regions with

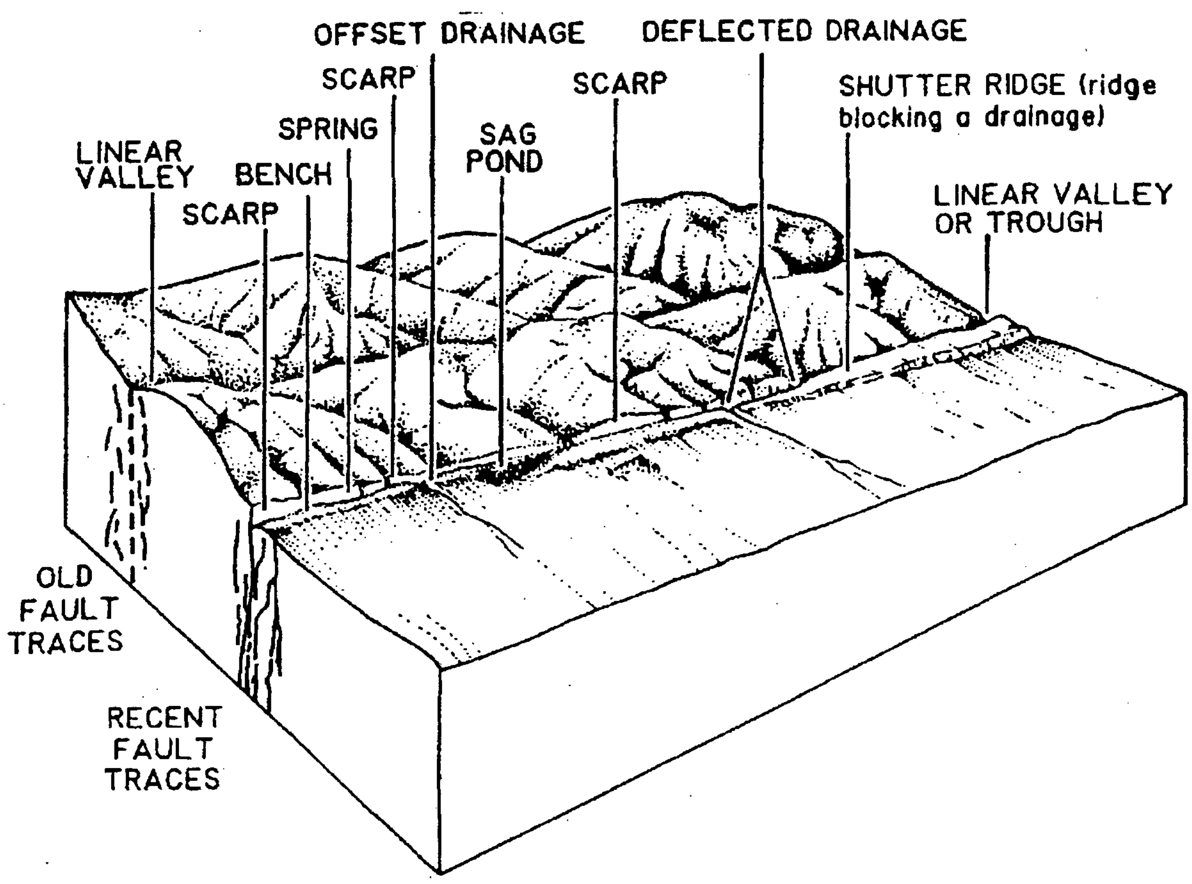


Figure 2.2 Block diagram of geomorphic structures associated with strike-slip faults (from Wesson et al., 1975).

moderate to high levels of seismic activity have received priority for the placement and maintenance of local arrays of seismic monitoring networks. Map and cross-section displays of recorded seismicity in these regions can be used to delineate individual active fault surfaces (e.g., the San Andreas fault in California and New Madrid seismic zone in the central United States).

However, in regions with low rates of seismicity, the instrumental record of earthquakes rarely is adequate in frequency or in the accuracy of epicentral locations to confidently identify an active fault. In these cases, the deployment of a local network to monitor and record instrumental seismicity may indicate active regional tectonic deformation and is valuable for assessing the rate of background seismicity, width of the seismogenic crust, and overall style of faulting in the region. It is important to note that the absence of instrumental seismicity also cannot be used to show that a fault is inactive. For example, the relatively low rate and broad distribution of contemporary seismicity in southern Oklahoma belies the occurrence of two large earthquakes within the past few thousand years (Kelson et al., 1990; Kelson and Swan, 1990; Swan et al., 1993). Even relatively active faults, such as the Wasatch fault in central Utah, which has an average Holocene slip rate of >1 mm/yr, has little associated instrumental seismicity (Arabasz et al., 1992).

2.2.5 Geodetic Evidence of Faulting

In the past two decades, significant advances have occurred in developing land-based and space-based geodetic data. Land-based geodetic surveys involve resurveying existing benchmarks and calculating their changes in relative location. Land-based geodetic studies involve measuring the change in relative position of two closely-spaced monuments that straddle an active surface fault (e.g., Thatcher, 1990), or re-leveling an entire sequence of stations that traverse a fold structure following an earthquake on an underlying blind-thrust to measure coseismic growth of the overlying fold structure (e.g., Ekstrom et al., 1992).

Space-based geodesy is a surveying technique that measures the three-dimensional location of a receiver station on the Earth's surface using an orbital network of satellites called the Global Positioning System (GPS). An areal GPS network that is continuously monitoring regional Earth movements can yield maps of crustal strain accumulation with 1 mm accuracy. Anomalous patterns of regional crustal movements determined from GPS studies are capable of measuring both coseismic regional displacements of the Earth's crust produced by earthquake slip (e.g., Hudnut et al., 1994), post-seismic displacements of crustal "relaxation" (e.g., Donnelan et al., 1994), and continuous aseismic regional displacements that represent the buildup of elastic strain in the upper crust (e.g., Donnellan et al., 1993).

2.3 Characterization of Capable Faults

Once a potentially active fault is recognized, an evaluation of capability is made by assessing its past earthquake behavior and current seismotectonic environment. An evaluation of past seismogenic behavior generally is conducted through paleoseismologic investigations, which typically include: (1) detailed mapping and delineation of fault traces via aerial reconnaissance, interpretation of aerial photography, and field mapping; (2) detailed stratigraphic analysis; (3) structural analysis of the observed deformation; and (4) detailed geomorphic analyses. In addition to mapping and use of aerial photography and satellite imagery, subsurface investigations, including exploratory trenching, drilling, and geophysical profiling also should be employed to conduct investigations of active faulting. Seismologic analyses and geodetic data also provide information that may be used to assess fault geometry and behavior. These studies are performed to characterize the following fault parameters:

- fault location, fault zone width, and fault trace complexity;
- fault geometry and dimensions, both along strike and downdip, including fault length, segmentation, strike and dip, and down-dip width in the crust;
- fault behavior, including sense of slip, average and maximum displacement per event, and slip rate;
- maximum earthquake potential; and
- recurrence interval between faulting events.

2.3.1 Investigative Techniques for Characterizing Surface Faults

2.3.1.1 Detailed Mapping

Detailed mapping is performed to accurately locate fault traces and to help estimate fault length and segmentation. Fault segments are defined by distinct changes in fault geometry (strike and dip), fault trace complexity (gaps or *en echelon* steps in the fault trace, anastamosing versus simple fault pattern, intersecting faults, etc)., and fault behavior (recency of activity, slip rate, sense of slip). Mapping also is used to identify sites for more detailed geomorphic analyses and subsurface paleoseismic investigations.

2.3.1.2 Detailed Geomorphic Analyses

In addition to being an excellent tool to identify and map active faults, as discussed above, geomorphic features can be used to assess past earthquake behavior on a fault. Geomorphic features such as stream channels, stream terraces, marine terraces, and glacial moraines form excellent, often well-preserved markers from which to assess fault slip rate, recency of activity, and the direction and amount of displacement during an earthquake. Higher scarps on progressively older stream terraces, for example, indicate recurrent fault movement. Measurements of scarp heights across these surfaces and estimates of surface ages based on soil development or other time-dependent properties may yield data on fault slip rate, and provide constraints on the timing of surface-rupture earthquakes.

2.3.1.3 Subsurface Investigations

Subsurface investigations often provide the most definitive information on fault location and fault behavior. Subsurface investigations include drilling large- and small-diameter boreholes, geophysical profiling, and exploratory trenching. Boreholes may be drilled to define the thickness and character of surficial deposits, and/or the depth and type of bedrock. These may be used to assess the presence or absence of fault-related deposits that may be used to evaluate earthquake timing. For example, Kelson et al. (1996a) collected data from large- and small-diameter boreholes to assess the age and location of a buried paleovalley margin, which provided a Holocene slip rate for the Calaveras fault in northern California.

Site-specific geophysical profiling may include many different techniques of investigation, including several configurations of seismic reflection and refraction designed to image various depth intervals, ground penetrating radar, magnetic surveys, and various types of electro-magnetic surveys. The primary purpose of these efforts commonly is to identify and provide preliminary characteristics of faults, folds, or fault-related deposits in the subsurface, without substantial ground disturbance. These profiles may provide critical data on fault location and geometry, and may be used to help choose specific locations for exploratory trenching.

Exploratory trenching is the most direct and most commonly used method for assessing paleoseismic fault activity. Sites for trenching are chosen carefully following geologic and geomorphic mapping, topographic profiling, and geophysical profiling, as mentioned above. Preferable sites include those with deposition of late Quaternary deposits across the fault trace and no episodes of erosion. Continuous deposition is preferred to provide a complete record of fault activity.

Periods of erosion and/or non-deposition may destroy or not preserve evidence of faulting.

The information exposed in trenches relevant to paleoseismic investigations consists of four main elements: (1) geologic units (rock, sediment); (2) surfaces (unconformity, diastem); (3) geologic contacts (stratigraphic); and (4) faults and fractures. One or more of these elements may undergo subsequent transformations, including weathering and soil formation, bioturbation (animal burrows, roots), liquefaction, subsequent folding or faulting, etc.

Eight stratigraphic and structural features commonly are used to recognize and evaluate the presence and timing of individual paleoseismic events (Figure 2.3). These include:

(1) deposits offset by a fault

(2) abrupt upward truncation of a fault strand, with younger faults extending to higher stratigraphic levels

(3) deposits and surfaces deformed by folding, tilting or warping

(4) colluvial wedge deposits formed by degradation of fault scarps

(5) transformed deposits including sediment sheared by faulting and liquefaction deposits

(6) systematic and abrupt or stepped increases in displacement downsection

(7) intruded material such as fissure fills and fault gouge

(8) fault planes exposed at the surface

In addition, lithologic variations across a fault, variation in thickness of stratigraphic units or soils across a fault, and open fissures or pockets along a fault plane may be used to identify paleoseismic events. Generally, the identification of two or more of these features are required to document the presence of a paleoseismic event and to determine the relative timing between past earthquakes.

2.3.1.4 Seismologic Analyses

Analyses of pre-instrumental and instrumental seismicity data provide important information on fault behavior. Focal mechanisms show the style and orientation of fault slip (Figure 2.4). Hypocentral depths provide information on the width of the seismogenic crust and, thus, maximum width of faulting in the crust, which in turn is used in estimating the maximum magnitude that can be expected along the fault. Where microseismicity and/or past historical earthquakes

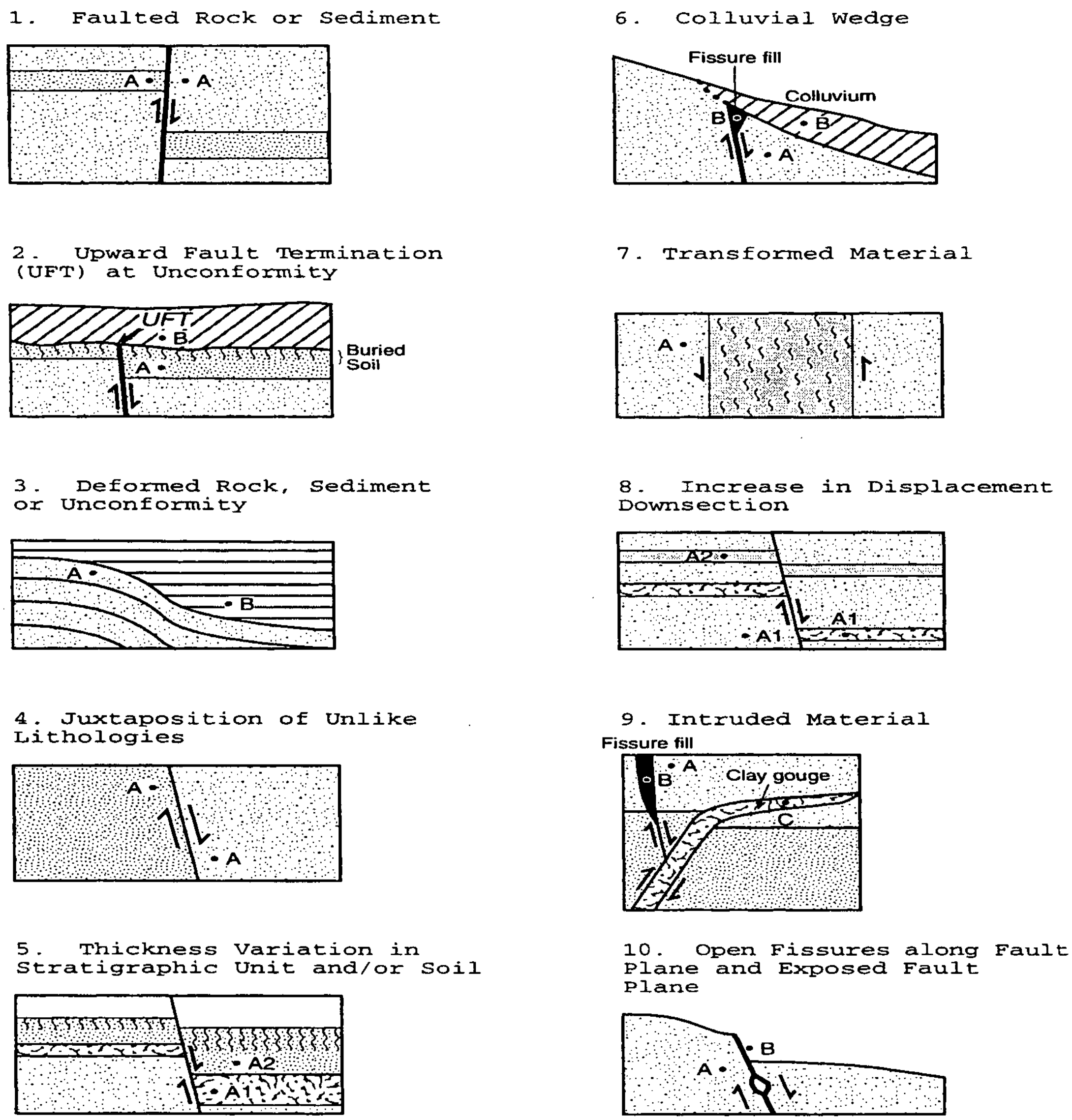


Figure 2.3 Schematic diagrams illustrating the principal stratigraphic and structural criteria used to identify the occurrence and timing of past earthquakes. Dated samples at locations A predate the earthquake. Dated samples at locations B post-date the earthquake. Dated samples at C are not helpful for deciphering chronology of past earthquakes because the sample may pre-date or post-date the earthquake depending on the geologic context. Dated samples at locations A and B, therefore, constrain the timing of the earthquake. Where evidence of multiple earthquakes is present, locations A1 pre-date the earlier event and locations A2 post-date the earlier event and pre-date the later event. From Lettis and Kelson, 1998.

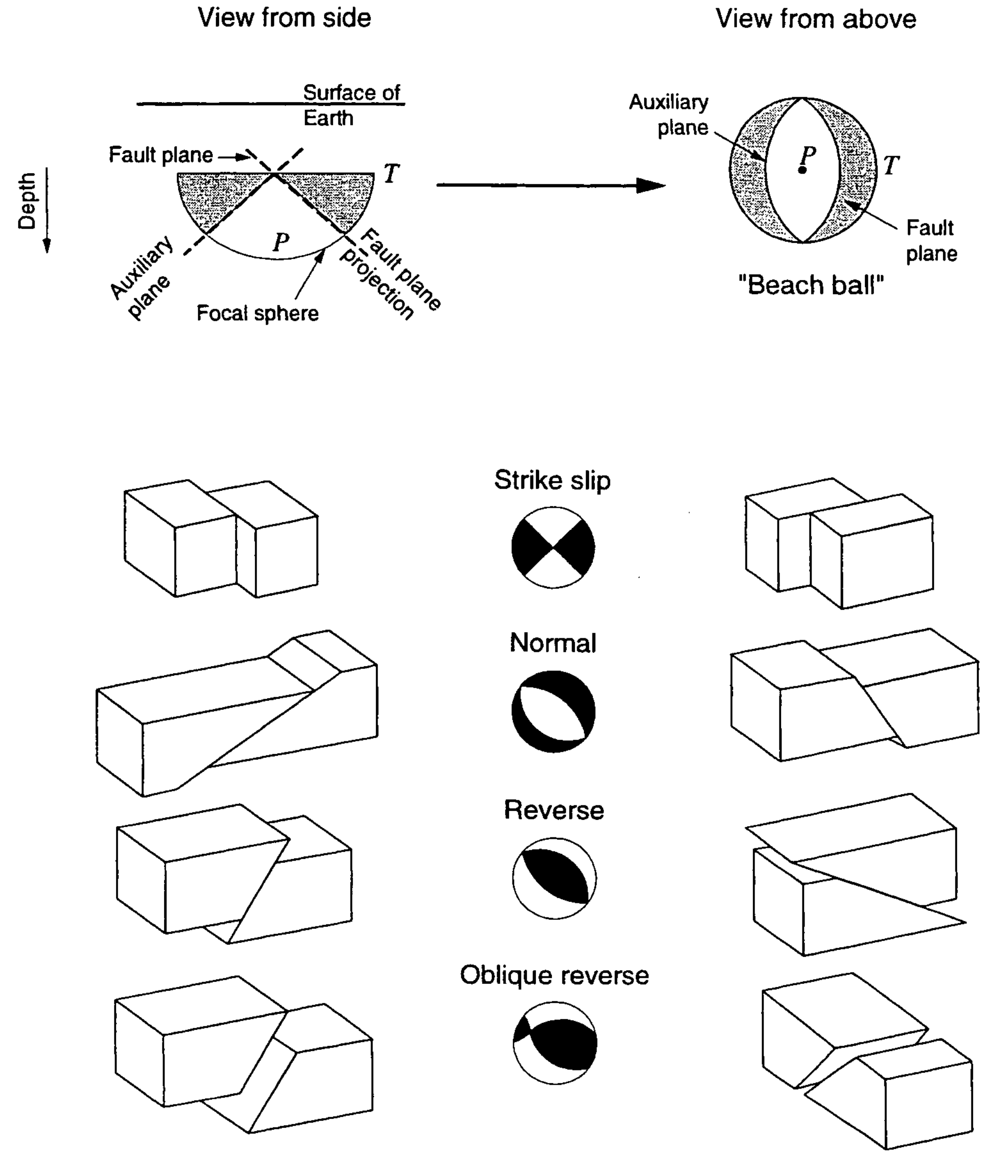


Figure 2.4 Schematic diagram of a focal mechanism. Reprinted from Walter et al., 1996.

occur on a specific fault, the seismicity data provide information on the rate of earthquake occurrence and, possibly, maximum earthquake potential. In general, however, seismicity data are used to assess the regional rate of background seismicity rather than fault-specific seismicity rates, primarily because the rate of seismicity or the accuracy of hypocentral locations commonly are inadequate for associating earthquakes with a specific fault plane or zone.

2.3.1.5 Geodetic Analyses

Geodetic arrays increasingly are used to define regional tectonic strain rates and rates of deformation across specific fault zones. Depending on the type of data, duration of recordings, and spacing of survey stations, rates of deformation can be detected with an accuracy of 0.1 to 1 millimeter per year. Geodetic data, for example, have been valuable in identifying locations of active faulting in the Walker Lane Belt and the Eastern Nevada Seismic Belt (Unruh et al., 1996). Although regional geodetic data seldom provide sufficient resolution to identify specific active faults, local geodetic surveys may be used to characterize a known active fault or fault system. Tectonic strain recorded by geodetic data can be used to identify active tectonic faults if the data resolution is sufficient, but cannot be used to define a fault as inactive, particularly in areas of low strain rates where faults may have long-term average slip rates of 0.1 millimeters per year or less.

2.3.2 Investigative Techniques for Characterizing Blind Faults

The role of blind-thrust faulting in earthquake hazards, particularly within the predominantly strike-slip tectonic domain of western California, came as a surprise to many geologists. There was an immediate response to apply the newly developed fault-bend fold theory to the assessment of seismogenic blind-thrust faults (e.g., Namson and Davis, 1988, Davis et al., 1989). This was accompanied by rapid development and application of innovative tools to locate and characterize blind-thrust faults. Some of these techniques existed already and were borrowed from other disciplines (e.g., geodesy, deep seismic reflection profiling), whereas others were developed specifically to characterize structures associated with active faults (e.g., tectonic geomorphology, growth fault-bend fold theory). All of the techniques when integrated with more traditional methods of geologic mapping and seismologic studies produce sophisticated and accurate interpretations of blind-thrust faults. This integration of techniques is extremely important because, when used alone, individual techniques commonly provide non-unique and highly interpretative solutions. In the following paragraphs we briefly describe the techniques most commonly used to characterize blind-thrust faults, and their role in seismic hazard assessment.

2.3.2.1 Subsurface Exploration Data

Analysis of deep borehole data and acquisition and interpretation of seismic reflection profiles are two methods of subsurface exploration developed by the oil and gas industry that have direct application to characterizing blind-thrust faults. Regional, deep borehole data can locate blind-thrust faults directly by penetrating the fault surface, or indirectly through obtaining structural data that define an overlying fold. Borehole data also provide detailed stratigraphic information that can be used to determine recency of faulting and slip rates (e.g., Yeats, 1988). Seismic reflection data have been used successfully to locate active blind-thrust faults by directly imaging the buried fault structure and indirectly by imaging youthful fold deformation resulting from displacement on the underlying fault (e.g., Namson and Davis, 1988; Medwedeff, 1989; Unruh and Moores, 1992; Unruh et al., 1997).

2.3.2.2 Balanced Cross Sections

The construction of balanced or retrodeformable cross sections also has proved to be a successful tool in locating and characterizing seismogenic blind-thrust faults in California (e.g., Davis et al., 1989; Unruh et al., 1992; Unruh, 1997; Angell et al., 1993; Shaw and Suppe, 1994). By application of fault-bend fold theory and restoration of the cross section to its configuration prior to onset of compressive deformation, the location, geometry and kinematic evolution of contractional structures along the line of section can be characterized. Although the solutions provided by the analysis are non-unique, when integrated with subsurface information such as seismicity, borehole, and seismic reflection data, the geometry and kinematics of the blind-thrust fault can be more tightly constrained. Also, the application of Quaternary geologic data provides valuable constraints on the timing and amount of shortening and uplift that can then be used to constrain the slip rate predicted by restoration of the cross section (e.g., Unruh et al., 1992; Bullard et al., 1994; Angell et al., 1997). The construction of regional balanced cross sections has been an important component to understanding seismic hazard in the Los Angeles basin area because significant contractional structures in that region are blind.

2.3.2.3 Seismologic Analyses

The distribution and kinematic nature of background microseismicity also may be used to characterize blind-thrust faults (e.g., Hauksson, 1990). Detailed analysis of the location and focal mechanisms of microseismicity in the Los Angeles basin region was used in conjunction with geologic data to outline the possible location of a regional system of blind-thrust ramps that are capable of generating large-magnitude earthquakes. An analysis of the seismic record of large reverse earthquakes worldwide by Lettis et al. (1997) is useful in demonstrating that regions of "blind" reverse faulting are most often characterized by recognizable surface deformation.

In the New Madrid, Missouri epicentral region of the 1811-1812 ~M8 earthquakes, the identification of a definitive fault source has been difficult using both geological (Van Arsdale, 1997; Kelson et al., 1996a) and geophysical methods (Hildenbrand, 1985; Hildenbrand and Hendricks, 1995; Langenheim and Hildenbrand, 1996). However, the excellent record of microseismic activity recorded in the epicentral region, primarily by Panda Network, provides valuable information on the structural character of the earthquake source. These data help define the mechanical process of generating large earthquakes and are used to characterize source geometry for seismic hazard analyses in the region.

2.3.2.4 Tectonic Geomorphology Analyses

Tectonic geomorphic analyses provide a means by which to assess the relative tectonic activity of a region (e.g., Bull and McFadden, 1977; Bull, 1984, 1990, 1991; Menges, 1988; Wells et al., 1988; Bullard and Lettis, 1993; Hitchcock et al., 1994). The assessments are based upon the quantification of geomorphic elements contained in the landscape that record a history of tectonic activity across long-term and short-term time spans. Therefore, these techniques are well-suited to characterizing regions underlain by blind faults. The geomorphic characteristics assessed are variations in the morphology of landscape components ranging from those generated as a direct result of fault rupture, such as fault scarps (e.g., Wallace, 1977; Nash, 1980, 1986; Hanks et al., 1984; Hanks and Andrews, 1989; Hanks, 1998; Arrowsmith, 1992; Avouac, 1993) and mountain fronts (e.g., Bull and McFadden, 1977; Menges, 1988), to components that result from the geomorphic modification of landforms indirectly related to tectonic activity (Bullard and Lettis, 1993). The latter include dissected mountain front facets (Menges et al., 1987), incised reaches of streams, and changes in stream channel pattern (Ouchi, 1985). Subtle characteristics in the landscape also result from surface folding related to displacement on structures in the subsurface and include deformed fluvial terraces, identifiable areas of anomalous topographic relief, and loci of deposition and erosion, loci of anomalous fluvial geomorphic parameters such as changes in stream slope and stream sinuosity (e.g., Bullard and Lettis, 1993; Hitchcock et al., 1994). Combined with detailed Quaternary geologic and geomorphic mapping, the tectonic geomorphology of a region provides powerful independent and supporting evidence for relative tectonic activity and variations in the tectonic framework such as fault segmentation, and styles of tectonic activity.

2.3.2.5 Paleoseismologic Analyses

Scarps and associated surface rupture produced by earthquakes on emergent thrust faults (e.g., the 1971 San Fernando earthquake) present an easy target for field-based investigative studies such as paleoseismic trenching that can provide direct access to information on past earthquakes recorded in the young deposits, including fault dip, sense of slip, stratigraphic offset and slip per event. Blind-thrust faults do not produce primary rupture scarps containing this information. However, secondary tectonic faulting associated with coseismic fold growth such as bedding-plane (flexural) slip faults and bending-moment faults have been unequivocally associated with coseismic folding events above blind-thrust faults (e.g., Yeats, 1986; *See* Section 2.5.2 Secondary Tectonic Faults). These structures may be expressed at the ground surface as recognizable geologic features that can be investigated by paleoseismologic trenching in the same way as a primary surface fault (e.g., Yeats, 1986; Cotton et al., 1990; Avouac et al., 1992, Treiman, 1994; Kelson et al., 1996a). Paleoseismologic investigations of the statigraphic relationships exhibited by unfaulted, tilted syntectonic sediments that overlie blind-thrust faults is currently being tested in the Los Angeles basin. This technique also may be used to characterize the location and temporal characteristics of blind-thrust faults (Mueller and Suppe, 1997).

2.3.2.6 Geodetic Analyses

Geodesy (measurement of the shape of Earth's surface) has emerged in recent years as one of the most promising tools for investigating tectonically active regions in general, and regions underlain by blind-thrust faults in particular (e.g., Feigl et al, 1993; Donnellan et al., 1993).

Coseismic surface displacements determined from both land-based and space-based geodetic studies can be used to model the geometry, location and slip of the underlying causative blind-thrust fault using an approach called dislocation modeling (e.g., Ekström et al., 1992; Marshall and Stein, 1994).

2.4 Seismogenic Tectonic Faults

Seismogenic tectonic faults include normal faults, reverse faults, strike-slip faults, and blind faults. Issues related to the characterization of these faults for seismic hazard assessment is provided in the following sections.

2.4.1 Normal Faults

Normal faults have dip-slip displacement in which the hanging wall has moved downward relative to the footwall. These faults are important sources of large destructive earthquakes in continental rifts and extensional tectonic provinces. Continental extensional settings include: (1) passive margins (e.g., the Atlantic passive margin), (2) discrete intracontinental rifts (e.g., the Rio Grande rift, the East African rift, the Baikal rift zone), (3) diffuse rifts or extensional provinces (e.g., the Basin and Range province, the Tibetan plateau rift system), (4) extension in strike-slip zones (e.g., pull-apart basins such as the Dead Sea rift, and basins along the San Andreas and Anatolian fault systems), and (5) extension in zones of net compression (e.g., continental extensional features formed in high plateaus or mountain belts, such as the Tibetan plateau and the Peruvian Andes) (Ruppel, 1995).

Theoretical models for extension of the lithosphere by continental rifting are described in terms of pure shear and simple shear. Pure shear involves coaxial horizontal extension of the crust and lithosphere that is coupled and symmetric about the spreading axis (McKenzie, 1978; Le Pichon and Sibuet, 1981). Extensional faulting occurs on normal faults that dip toward the rift axis and extend through the seismogenic crust. Crustal extension by magmatic dilation, which takes place primarily by intrusion of dikes parallel to the rift axis, generally is restricted to the region thinned by normal faulting. An alternative model proposed by Wernicke (1981, 1985) invokes non-coaxial simple shear as the dominant mechanism for deformation. In the simple shear model, crustal extension is achieved by heterogeneous, simple shear displacement along a low-angle detachment fault that extends through the crust. The detachment is rooted into a zone of mantle upwelling that is offset from the locus of extensional faulting in the upper crust. The area of magmatic intrusion, therefore, is significantly offset from the surface expression of intense extensional deformation. This basic element of the model results in a characteristic asymmetry in the simple shear rift system. In the simple shear model, much of the extensional deformation may be accomplished by aseismic slip along the low-angle detachment fault. According to both models, extension is achieved primarily by normal faulting in the upper crust and dike intrusion from a lithospheric mantle source in the lower crust. The relative timing and amounts of extension by intrusion or faulting vary among different rift systems.

Fault-bounded sedimentary basins, called **graben**, are a fundamental manifestation of continental extension (e.g., Bally, 1982; Wernicke and Burchfiel, 1982; Anderson et al., 1983; Jackson and McKenzie, 1983; Gibbs, 1984; Rosendahl, 1987). Like the large-scale modes of continental extension described above, graben structures have cross-sectional geometries that can be either symmetric (graben) or asymmetric (half-graben) depending on the relative amounts of slip on the faults that bound the basin. Half-graben are characterized by a master fault or fault zone that dips inward beneath the basin floor (Figure 2.5). Seismologic and geologic studies of normal-faulting earthquakes demonstrate that rupturing commonly occurs on multiple fault strands within half-graben (e.g., Bruhn and Schulz, 1996). The primary fault or rupture zone commonly break into multiple strands along splay faults that presumably intersect at depth (Bruhn et al., 1987, 1990, 1992). Antithetic faults within the interior and along the opposite margin of fault-bounded basins dip toward the primary fault zones.

The geometry of normal faults, as well as the processes of continental extension have been the focus of much research and debate. Recent review papers have been published that describe extensional processes (passive and active) in continental crust (e.g., Lister et al., 1991; Ruppel, 1995), the resulting structures formed (e.g., Thompson et al., 1989; Schlische, 1993), and the structural, fluid-chemical, and mechanical characteristics of seismogenic normal faults (Bruhn et al., 1990; Bruhn and Schultz, 1996).

In general, three styles of normal faults are important for accommodating extension of the brittle upper crust as inferred from geologic, seismologic, and geophysical data: (1) **steep, planar faults**, (2) **listric faults**, and (3) **low-angle detachment faults** (Anderson et al., 1983; Zoback and Anderson, 1983). A brief summary of the geometry of each of these styles of faults is provided below.

(1) Crustal-scale, moderately steep to steep, planar normal faults. Geophysical (e.g., Okaya and Thompson, 1985; Zoback and Anderson, 1983; Anderson et al., 1983), seismologic (e.g., Vetter et al., 1983; Smith et al., 1985; 1989; Arabasz and Julander, 1986; Jackson, 1987; Doser and Smith, 1989), and geodetic studies (e.g., Stein and Barrientos, 1985), principally within the Basin and

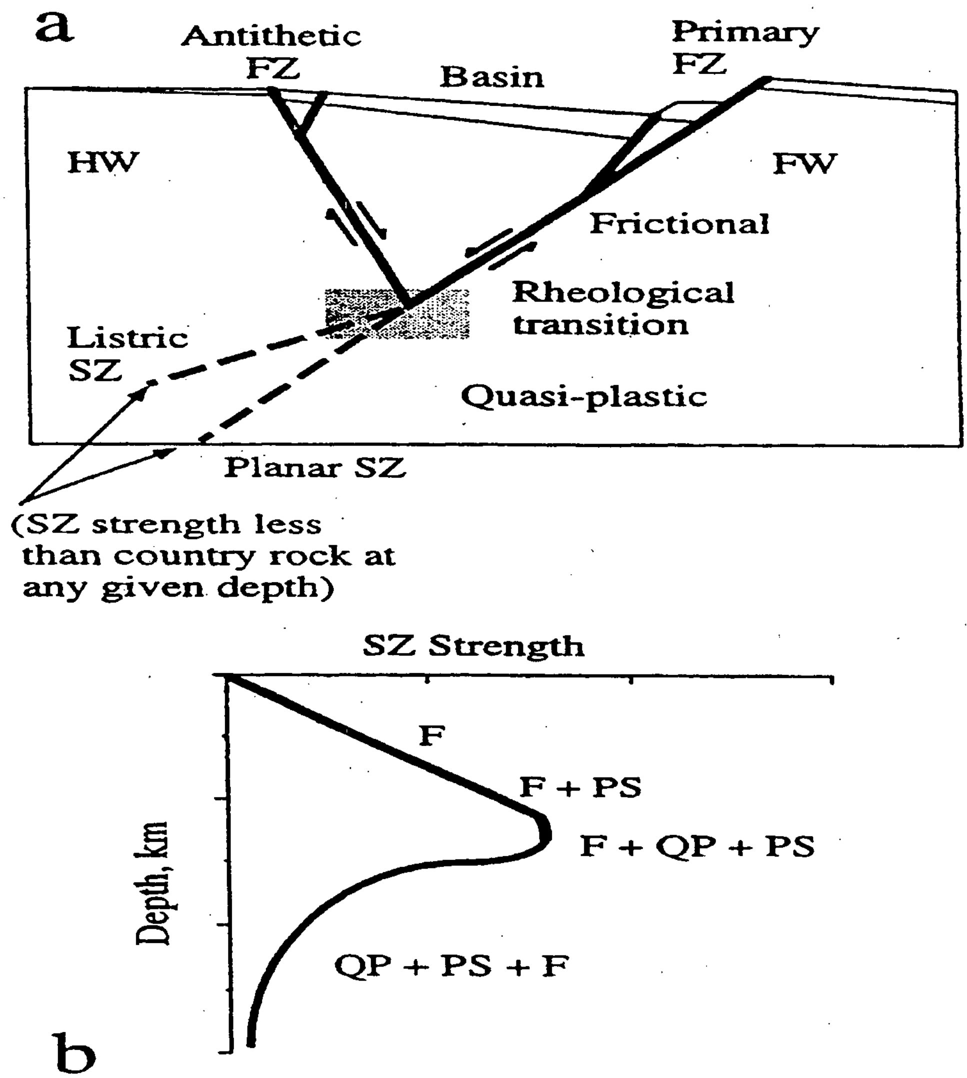


Figure 2.5 (a) Schematic cross section of a half-graben normal fault system showing primary and antithetic fault zones (FZ) extending from the surface through the rheological transition zone. Primary fault becomes a weaker, partially creeping shear zone (SZ) as crystal plasticity increases within and below the transition zone. Two geometries are indicated between the upper, frictional fault zone and lower, quasi-plastically deforming shear zone: case 1, shear zone coplanar with overlying fault zone; case 2, shear zone forms listric geometry. HW, hangingwall; FW, footwall. (b) Schematic diagram of fault zone strength and dominant deformation mechanisms as a function of depth. F, frictional sliding; PS, pressure-solution flow; PQ, quasi-plastic flow created by crystal plasticity. Two or more mechanisms may operate simultaneously within a depth interval. From Bruhn and Schultz, "Geometry and Slip Distribution in Normal Fault Systems: Implications for Mechanics and Fault-Related Hazards," *Journal of Geophysical Research*, 101-B2:3401-3412, 1996, © by the American Geophysical Union. Used with permission.

(2) Range province of western United States, have demonstrated that moderately steep to steep, planar normal faults extend to depths of about 15 km. Jackson (1987) and Jackson and White (1989) present a worldwide review of focal depths and fault-plane solutions for large normal-fault earthquakes on continents that demonstrates the overwhelming majority of earthquakes with magnitudes > 6.0 nucleate in the depth range of 6 to 15 km on faults dipping greater than 45° to 60°.

(3) Moderately to deeply penetrating listric normal faults. Listric fault geometries have been imaged on seismic reflection profiles in the Basin and Range (e.g., Anderson et al., 1983; Smith and Bruhn, 1984; Gans et al., 1985) and in the Rio Grande rift (e.g., Cape et al., 1983; Russell and Snelson, 1994). Such geometries commonly flatten into underlying zones of subhorizontal detachment within the upper 5 to 10 km of the crust (McDonald, 1976; Anderson et al., 1983; Van Tish et al., 1985). Listric boundary fault zones are known or inferred for many of the Mesozoic rift basins in eastern North America (Figure 2.6) (Schlische, 1992).

(4) Low-angle normal faults (detachments). Geologic reconstructions, thermochronology, paleomagnetism, and seismic reflection profiles indicate that low-angle normal faults in the upper continental crust are common in the geologic record (Wernicke, 1995). These detachment faults commonly are associated with metamorphic core complexes and have been recognized in oceanic as well as continental lithosphere (e.g., Mutter and Karson, 1992). Many authors have suggested that detachment faults pass downward into gently dipping crustal shear zones (Wernicke, 1981, 1985; Reynolds, 1985; Davis, 1983; Davis et al., 1983, 1986; Lister et al., 1984; Davis and Lister, 1988).

The relative importance of moderate- to high-angle (> 30° to 60°) and low-angle (<30°) faulting in accommodating large-scale extension has been the focus of much discussion in the past decade. Arguments supporting displacement on high-angle normal faults rely on Anderson's (1942) fault theory, the paucity of seismically active low-angle normal faults (Jackson, 1987), laboratory models (Brun et al., 1994), and observations of fault rotation over time from high angles to low angles (Buck, 1988, 1993). Thompson et al. (1989) conclude that high-angle normal faults may be the most important structures controlling extension in the brittle, upper crust of the Basin and Range. Proponents of extension accommodated by displacement along low-angle normal faults argue that such faulting does not violate Anderson's (1942) theory (Forsyth, 1992), that elevated pore fluid pressures and concomitant weakening in both the ductile and brittle regimes actually render low-angle faulting favorable (Axen, 1992), that some low-angle faults may indeed be seismogenic (Abers, 1991; Johnson and Loy, 1992), and that displacement on low-angle faults is the only logical way to accommodate the very large amounts of extension observed at passive margins and in the Basin and Range province (Wernicke, 1981; Scott and Lister, 1992). Wernicke (1995) presents a mechanical model relating fault dip to earthquake recurrence model that explains both the common occurrence of low-angle faults and the lack of large faults dipping more than 60°. This model argues for the initiation of and slip on low-angle normal faults, and suggests that the paradoxically low ratios of shallow and steep dipping focal planes to moderate ones in global seismicity may be resolved by a simple recurrence model, where the larger size and greater efficiency of shallow dip-slip faults cause them to fail much less frequently. This model contrasts with the conclusions of other geophysicists who argue that

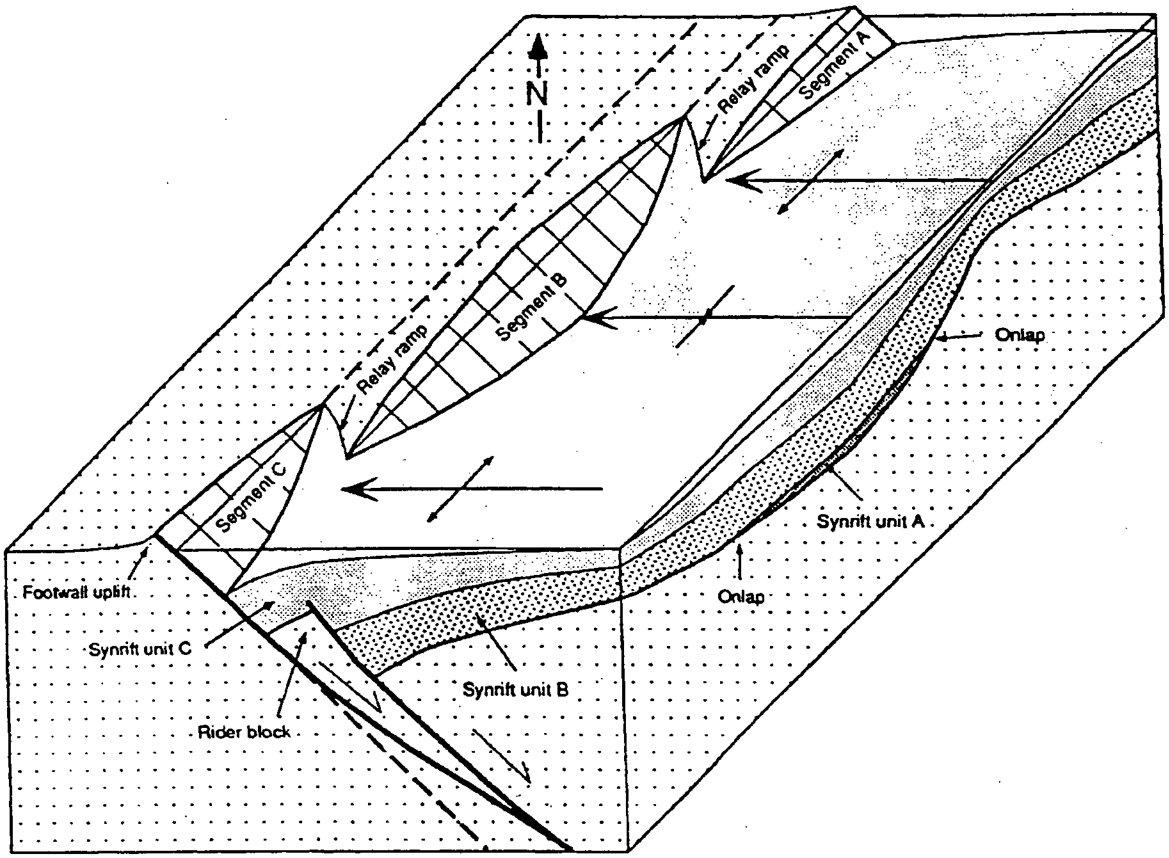


Figure 2.6 Block diagram illustrating the geometric relationships among border fault segmentation, relay ramps, transverse folds, and rider blocks. Synrift unit A forms a restricted wedge, suggesting that fault Segment B lengthened through time. Synrift unit B is absent from the hanging wall block of fault Segment C, suggesting that Segment C is younger that Segment B. Segments B and C may merge at depth, forming ends of Segments B and C if they involve only partial deactivation of preexisting structure. From: R.W. Schlische, "Anatomy and Evolution of the Triassic-Jurassic Continental Rift System, Eastern North America," *Tectonics*, 12-4:1026-1042, 1993, © by the American Geophysical Union. Used with permission.

large seismogenic low-angle normal faults likely do not exist (e.g., Jackson and McKenzie, 1983; Stein et al., 1988; Buck, 1988, Jackson and White, 1989; King and Ellis, 1990).

Research related to the characterization of active normal faults for seismic hazard evaluations over the past twenty years, has resulted in a wealth of data regarding the geologic, geomorphic, structural, and seismologic characteristics of seismogenic normal faults, as summarized below.

2.4.1.1 Physiographic Features

Qualitative and quantitative analyses of the morphology of large-scale landforms associated with normal faults (i.e., range fronts; fault-generated hillslopes, including piedmont fault scarps, basal triangular facets, etc.) and associated alluvial and fluvial systems have been used to identify and assess the relative activity of normal faults and fault segments (e.g., Menges, 1988; 1990; Briais et al., 1990). Morphologic data used for these assessments can be derived from analysis of satellite images, aerial photographs, topographic maps, and fieldwork. Field data and morphometric measurements may be analyzed with a variety of data including: field- and office-based topographic profiling; range-front projections; topologic ordering of facets and ridgelines; linear and curvilinear regressions; diffusion modeling of scarp degradation, time-series analysis, and parametric and nonparametric analyses of variance (e.g., Menges, 1988).

Analysis of fault scarps from both historic and prehistoric earthquakes provides important constraints on many fault parameters (e.g., displacement per event, recurrence, segmentation) that are of concern to earthquake hazard assessments (Schwartz, 1988). McCalpin (1987, 1996) provides a summary of the use of relationships between fault scarps and geomorphic surfaces and fault scarp morphology to identify and assess the age of paleoseismic events. Conceptual models of fault scarp degradation (Wallace, 1977) and empirical relations between scarp-height, slope-angle, and age (Bucknam and Anderson, 1979) provided the basis for morphologic dating techniques that were later refined by modeling fault-scarp degradation with diffusion-equation model mathematics (e.g., Nash, 1980; Hanks et al., 1984; Mayer, 1984; Hanks and Wallace, 1985). Discussion of the applications and limitations of this technique are provided in a proceedings volume for a 1987 workshop "Directions in Paleoseismology" (Crone, 1987; Hanks and Andrews, 1987; Mayer, 1987; Pierce and Coleman, 1987; and Nash, 1987). A review of developments in the quantitative modeling of fault-scarp morphology recently has been presented by Hanks (1998).

Scarp displacement data also can be used to assess the location and seismic moment of earthquakes, to constrain seismologic and dislocation models of fault orientation, slip distribution, and moment energy release computed from geodetic data and seismic coda (Stein and Barrientos, 1985), and to infer the spatial distribution and form of potential geodetic anomalies that precede earthquakes (Scholz, 1990; Bruhn and Schultz, 1996).

2.4.1.2 Structural Features

As noted above, normal faults in extensional continental settings typically are associated with asymmetric half graben that are characterized by a master fault or fault zone that dips toward the basin floor and one or more antithetic faults (Figures 2.5 and 2.6). Bruhn and Schultz (1996) note the following with regard to antithetic and splay faults in normal systems:

(1) Antithetic faults, which may intercept the master fault at depth, apparently can nucleate either at or near the surface, due to bending of the hanging wall as it progressively is displaced downward along the master fault zone, or at depth due to either large stress gradients generated by spatial changes in slip along the master fault or changes in master fault dip (*See* Cloos, 1968, and discussions by Rosendahl, 1987, Melosh and Williams, 1989, Weissel and Karner, 1989, Schultz, 1991, and Dresden et al., 1991).

(2) Bends, steps, and jogs in the master fault presumably cause local stress concentrations that contribute to the development of splay faults of variable size and orientation (e.g., Newhouse, 1940; Stewart and Hancock, 1991).

(3) Antithetic and splay faults may have only limited strike length (few kilometers) and depth or they may extend along strike for tens of kilometers and extend deep into the seismogenic crust.

A discussion of the large-scale geometry and internal structure of extensional and transtensional fault zones, and the influence of these structures on fault rupture during large earthquakes is provided by Bruhn et al. (1990).

2.4.1.3 Stratigraphic Features

In arid to semiarid regions, such as the Basin and Range province of the western United States, individual paleoseismic events along normal faults can be identified through stratigraphic studies of deposits adjacent to the scarps of young normal faults in trench and natural fault exposures (e.g., Malde, 1971; Swan et al., 1980; McCalpin, 1982; Schwartz and Coppersmith, 1984; Machette, 1988; Forman et al., 1991; Machètte et al., 1991, 1992; Kelson et al., 1996b; McCalpin, 1996). Paleoseismologic investigations of normal faults generally involve the assessment of geomorphic and structural-stratigraphic relationships exposed in natural exposures or trenches. Many of the early detailed paleoseismic studies of normal faults were conducted along the Wasatch fault zone, and from these studies stratigraphic and structural relationships and criteria (i.e., colluvial wedges) were developed to identify geologic evidence for the size and timing of large-magnitude paleoearthquakes on normal faults (e.g., Swan et al., 1980; Schwartz and Coppersmith, 1984; Schwartz, 1988). Subsequent studies have developed more detailed facies models of colluvial sedimentation adjacent to normal-fault scarps (Nelson, 1987; 1992) (Figure 2.7) and have discussed conditions (both tectonic and sedimentary) that complicate the interpretation of the colluvial-wedge geometry and stratigraphy (McCalpin, 1987).

2.4.1.4 Seismologic Features

As noted above, seismologic data from extensional regimes both in the western United States and worldwide have been used to assess the downdip geometry of normal faults (e.g., Arabasz and Julander, 1986; Jackson, 1987; Jackson and White, 1989; Doser and Smith, 1989). Using seismologic data (first-motion analyses, body and surface waveform modeling) combined with geodetic and geologic studies of surface faulting, Doser and Smith (1989) examined the source parameters of earthquakes of magnitude 5.5 to 7.8 that occurred in the western Cordillera of the United States between 1915 and 1988. These events consisted of approximately equal numbers of normal, strike-slip and oblique-slip faulting. The principal results of their analyses were: (1) all earthquakes occurred on faults dipping 38° or more; (2) all $M_W \geq 7.0$ earthquakes occurred at depths 12 km and were composed of multiple subevents, (3) most earthquakes (> 70 percent) had unilateral ruptures; and (4) no individual subevent had a rupture length > 21 km.

Active normal faults commonly do not produce prominent alignments of microseismicity. The Wasatch fault zone in Utah, for example, with an average Holocene slip rate of > 1 mm/yr has little associated microseismicity (Arabasz et al., 1992).

2.4.2 Reverse Faults

Reverse and thrust faults are dip-slip faults in which the hanging wall block has moved up relative to the footwall block (*See* Section 2.1, above). In general, both of these types of faults place older rocks or sediments over younger rocks or sediments, resulting in a repetition of stratigraphy observed in cross section across the fault. For imbricate reverse faults, stratigraphic repetition also can be observed in map view. Reverse and thrust faults both accommodate contraction; the distinction being that reverse faults have dips greater than 45°, whereas thrust

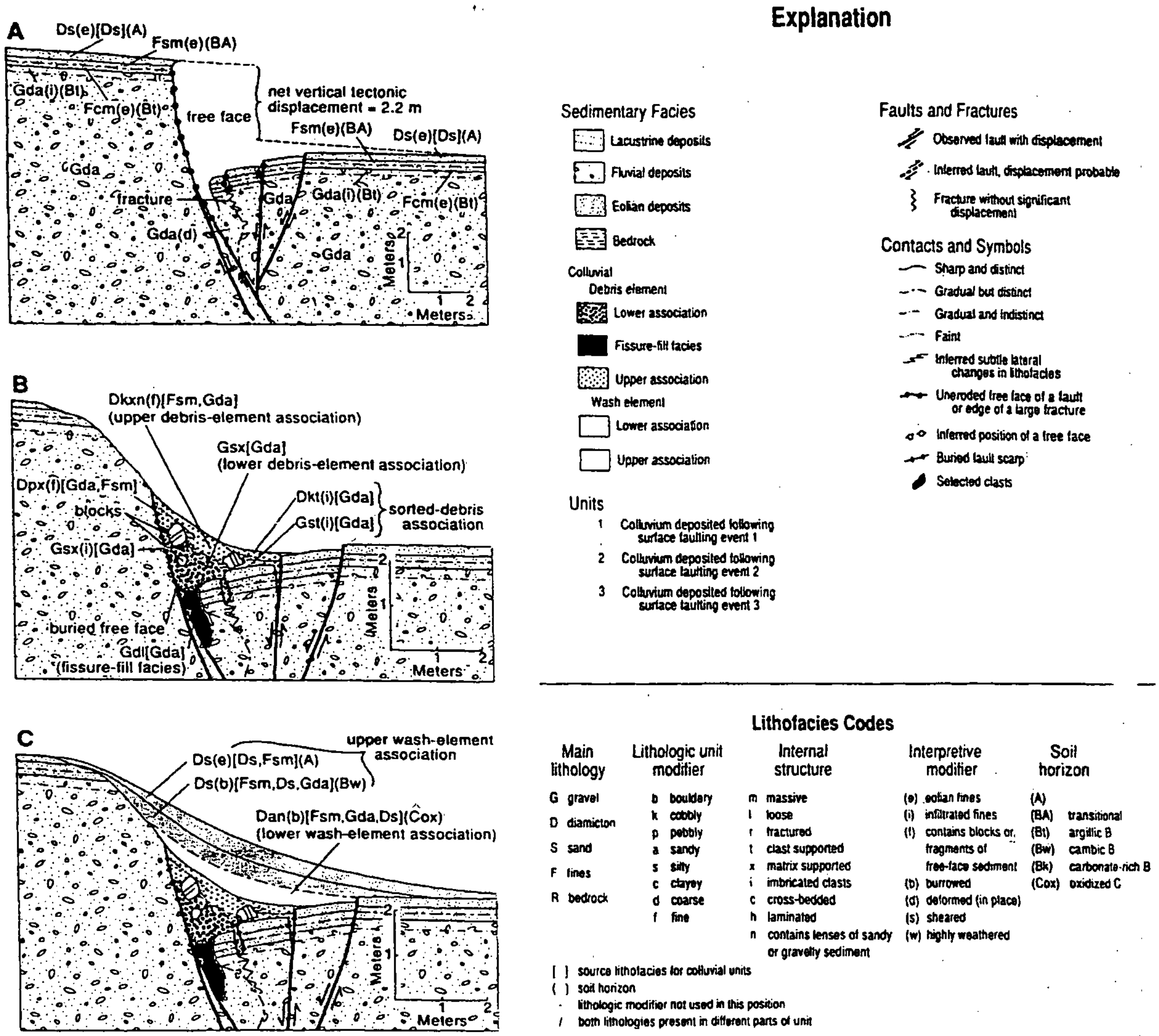


Figure 2.7 Sequential diagrams showing the deposition of colluvial lithofacies adjacent to a fault scarp on gravelly alluvium following a 2.2-m-displacement event on a Basin and Range normal fault - modified from Nelson, A.R., A Facies Model of Colluvial Sedimentation Adjacent to a Single-Event Normal-Fault Scarp, Basin and Range Province, Western United States, (Crone, A.J., and E.M. Omdahl, eds.), Directions in Paleoseismology, USGS Open File Report, 87-673:136-145, 1987. Colluvial lithofacies are grouped into debris and wash facies architectural elements, and upper and lower facies associations are recognized within each element. Only major facies relations are shown in cross sections; most exposures of faults are more complex than this model. (A) Scarp on fluvial and eolian deposits (with a soil developed on them) immediately after faulting, before deposition of colluvium. (B) Facies relations after deposition of the debris archtectural element. (C) Facies relations after deposition of the wash architectural element (present-day conditions). Modified from Nelson, 1987.

faults have dips less than 45°. The extent and amount of displacement along reverse and thrust faults are at all scales (Twiss and Moores, 1992). Displacements may occur on the order of millimeters and meters at individual outcrops or exposures, to tens or hundreds of kilometers at the scale of mountain ranges. Complex zones of thrust faults along convergent plate margins have accommodated thousands of kilometers of displacement. All of these scales are pertinent to the assessment of seismic hazards. The larger faults and fault systems may produce moderate to great earthquakes or may be reactivated as normal faults in extensional environments (e.g., Arabasz and Julander, 1986; West, 1992, 1993), and the smaller, outcrop-scale features may yield information on the style, timing, and displacements of individual paleoearthquakes (e.g., Kelson et al., 1996a). Well-known examples of active reverse and thrust faults in the United States include the San Fernando, Coalinga and Northridge faults in southern California (Proctor et al., 1972; Oakesshott, 1975; Teng and Aki, 1996), the Cascadia subduction zone in the Pacific Northwest (Atwater, 1987; Clarke and Carver, 1992; Nelson et al., 1996), and the Reelfoot fault in the New Madrid seismic zone (Russ, 1982; Chiu et al., 1992; Kelson et al., 1996a). Many active thrust faults have little or no prominent surficial expression (i.e., "blind" thrusts); these structures are treated in more detail in Section 2.4.4 of this report.

Reverse and thrust faults are recognized through identification of features common to all faults, including the presence of distinct physiographic features, truncated geologic features, large- and small-scale stratigraphic relations, historical seismicity, and geodetic strain. However, the crustal shortening accommodated by reverse and thrust faults commonly produces many unique features, regardless of whether the displacement is related to tectonic or nontectonic processes. The following sections provide general characteristics of reverse and thrust faults based on physiography, structural features, stratigraphic relations, and contemporary seismicity.

2.4.2.1 Physiographic Features

In general, historic reverse and thrust earthquakes have resulted in complex patterns of surficial deformation. Large-magnitude earthquakes associated with convergent plate boundary subduction zones commonly produce large areas of subsidence and uplift without surface rupture (Plafker, 1969; Atwater, 1987). Nelson et al. (1996) note that during great earthquakes along subduction zones, regional coseismic subsidence up to about 2.5 m may occur over as much as 1000 km of coastline, and local coseismic subsidence of up to about 2 m may occur over as much as 10 km of coastline. Local areas in the zone of coseismic uplift can be thrust upward as much as 15 m during slip on imbricate thrusts or from growth of folds within the upper plate (Plafker, 1969; Nelson et al., 1996). Subhorizontal datums such as marine and lacustrine shorelines, fluvial and alluvial-fan surface, and cultural features can reveal the patterns of surface deformation (Hull, 1987; Hanson et al., 1992, 1994).

Physiographic features resulting from shallow crustal faults and folds related to the uplift of mountain ranges may also have complex patterns. The 1952 Arvin-Techachapi, California (Buwalda and St. Amand, 1955), the 1980 El Asnam, Algeria (Philip and Meghraoui, 1983), and the 1988 Spitak, Armenia earthquakes (Philip et al., 1992) produced very complex range-front surface ruptures that have been well documented. Complex reverse-fault ground rupture patterns are not restricted to range-front localities and mountainous terrain. For example, the 1968 Meckering (Australia) earthquake produced an array of surficial features, including prominent topographic scarps, linear ridges, and folds in near-surface materials (Gordon, 1971; Hull, 1987). The form of these features depended on the thickness and moisture content of the near-surface materials, and the pre-existing slope (Gordon, 1971). The 1987 Armenia earthquakes also produced numerous types of scarps and folds whose variability appears to be related to the type and thickness of near-surface materials, and the presence or absence of consolidated bedrock (Phillip et al., 1992). Uplift during the 1811-12 earthquakes in the New Madrid seismic zone resulted in the formation of the Reelfoot scarp and the ponding of Reelfoot Lake in northwestern Tennessee (Fuller, 1912; Russ, 1982; VanArsdale et al., 1991). Russ (1982) also showed that the Lake County uplift, which has about 8 m of Holocene structural relief, is spatially associated with a westward-dipping reverse fault that likely was responsible for at least one of the three great New Madrid earthquakes. Geomorphic analyses of deformation (i.e., analyses of stream profiles, topographic residual maps) in the vicinity of the 1987 Whittier Narrows earthquake (southern California) suggest long-term localized uplift and the presence of multiple thrust sheets separated by tear faults (Bullard and Lettis, 1993). Hitchcock et al. (1994) and Hitchcock and Kelson (1996) also used stream gradients and the pattern of alluvial-fan apices to infer the presence of multiple thrust faults in the San Francisco Bay and Northridge regions respectively.

In many cases, the topographic expression of surface rupture along a reverse or thrust fault may be quickly destroyed. Hull (1987) notes that the presence of a shallow-dipping fault plane near the base of a steep-sided mountain range makes the recognition of reverse and thrust fault traces difficult because of active slope processes, stream degradation, and alluvial-fan deposition. As an example, Hull (1987) shows that surface deformation produced by the 1931 Hawke's Bay (New Zealand) earthquake was barely perceptible only 53 years later and, therefore, the lack of recognition of an active fault prior to 1931 in this area is not surprising. This is especially relevant in urbanized areas, where cultural modifications may destroy critical but subtle geomorphic evidence of thrust or reverse faulting. In the case of the 1980 El Asnam, Algeria earthquake, the physiographic expression of coseismic "secondary" structures (i.e., folding and minor normal faulting in the hangingwall) was much more prominent than the surface expression of primary reverse fault rupture, even though the displacements were 5-10 times larger on the primary fault. Whereas topographic growth of the hangingwall fold produced dramatic observable results (including damming and flooding of the Ech Cheliff River) the subtle physiographic expression of the shallow dipping primary reverse fault that emerged in a freshly plowed field went unnoticed for several days (Phillip and Meghraoui, 1983; F. H. Swan, personal communication).

2.4.2.2 Structural Features

Reverse and thrust faults characteristically emplace older (deeper) rocks and sediment on top of younger (shallower) rocks and sediment, as demonstrated in vertical sections across the fault (Figure 2.1). Horizontal separation across reverse or thrust faults will vary along strike of individual faults, depending on the attitude of displaced strata, the net direction of fault slip and the dip of the fault (Figure 2.8) (Twiss and Moores, 1992). Either left-lateral or right-lateral strike separation may be produced by pure reverse dip slip, depending on the orientation of the bedding relative to the slip direction. As with strike-slip faults, the only definitive method to assess the net amount and direction of slip along a reverse or thrust fault is to identify and characterize one or more piercing line(s) across the fault. Unfortunately, such linear markers across reverse faults are difficult to detect in the geologic record, primarily because reverse displacement will result in erosion of the hangingwall and burial of the footwall in the near-surface environment. Carver (1987) notes that localized folding or displacement of initially horizontal strata across reverse and

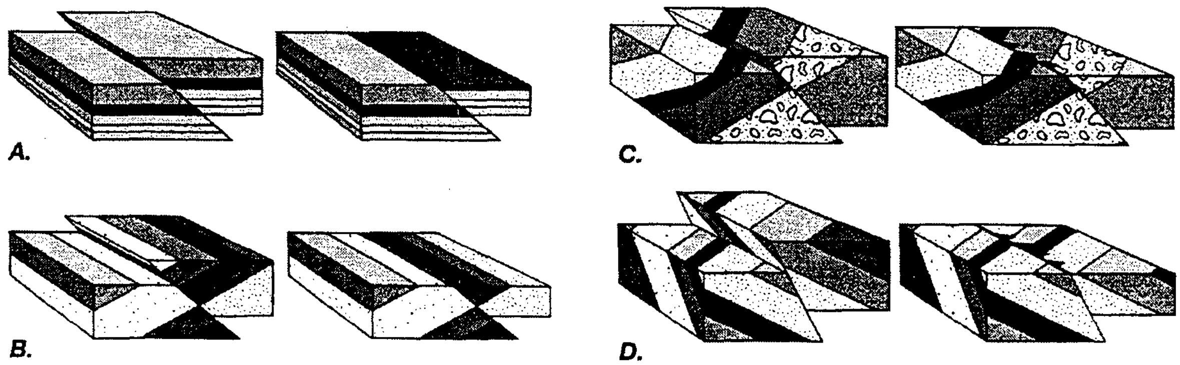


Figure 2.8 The effect of the dip of strata on the separations developed as a result of thrust faulting. The right diagram of each pair shows the hanging wall block eroded down to the same level as the footwall block. On the top surfaces (map views), A. shows a simple discontinuity, B. shows the cutting out of strata, and C. and D. show left and right lateral separations, respectively. faults From: STRUCTURAL GEOLOGY by Twiss and Moores © 1992 by W.H. Freeman and Company. Used with permission.

thrust faults provides the most obvious and easily identifiable evidence of contractional deformation. Thrust faults commonly have several structural characteristics that allow for their recognition. For example, the map trace of a thrust fault that reaches the surface usually is highly sinuous because of the intersection of the shallow-dipping fault plane with topographic irregularities (e.g., Lettis and Hall, 1994). At depth, thrust faults generally are listric faults that curve toward shallow or horizontal dips, although some faults may continue through the crust at a shallow dip (Chiu et al., 1992). Thrust and reverse faults associated with contractional strike-slip faults (i.e., positive flower structures) often have very shallow dips in the near surface that steepen with depth to join the main, strike-slip fault (e.g., Sylvester and Smith, 1976; Woodcock and Fisher, 1986). Some thrust faults are associated with a **thrust sheet**, which is a hanging wall block with a large areal extent relative to its thickness. A thrust sheet that has been moved a great distance and thus is geologically out of place is called **allochthonous** (e.g., the Pemine Alps of western Europe and the inner Piedmont of southwestern North America), in contrast to **autochthonous** folding and faulting of a body of rock that has formed largely *in situ* (e.g., are Rocky Mountain foreland of western-central North America) (Twiss and Moores, 1992). The Hartz Mountain thrust in Wyoming underlies an allochthonous thrust sheet that likely is a result of large-scale gravity-induced movement (*See* Cotton, Appendix A of this report).

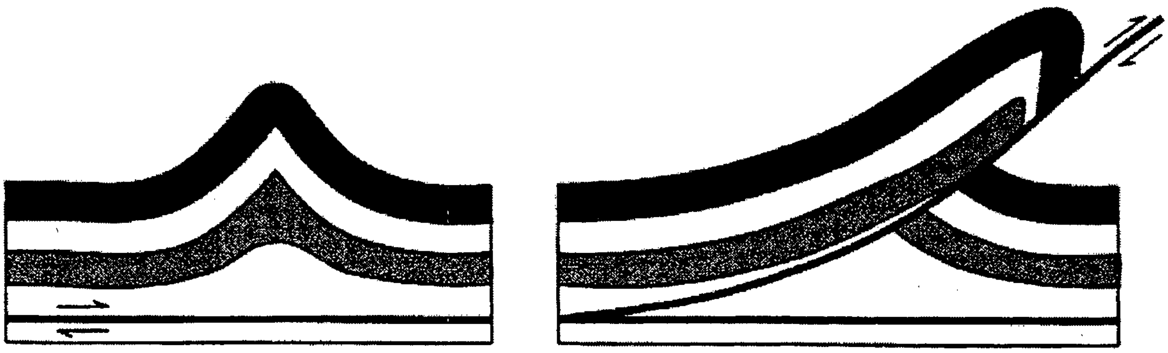
Thrust faults typically do not consist of a smooth, simple surface, but instead are characterized by flat sections of the fault that may parallel stratigraphy as well as **fault ramps**, in which the fault surface cuts upsection across stratigraphy. This **ramp-flat geometry** may be restricted to moderate to shallow crustal levels where the fault cuts through rocks of substantially different strength. The occurrence of ramps and flats near the ground surface at the scale of surficial deposits relevant to assessing paleoseismic characteristics is unlikely because of the low strengths and unconsolidated nature of surficial deposits. Thrust faults commonly also have **lateral ramps**, which represent the margins of a thrust sheet that is (sub)parallel to the direction of transport. Lateral ramps also may accommodate the along-strike linkage of thrust movement at different stratigraphic horizons. Twiss and Moores (1992) note that if the lateral ramps (also known as **sidewall faults**) are steeply dipping, they typically exhibit strike-slip displacement and are termed **tear faults** or **transfer faults**. These faults accommodate differential movement within a thrust sheet, or connect different sections of the thrust fault system.

Many thrust faults are complex systems that include numerous fault strands, commonly underlain by a detachment thrust or **décollement**. Deformation associated with the thrust is confined to the rocks above the décollement, with individual listric thrust planes joining the décollement in the subsurface. Many thrust systems include an **imbricate fan**, in which several listric thrust faults overlap to form a series of thrust sheets (Butler, 1982). In imbricate thrust systems, the thrusts branch up from the décollement in the direction of relative movement of the thrust sheet. Thrust systems may also include a **thrust duplex**, in which the individual faults that branch from the basal detachment also curve upward to merge along an upper, **roof thrust**. Duplexes differ from imbricate fan systems because they are confined within the stratigraphic section.

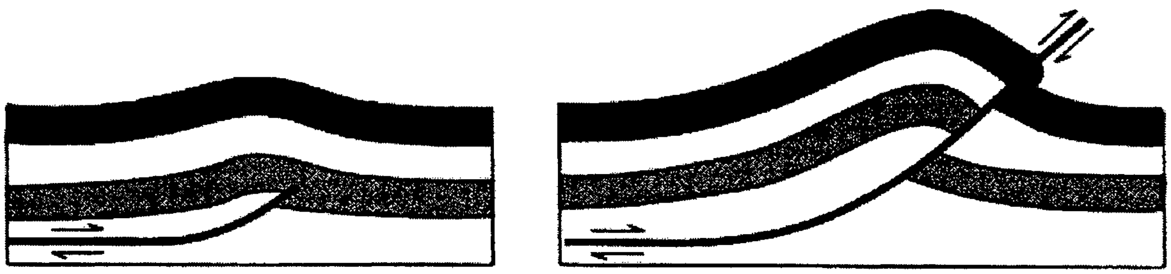
Thrust faults commonly are associated with folds, such that the term **fold and thrust belt** is common in the geologic literature. This deformation may include folds in surficial deposits overlying active thrust faults. Thrust faulting also may produce **fault-propagation folds**, which accommodate deformation near the tip line of a thrust (e.g., Mitra, 1990). Ramps in the fault surface require that the thrust sheet deforms as it moves, with movement over the ramp resulting in a **fault-ramp fold** or **fault-bend fold** (Suppe, 1983). The recognition of these fold types may provide solid evidence for the existence of adjacent reverse or thrust faults. In contrast, **lift-off** or **detachment** folds form by flow of ductile material such as salt or overpressured shale into the fold core. Detachment folds, therefore, do not require the presence of an underlying fault ramp (e.g., Mitra and Namson, 1989), although, thrust ramps may form subsequent to fold development. At all scales of contraction, folds typically are asymmetric and their development may continue until the limbs of a fold cannot be rotated any closer together. This results in the development of a thrust fault that generally cuts the steep or overturned limb of the fold (Figure 2.9). In some cases, the steep or inverted limb of the fold becomes progressively sheared and thinned, until the limb itself becomes a zone of thrust faulting called an **out-of-syncline** thrust. Similar late-stage thrust faults effect the limbs of detachment folds (**break-thrusts**) and fault-bend folds (**break-through faults**).

2.4.2.3 Stratigraphic Features

There are several types of stratigraphic contrasts that may indicate the presence of reverse or thrust faults. In addition to stratigraphic inversion or repetition that results simply from older rocks placed over younger, there is the inverse juxtaposition of metamorphic rocks of substantially different grade, whereby high-grade rocks, which are generally associated with deep structural levels, overlie unmetamorphosed or low-grade metamorphic rocks. Thrust faults also may juxtapose stratigraphic sequences of the same age but different sedimentary facies. For example, juxtaposition of shales, cherts, or other rocks formed in deep-water environments with shallow-water limestones or sandstones suggests that the contact between the two sequences is a thrust fault with large displacement (Twiss and Moores, 1992). The presence of highly deformed rocks overlying relatively undeformed rocks of the same stratigraphic sequence probably indicates the presence of thrust fault between the two



A.



B.

C.

Figure 2.9 Diagrammatic cross sections illustrating relationships between folds and thrust faults. A. Thrust fault cuts up from the décollement through the foreland limb of a fold when the fold becomes too tight to accommodate further shortening. B. Fold forms in association with the propagation of a thrust fault. C. Formation of a fold by ductile flow can result in the shearing out of one limb to form a ductile thrust fault. From: STRUCTURAL GEOLOGY by Twiss and Moores © 1992 by W.H. Freeman and Company. Used with permission.

sequences. Where thrusting reactivates older normal faults (**structural inversion**) normal stratigraphic may persist, but anticlinal deformation may be recognizable.

In addition to faulted strata there are recognizable stratigraphic features associated with Quaternary reverse or thrust faults related to vertical offset of the ground surface (Carver, 1987). These features are related to changes in the geomorphic or sedimentary processes triggered by coseismic deformation. A considerable amount of literature has been produced in the past decade concerning stratigraphic evidence of great earthquakes related to subduction in the Pacific Northwest (*See* Nelson et al., 1996 for references). Criteria identified by Nelson et al. (1996) that allow identification of rapid coastal subsidence during great subduction-zone earthquakes address: (1) the suddenness of submergence, (2) the amount of submergence, (3) the lateral extent of peat-mud contacts, (4) evidence of tsunami-related stratigraphy, and (5) the synchroneity of submergence over large sections of the coast. Nelson et al. (1996) note that deciphering a record of great earthquakes on the Cascadia subduction zone (and therefore other subduction zones worldwide) will require detailed studies of intertidal stratigraphy, sedimentology and paleoecology, and application of improved dating techniques.

Earthquakes produced by shallow crustal reverse and thrust faults also produce stratigraphic features that can be investigated during paleoseismic studies. Because the topographic changes may be on the order of less than a meter or so, paleoseismic indicators of individual earthquakes usually requires detailed stratigraphic or geomorphic analyses (Carver, 1987). The rapid development of fault scarps during large earthquakes commonly yields colluvial deposits shed onto the downthrown block, which may provide temporal information on fault movement if they contain datable material or are associated with pedogenic soil development. As noted above, the style of surface deformation along the fault trace may vary considerably, which results in different types and locations of fault-related surficial deposits. Carver (1987) shows that colluvial stratigraphy on the downthrown side of reverse faults can provide detailed information on the timing and style of paleoseismic deformation. Swan (1987) also documents displaced, scarp-derived colluvial deposits on the downthrown side of the thrust fault that ruptured during the 1980 El Asnam (Algeria) earthquake. Using a slightly different approach, Kelson et al. (1996a) excavated across a small graben in the hanging wall of the Reelfoot fault in the New Madrid seismic zone, and interpret the occurrence of paleoearthquakes on the basis of colluvial strata preserved within the graben. This study suggests that many of the different scarp morphologies produced by reverse or thrusting noted by Gordon (1971) and Phillip et al. (1992) may yield paleoseismic information. In addition, King and Vita-Finzi (1981) show that drainage ponding and lacustrine deposition may also yield information on paleoearthquake deformation.

Carver (1987) and Hull (1987) both note that folding may be the dominant mode of surface deformation during reverse earthquakes. King and Stein (1983) show that deformed stream profiles and fluvial-terrace profiles may be coincident with areas of localized coseismic deformation. Kelson et al. (1992, 1996a) show that folding during paleoearthquakes along the Reelfoot fault in the New Madrid seismic zone resulted in the deposition of colluvium that can be used to assess earthquake timing.

Unfortunately, there are wide variations in the amount of displacement along the strike of a reverse or thrust fault during individual earthquakes (Hull, 1987), suggesting that products of scarp degradation may vary considerably along strike. This characteristic of reverse faulting suggests that empirical relations between fault displacement and earthquake magnitude may have greater uncertainty than generally believed, and that the segmentation of reverse faults based on amounts of slip per event along strike may be problematic.

2.4.2.4 Seismologic Features

Contemporary seismicity, where present, may provide excellent evidence of active reverse and thrustfaulting. However, seismicity can only provide evidence for the presence of an active tectonic source; the absence of contemporary seismicity does not prove the absence of an active source. There are at least three main features of contemporary seismicity that are strongly suggestive of reverse faulting: (1) diffuse alignment of epicenters in map view, (2) reverse focal mechanisms, and (3) the alignment of hypocenters that define a dipping fault plane in cross section. The association of contemporary seismicity with active faulting, rather than a nontectonic process (e.g., man-made explosions, reservoir-triggered seismicity, hydrologic loading) may be indicated by the temporal and spatial pattern of events.

Focal mechanisms provide additional information to evaluate style and orientation of fault-slip. A reverse earthquake focal mechanism can be used for identifying causative seismogenic structures by allowing interpretations of two directions of maximum compression, and the orientation of two possible fault planes (Figure 2.4). Further evaluation of the seismicity pattern is necessary to correlate seismic evidence with the type and attitude of the fault rupture (Haucksson, 1994).

Reverse faults typically produce a diffuse seismicity pattern in map view because of their low-angle fault planes (Chiu et al., 1992). Identifying a causative fault or active seismogenic structure from the map pattern of contemporary seismicity requires review of a cross section oriented perpendicular to the strike of the fault showing earthquake hypocenters. In cross section, hypocenters typically form a dipping alignment representative of the inferred fault plane (Hauksson, 1994). If the orientation of the cross-section is not perpendicular to the fault plane, the alignment of hypocenters shows an apparent dip of the fault plane, and the distribution of hypocenters may falsely suggest a wider and/or shallower-dipping fault plane (Chiu et al., 1992). For reverse earthquakes without surface rupture, seismicity cross-sections suggest where surface deformation may occur by projecting the dipping fault plane to the ground surface. For instance, the causative structures for many of the recent moderate-sized earthquakes in the Los Angeles region, such as the 1994 Northridge earthquake, were delineated by plotting the earthquake hypocenters in cross-section.

2.4.3 Strike-Slip Faults

A strike-slip fault is a fault along which most of the movement is parallel to the fault strike (Bates and Jackson, 1987). As described in Section 2.1, strike-slip faults are either right-lateral (dextral) or left-lateral (sinistral) depending on the relative motion of the block on the opposite side of the fault. Sylvester (1988) subdivides strike-slip faults into two broad classes: interplate **transform** faults, which are regional, plate-bounding strike-slip faults that extend through the lithosphere and accommodate large amounts of relative plate motions; and intraplate **transcurrent** faults, which do not cut through the lithosphere and juxtapose pieces of the continental crust.

Sylvester (1988) defines three types of transform strike-slip faults:

(1) **ridge** transform faults, which are confined to the oceanic crust and link the offset segments of oceanic ridge spreading centers (e.g., Kane fracture zone, North Atlantic)

(2) **boundary** transform faults, which separate unlike plates and generally are more parallel to the plate boundary (e.g., the San Andreas fault, California)

(3) **trench-linked** transform faults, which accommodate the horizontal component of oblique subduction at convergent plate margins and typically are parallel to the trench and associated magmatic arc (e.g., the Median Tectonic Line, Japan)

Sylvester (1988) further defines four types of transcurrent strike-slip faults:

(1) **indent-linked** strike-slip faults, which separate continental blocks that move with respect to one another in convergent plate settings (e.g., the North Anatolian fault, Turkey)

(2) **tear** faults, which accommodate displacement between larger structural features, such as regional strike-slip, reverse, or normal faults (e.g., the Kalabagh fault, Pakistan)

(3) **transfer** faults, which link en echelon segments of major strike-slip faults across step-over zones (e.g., the Kickapoo fault between the Johnson Valley and Homestead Valley faults, California)

(4) **intracontinental transform** faults, which juxtapose allochthons of different tectonic origins (e.g., the Garlock fault, California)

Because of the large displacements and regional nature of transform strike-slip faults (as defined above), most of the uncertainty in differentiating between tectonic and nontectonic strike-slip faults are associated with the transcurrent style of strike-slip faults. For this reason, we focus below on the recognition of transcurrent strike-slip faults, although many of the characteristics of strike-slip faults in general were identified via study of onshore and near-shore transform strike-slip faults such as the San Andreas (California), El Pilar (Venezuela), and Chugach-Fairweather-Queen Charlotte (Alaska and British Columbia) faults.

In many respects, strike-slip faults are recognized through identification of features common to all faults, including the presence of distinct physiographic features, truncated geologic features, large- and small-scale stratigraphic relations, historical seismicity, and geodetic strain. However, because the primary direction of displacement along strike-slip faults, by definition, is parallel to fault strike, many features are unique to strike-slip faults, regardless of whether the fault origin is related to tectonic or nontectonic processes. The following sections provide general characteristics of strike-slip faults based on physiography, structural features, stratigraphic relations, and contemporary seismicity.

2.4.3.1 Physiographic Features

Perhaps some of the most useful criteria for recognizing strike-slip faults are related to the topographic expression that results from surface displacement and differential erosion of displaced rock types or geologic features (e.g., Vedder and Wallace, 1970). These features are manifested in the present-day topography if the rate of surface deformation is sufficient to overcome the rate of degradation of the land surface, which may vary greatly depending on climate, geologic materials (e.g., rock types, surficial deposits, degree of soil development), topographic relief, and other factors. Thus, given appropriate conditions under which surface deformation is preserved, geomorphic features may provide excellent evidence of strike-slip faulting. Sylvester (1988) notes that the most distinctive characteristics of active or recently active strike-slip faults is their structural and topographic linearity, even in heavily vegetated areas with presumably high degradation rates. Strike-slip faults commonly are associated with **linear "rift" valleys** (Willis, 1937; Biq, 1959; Allen, 1965) that may be up to 10 km wide (Gilbert, 1907; Lawson et al., 1908; Noble, 1927; Davis, 1927). These valleys typically are present along strike-slip faults where the amount and rate of deformation are sufficient to dominate geomorphic processes (as along major faults such as those within the San Andreas fault system or the Bocono [Venezuela] fault). In contrast, some strike-slip faults with a presumably lesser influence on late Quaternary surficial processes, such as the Meers fault in southern Oklahoma, transect valleys and ridges and are not associated with a well-developed "rift" valley (Kelson and Swan, 1990; Swan et al., 1993).

Other distinctive landforms along active or recently active strike-slip faults include **pressure ridges**, closed depressions ("**sag ponds**"), and **side-hill benches**, which form as a result of localized uplift, subsidence, and/or translation of small blocks within the fault zone (*See* Figure 2.2). These features may be transient through time because of variations in the locations, amounts, and senses of vertical separation during a series of surface-rupture earthquakes. Other geomorphic features associated with strike-slip faulting include **shutter ridges**, which are ridges that have been displaced laterally so that they block drainages, and **deflected stream channels**, which have abrupt changes in orientation at a fault trace as a result of lateral offset (Wallace, 1976). Stream channels may have deflections to the left or right, although the sense of deflection may not represent the true sense of lateral fault offset because of stream capture, fault-line erosion , and other geomorphic processes.

Fault scarps are common along strike-slip faults. These scarps result from a component of vertical displacement along the fault and/or the lateral juxtaposition of topography. Both the height and orientation of the scarp typically vary along a strike-slip fault because of variation in the sense of throw along strike (i.e., "scissors" geometry) and irregularity in topography. Changes in the sense of vertical displacement along strike may be one of the most diagnostic geomorphic indicators of strike-slip faulting. Scarp heights also may vary along a strike-slip fault because of the juxtaposition of different elements within a rugged topography. For example, oblique-lateral displacement of rolling topography along the Meers fault in southern Oklahoma produced accentuated fault scarps along the southeastern sides of ridges, but diminished scarps on the northwestern sides of ridges (Ramelli, 1988; Kelson et al., 1990; Swan et al., 1993). As with dip-slip, faults strike-slip faults also may produce **fault-line scarps**, where blocks of more resistant rocks are translated adjacent to rocks of lesser erosional resistance, commonly resulting in prominent, linear scarps from erosion of the weaker rocks. Strike-slip fault-line scarps are especially distinctive due to their highly linear and therefore easily recognizable nature. It is important to differentiate between fault-line scarps and fault scarps in assessments of fault activity and sense of slip, because fault-line scarps are the result of erosion rather than recent displacement (*See* Section 2.7).

2.4.3.2 Structural Features

In general, structural features related to strike-slip faulting are comparable to those produced by all types of faults, although several structural characteristics are diagnostic of strike-slip faulting. Most strike-slip faults are approximately planar and vertical, and thus, as noted above, their map traces tend to be linear. The truncation of geologic features such as stratigraphy, foliation, dikes, sills, folds, various landforms, and other faults all indicate displacement along a fault. Demonstration of the absence of these features is perhaps the only means to show an absence of faulting. Direct evidence of strike-slip displacement of structural features is best documented by delineating the piercing points of a geologic line (Sylvester, 1988), or more succinctly, a **piercing line**. Such a line is identified on both sides of a fault and used to evaluate the amount and sense of lateral offset. In theory, piercing lines also may yield a net slip vector for all types of faults, but the identification and delineation of piercing lines across dip-slip faults is generally more difficult because of post-faulting erosion or deposition. Piercing lines previously used to document strike-slip faulting include marine strand lines and shelf-to-basin transition zones (Addicott, 1968); sedimentary and metamorphic rock facies boundaries (Roddick, 1967); formation pinch-outs and isopachous lines (Stewart, 1983); and intersections of surfaces of unique isotopic and geochemical trends (Silver and Mattinson, 1986). Late Quaternary piercing lines recently used to assess active strike-slip faulting include small channels or rills (e.g., Sieh and Jahns, 1984; Niemi and Hall, 1992; Lienkaemper and Borchardt, 1996), the back-edges of marine terraces (Weber and Cotton, 1981; Hanson and Lettis, 1994), a paleovalley margin (Kelson et al., 1996b), and the crests of glacial moraines (Kelson et al., 1996c). Other direct evidence of lateral fault slip includes the presence of subhorizontal slickenlines along a fault plane, or other small-scale structural features such as mullions, striae, and ridge-in-groove lineations (e.g., Petit, 1987; Hancock, 1985; Twiss and Moores, 1992).

Indirect evidence of lateral faulting includes the *en echelon* pattern of faults, fractures, and folds commonly present in narrow, elongate zones along strike-slip faults (Figure 2.10). A variety of shear fractures, folds, normal faults, and reverse faults commonly are associated with strike-slip faults, with the pattern depending on the sense of shear and the strength of faulted material (*See* Sylvester, 1988). Subsidiary shear fractures, or **Reidel shears**, develop at a small angle (about 10° to 20°) to the main fault in an *en echelon* array. Folds and thrust faults also form in an *en echelon* pattern above or beside strike-slip faults, with the trend of the fold hinges and the strike of the thrust faults oriented about 30° to 45° to the main fault. Where strike-slip faults contain bends or stepovers, lateral slip produces either uplift within a **restraining bend** or **restraining stepover**, or subsidence within a **releasing bend** or **releasing stepover** (e.g., Crowell, 1974; Barka and Kadinsky-Cade, 1989; Kadinsky-Cade and Barka, 1989; Brown and Simpson, 1989). A common result of these irregularities in the fault trace is a **strike-slip duplex**, which is a set of fault slivers bordered by strands of the main fault (Woodcock and Fischer, 1986; Twiss and Moores, 1992; Figures 2.11 and 2.12). These duplexes differ from those formed by dip-slip faults, which produce thickening or thinning of the continental crust via dip-slip movement. Because of their vertical orientation, strike-slip duplexes thicken or thin the crust via oblique movement. Restraining bends typically produce contraction via a series of thrust faults that diverge upward, termed a **positive flower structure** or **palm tree structure** (Sylvester, 1988). Releasing bends typically produce extension via a series of thrust faults that diverge upward, termed a **negative flower structure** or **tulip structure**.

2.4.3.3 Stratigraphic Features

Although sedimentary basins associated with strike-slip faulting are present in a variety of tectonic settings, several stratigraphic characteristics of these basins appear to be distinctive (Christie-Blick and Biddle, 1985). These are: (1) geologic mismatches within and at the boundaries of basins; (2) longitudinal and lateral basin asymmetry; (3) evidence of rapid subsidence; (4) abrupt lateral facies changes and local unconformities at basin margins as a result of pronounced topographic relief, and (5) marked differences in stratigraphic thickness, facies geometry, and the occurrence of unconformities from one basin to another in the same region (Christie-Blick and Biddle, 1985).

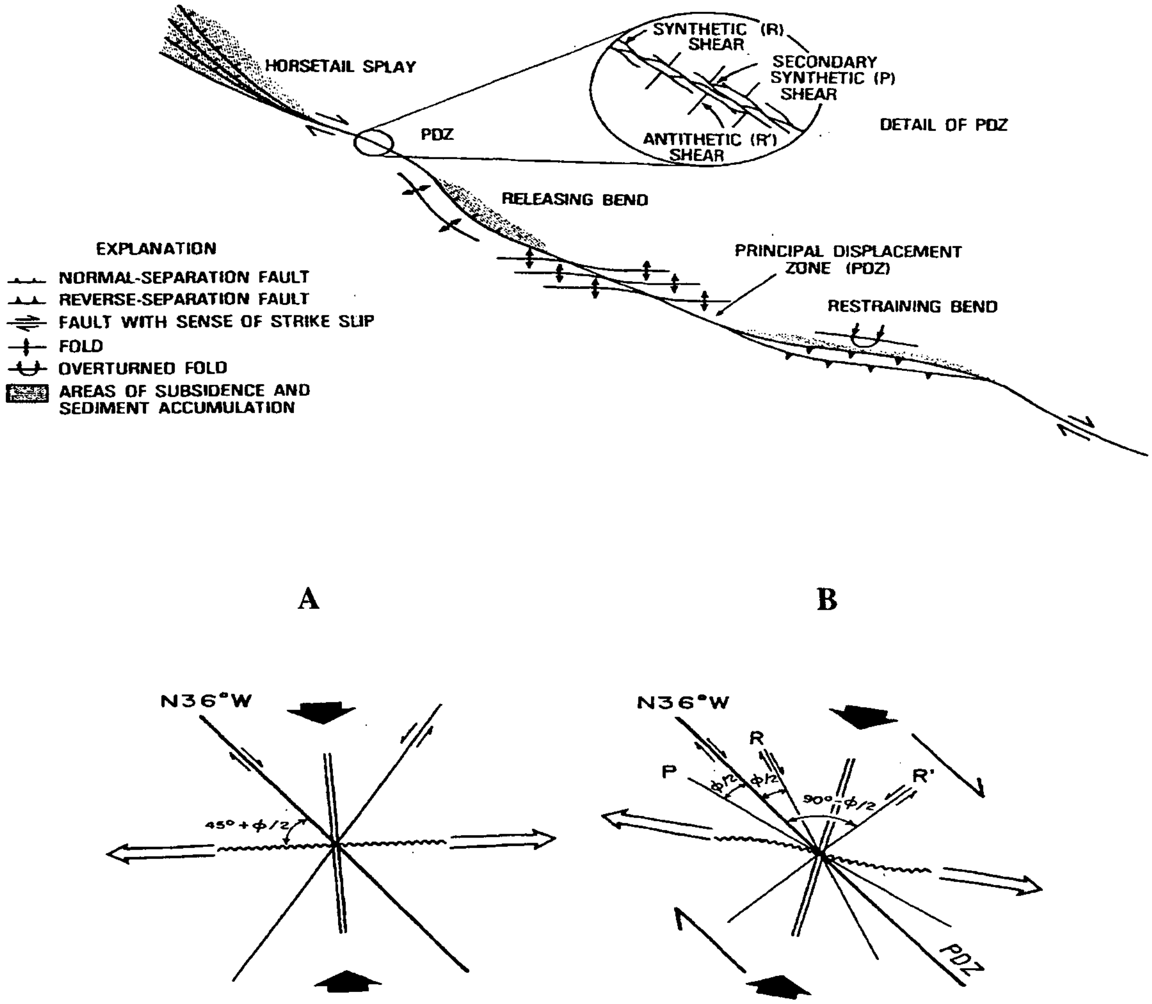


Figure 2.10 a. The spatial arrangement, in map view, of structures associated with an idealized right-slip fault. From Christie-Blick and Biddle, 1985. Reprinted with permission of the publisher, the Society of Economic Paleontologists and Mineralogists (Copyright © 1985 SEPM). b. Plan view of geometric relations among structures according to two-dimensional, strike-slip, tectonic models for a vertical fault which strikes N36 W (adapted with modifications from Aydin and Page, 1984). Reprinted with permission of the publisher, the Geological Society of America, Boulder, Colorado, USA. (Copyright © 1984 Geological Society of America). A) Coulomb-Anderson model of pure shear; B) Riedel model of right simple shear. Double parallel line represents orientation of extension (T) fractures; wavy line represents orientation of fold axes. P = P fracture, R and R' are synthetic and antithetic shears, respectively; PDZ = principal displacement zone; = angle of internal friction. Short black arrows = shortening axis; open arrows = axis of lengthening. Permission to use this copyrighted material is granted by the Society for Sedimentary Geology.

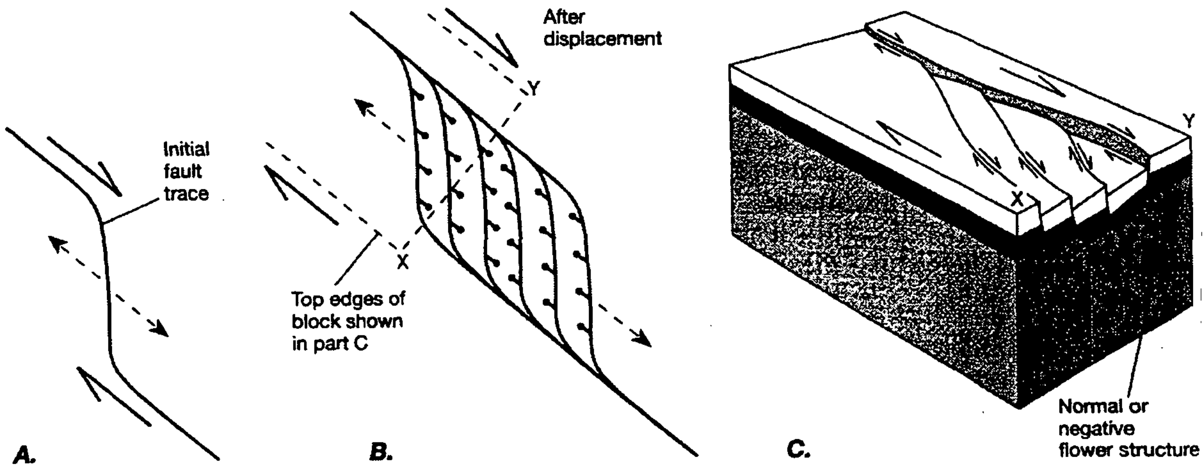


Figure 2.11 Formation of an extensional duplex at an extensional (releasing) bend. Large arrows indicate the dominant shear sense of the fault zone; small arrows indicate the sense of strike-slip and normal components of motion on the fault splays. A. Extensional bend on a dextral strike-slip fault. B. An extensional duplex developed from the bend in part A. C. A block diagram showing a normal, negative, flower structure in three dimensions. The block faces are vertical planes along the dashed lines in part B. From: STRUCTURAL GEOLOGY by Twiss and Moores © 1992 by W.H. Freeman and Company. Used with permission.

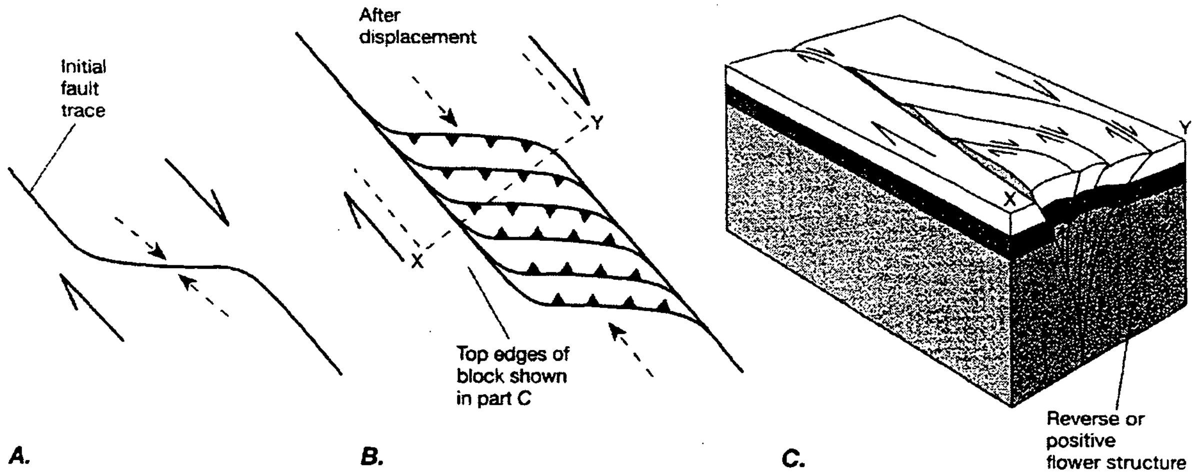


Figure 2.12 Formation of a contractional duplex at a contractional (restraining) jog. Large arrows indicate the dominant shear sense of the fault zone; small arrows indicate the sense of strike-slip and reverse components of motion on the fault splays. A. Contractional bend on a dextral fault. B. A contractional duplex developed from the bend in part A. C. A block diagram showing reverse, or positive, flower structure in three dimensions. The block faces are vertical planes along the dashed lines in part B. From: STRUCTURAL GEOLOGY by Twiss and Moores © 1992 by W.H. Freeman and Company. Used with permission.

Geologic mismatches are common in regions deformed by major strike-slip faults. For example, the Ridge basin along the San Gabriel fault in southern California contains a mismatch between sediments and suitable source rocks, which is explainable by removing 35 to 60 km of right slip on the fault (Crowell, 1982). However, the presence of similar geologic source rocks on opposite sides of a fault at a particular locality does not preclude strike-slip deformation (Christie-Blick and Biddle, 1985).

A particularly distinctive characteristic of strike-slip basins is the tendency to have longitudinal asymmetry (Christie-Blick and Biddle, 1985), perhaps as a result of intrabasin deformation and continued translation of source areas and sediment depocenters. Aydin and Nur (1982) note that typical pull-apart basins have an aspect ratio (length to width) of 3:1 in map view, although the ratio may vary considerably, depending on whether the structural, physiographic, or active dimensions of the basin are measured (Sylvester, 1988). The sense of basin asymmetry also may change from one transverse profile to another (Ben-Avraham et al., 1979). Notably, strike-slip deformation may result in basins bordered by normal faults (e.g., in a releasing stepover; Zak and Fruend, 1981), by reverse faults (e.g., in a restraining stepover; Yeats, 1983), or by a combination of both normal and reverse faulting (Nilsen and McLaughlin, 1985).

Strike-slip basins are characterized by extremely rapid rates of subsidence, and thus typically have very thick stratigraphic sections compared to lateral basin dimensions (Johnson, 1985; Nilsen and McLaughlin, 1985). However, variations in local patterns of deformation may produce episodic subsidence thereby producing highly variable rates of sedimentation. The stratigraphic record in strike-slip basins thus typically contains abrupt vertical and lateral facies variations. Facies changes may be highly localized, and the stratigraphy commonly contains laterally discontinuous unconformities.

Lastly, local tectonic controls in strike-slip basins frequently yield varied patterns of sedimentation among basins within the same region (Christie-Blick and Biddle, 1985; Johnson, 1985). Because deformation in any given basin may vary considerable in time and space, basin sediments tend to be more heterogeneous than those deposited in predominantly extensional or contractional terranes. Basins may be distinct, rhomb-shaped features (Crowell, 1974; Garfunkel, 1981) or have "lazy S" or "lazy Z" shapes (Schubert, 1980; Mann et al., 1983), and stratigraphic packages among these basins within a strike-slip region may only be crudely correlative.

2.4.3.4 Seismologic Features

Contemporary seismicity may provide excellent evidence of active strike-slip faulting. Three primary characteristics of contemporary seismicity are indicative or strongly suggestive of strike-slip faulting: (1) a linear alignment of earthquake epicenters in map view, (2) a near-vertical alignment of hypocenters in cross section, and (3) focal mechanisms that help interpret the direction of slip and the orientation of fault planes along which movement may have occurred during an earthquake. The association of contemporary seismicity with active faulting, rather than a nontectonic process (e.g., man-made explosions, reservoir-triggered seismicity, hydrologic loading) may be indicated by the temporal and spatial pattern of events.

Earthquakes produced by strike-slip faults typically form a linear pattern of epicenters, as shown in map view. The most prominent alignments of epicenters typically are associated with discrete moderate to large events and their associated aftershocks. However, numerous smaller events occurring intermittently over years or decades may produce broader, more diffuse linear zones suggestive of strike-slip faulting. In general, the inference is made that where concentrations of earthquakes locate beneath mapped faults, the seismicity defines the subsurface orientations of these faults (Walters et al., 1996). Where concentrations of earthquakes do not underlie mapped faults, it is also commonly inferred that the events represent active, but as yet unnamed or unidentified faults. Minor differences between fault traces mapped at the ground surface and seismicity commonly result from use of one-dimensional crustal velocity models to locate earthquakes in a crust where the velocity varies in three dimensions (Walters et al., 1996). Earthquake concentrations along the Hayward and Calaveras faults in the San Francisco Bay region, northern California, are perhaps some of the best examples of the association of contemporary seismicity with active strike-slip faults (Walters et al., 1996). Again, it should be noted that an absence of seismicity does not indicate that a fault is inactive (e.g., San Francisco Peninsula segment of the San Andreas fault).

Well-located hypocenters of contemporary seismicity along active strike-slip faults typically have vertical or near-vertical patterns in cross section. Slight variations from vertical orientations commonly are a result of slight mislocations of event hypocenters, which also may result from use of one-dimensional crustal velocity models in a heterogeneous crust. For example, Walters et al. (1996) provide cross sections of the Calaveras fault with a zone of hypocenters dipping about 85° NE, although detailed studies of the seismicity indicate that the fault probably is vertical.

Focal mechanisms with horizontal T- and P-axes may provide direct evidence of strike-slip faulting during individual earthquakes. For example, the Calaveras fault in northern California is associated with many strike-slip focal mechanisms that plot along its mapped trace. This fault also is associated with a few reverse focal mechanisms adjacent to its main trace (Walters et al., 1996) that probably reflect oblique thrusting as might be expected in a positive flower structure.

2.4.3.5 Geodetic Strain Characteristics

Geodetically measured strain can help identify regions that are undergoing horizontal shear and thus may contain active strike-slip faults. Geodetic arrays are used to determine the spatial distribution or contemporary rates of deformation (strain) across specific fault zones by using the average rate of change in distance between survey markers within a given area. The average rate of deformation between two survey markers is determined by the slope of linear fit to a distance-time plot for either preseismic or postseismic observations (Lisowski et al., 1991). The map pattern of strain vectors represent the velocity field of deformation. In strike-slip environments, relative velocities between points on opposite sides of major faults generally are parallel to the fault traces. Notably, geodetic deformation rates usually are lower than observed geologic rates, because: (1) geodetic networks may not span the entire zone of strain, and (2) geologic rates generally represent an average rate over several earthquake cycles, whereas geodetic rates represent a limited time window and instead reflect only a short period (Lisowski et al., 1991). Geodetically determined displacement vectors are useful for estimating regional strain, but may be difficult to correlate with specific faults even though active faults may be recognized in one area (e.g., Thatcher, 1995).

2.4.4 Blind Faults

Faults for which the primary rupture surface does not break the overlying syntectonic cover are referred to as "blind faults" (Lettis et al., 1997). This definition includes two "modes" of blind faults: those that could break the surface but are located in environments with high ratios of sedimentation to displacement, and therefore, constantly are concealed by depositional processes; and those that are blind due to their geometry and mechanical behavior. These two modes, or causes for fault blindness, may overlap in situations where propagation of a reverse fault to the surface is suppressed by an overlying sedimentary load.

Although all styles of active, primary faulting may be blind, the most widely recognized blind faults are low-angle thrust faults, which are common features of fold-and-thrust regions of contractional orogenic belts where they were first described. The following discussion, therefore, is based predominantly on the characteristics of active blind-thrust faults.

Blind-thrust faults are low-angle, reverse-slip faults that do not reach the ground surface, even though they may be active structures having kilometers of displacement. The relationship between folds and underlying faults was recognized more than 60 years ago, when a buried thrust ramp was postulated as the cause for generation of the Powell Valley anticline in the Appalachian fold and thrust belt (Rich, 1934). Quantitative techniques for representing the deep structure within fold and thrust belts began with the development of cross section balancing techniques in the oil and gas industry (Bally, 1966; Dahlstrom, 1969). In the early 1980's a quantitative theory relating faulting and folding, termed fault-bend fold theory, was developed (Suppe, 1983). Fault-bend fold theory relates the geometric and kinematic properties of near-surface folds to the location, geometry and slip of the underlying, hidden thrust fault using the axioms of cross section balancing, a flexural-slip fold mechanism, simple trigonometric relationships, and the assumption that slip over bends in a fault surface will induce folding of the overlying rock to conserve volume and prevent the development of voids or overlap between blocks. The theory provides a powerful tool for predicting the location and slip attributes of a blind-thrust fault from the geometry and kinematics of the overlying fold.

A series of recent moderate- to large-magnitude blind-thrust earthquakes that struck California (1982 New Idria, M_w5.4; 1983 Coalinga, M_w6.5; 1985 Kettleman Hills, M_w6.1; 1987 Whittier Narrows, M_w6.0; and 1994 M_w6.7 Northridge earthquakes) confirms the seismic potential of blind-thrust faults in the Coast Ranges and Transverse Ranges of western California, including the Los Angeles basin region. Blind-thrust faults were increasingly recognized through the 1980's as a class of earthquake sources that needed to be addressed (Stein and Yeats, 1989).

2.4.4.1 Physiographic Features

Unlike emergent reverse faults, blind-thrust faults remain completely buried within the Earth's crust, often to depths greater than 10 km (6 mi). Instead of producing localized, intense surface deformation in the form of scarps and mountain fronts associated with emergent faults, blind-thrust faults commonly are expressed at the surface or near-surface as broad anticlinal folds that grow progressively during earthquake events (Stein and Yeats, 1989; Ekstrom et al., 1992). Although the term "blind" is derived from the characteristic that the fault itself does not reach the surface and, therefore, cannot be seen, the presence of surface deformation in tectonically active regions can be exploited for characterizing potential seismic sources (e.g., Davis et al., 1989; Unruh et al., 1992; Bullard and Lettis, 1993; Bullard et al., 1994; Hitchcock et al., 1994; Shaw and Suppe, 1996; Kelson et al., 1996a).

One of the most important aspects of blind-thrusts in terms of their identification is that they are not truly "blind". In nearly all cases where active thrust faults can be identified on the basis of seismicity in interplate regions, their existence also is indicated by recognizable geomorphic features at the surface (Wells and Lettis, 1990; Lettis et al., 1997). These mappable features include a large range of structural and geomorphic features associated with localized uplift and tilting of the surficial deposits and underlying bedrock that lie in the hangingwall of a blind-thrust fault. The most obvious feature is an anticlinal fold affecting bedrock and overlying surficial deposits, such as the Coalinga and Kettleman Hills anticlines associated with the 1983 and 1985 blind-thrust earthquakes, respectively (Wentworth and Zoback, 1989, 1990; Ekstrom et al., 1992). Other associated features that are less obvious include anomalous morphometry of overlying fluvial systems, including basin asymmetry and longitudinal stream profiles, uplifted and preserved geomorphic surfaces such as fluvial and marine terraces, and range-front sinuosity (e.g., Bullard and Lettis, 1993; Hitchcock et al., 1994).

Recent advances in landform analysis techniques using a landscape modeling approach combined with analysis of geodetic data, digital elevation models, quantitative tectonic geomorphology, and numerical modeling of fault deformation has resulted in a promising multidisciplinary approach to characterization of blind faults (Arrowsmith and Rhodes, 1994; Burgmann et al., 1994; Arrowsmith, 1995; Arrowsmith et al., 1996). In this analysis, digital elevation models can provide morphometric parameters of the landscape such as stream gradient patterns, drainage patterns and local topographic relief. These can be related to fault parameters using numerical models of tectonic and geomorphic processes (e.g., linear elastic half-space models of fault deformation, and numerical models of landscape diffusion/scarp degradation). The resulting relationships can be used to assess changes in a synthetic, gridded landscape by altering both tectonic and geomorphic input parameters.

2.4.4.2 Structural Features

Because blind faults do not crop out, their surface expression is produced primarily by secondary effects of the deforming rock mass surrounding the primary fault. These secondary effects are principally anticlinal folding and development of minor faults and ground fractures. The anticlinal folds that form above blind thrusts fall into the three categories already described in Section 2.4.2.2 on characteristics of reverse faults: fault-bend folds; fault-propagation folds; and detachment folds. During fault-propagation and detachment folding, complex structures are formed in the highly strained region of fault tips, where slip decreases to zero and the transition from faulted to unfaulted rock is expressed as a zone of "continuous, heterogeneous shear" (i.e., fold deformation). Alternatively, anticlinal fault-bend folds are formed primarily by passive response of the hangingwall to translation over flat-and-ramp geometry of the fault surface (e.g., Suppe, 1983). Anticline back limbs are formed over ramp regions where uplift occurs, and firelimbs are formed over the ramp-to-flat transition where the rock mass no longer experiences uplift. Where tips and bends are below ground in active fault systems, this zone of hangingwall deformation may affect the earth's surface, producing geomorphic features that reflect movement on the underlying fault.

Intracontinental regions contain characteristics that may influence fault blindness aside from the influence of the presence or absence of surficial geology on the recognition of fault rupture (*See* Section 2.4.4.3). Due to the long deformational history and relatively thick seismogenic crust, faults that are not genetically blind but that lie within basement beneath a lithified sedimentary cover may be reactivated at depths from which rupture may be unable to propagate upward through the basement/cover interface. Similarly, in actively deforming regions of anomalously thick seismogenic crust, such as collisional belts experiencing continental subduction where seismicity persists to great depths, fault ruptures may be blind simply due to their great nucleation depths.

2.4.4.3 Stratigraphic Features

The interaction between blind faults and surficial stratigraphic processes are not as complex as for emergent faults, but are more subtle and, as a result of the distributed nature of the deformation, can have an effect over a very broad region. For blind reverse and thrust faults, the effects can be seen in both the stratigraphic architecture formed over a growing fold (i.e., growth stratigraphy) (e.g., Medwedeff, 1989; Suppe et al., 1992; Hardy and Poblet, 1995; Poblet et al., 1997) and its influence on both regional basin formation and migration (i.e., piggy-back basins and foreland basins) (e.g., Burbank and Tahirkheli, 1985; Johnson et al., 1986; Burbank et al., 1992), and local fluvial and depositional systems (i.e., stream sinuosity, terrace formation and uplift, (e.g., Ouchi, 1985; Bullard and Lettis, 1993; Merritts et al., 1994).

The relative lack of observed surface rupture during moderate to large intraplate earthquakes suggests that the intraplate environment may be conducive to blind faulting. Many intraplate regions, such as shield regions (e.g., Canadian and Arabian shields) or exhumed mountain belts with only a thin Holocene cover (e.g., Appalachian Mountains), are characterized by a lack of Quaternary deposits and landforms of sufficient age that could be used to identify surficial deformation related to active blind faulting. Similarly, in regions where unconsolidated surficial deposits are very young and thick, individual ruptures and even cumulative deformation may not be recorded due to the low strain rates typical of intracontinental regions relative to the high rates of sedimentation (e.g., Mississippi embayment, coastal plain of the southeastern United States).

These relationships also are supported by the empirical analyses of Lettis et al. (1997). These authors present data on reverse earthquakes worldwide indicating that for interplate regions, reverse earthquakes are most often associated with young contractional deformation that is recognizable from the surrounding geology on a regional scale. For the majority of intraplate reverse earthquakes, however, an association could not be made with recognizable, young deformation.

2.5 Nonseismogenic-Tectonic Faults

In addition to ground rupture on the causative fault (primary faulting), permanent surface deformation may result from earthquakes as a result of triggered slip, displacement on nearby faults (sympathetic or secondary faulting), bedding-plane slip and extension to accommodate coseismic folding, and shaking-induced ground failures related to liquefaction and landsliding. Although these faults are produced by earthquakes, the faults themselves are not capable of generating a moderate to large earthquake (i.e., they are nonseismogenic). These nonseismogenic tectonic faults are summarized in the following sections.

We have compiled relevant characteristics for each of these faults, and grouped the characteristics according to regional context, local context, or fault-specific characteristics. Regional context includes characteristics or associations with the regional tectonic setting, the regional geologic setting, the present-day stress or strain fields, the regional geophysical setting, and the regional seismologic setting. These characteristics are important to note, because some faults are more common in (or restricted to) certain regional settings.

Local context includes characteristics of the fault associated with the local topographic, geomorphic, stratigraphic, structural, geophysical, seismologic, and hydrologic settings. Fault-specific characteristics are grouped into spatial and temporal characteristics. For each of the faults described in Sections 2.5 and 2.6, these tabulations of characteristics provide a basis for differentiating between tectonic and nontectonic faults, as presented in Section 3.

2.5.1 Triggered Slip

Triggered slip, or sympathetic surface rupture, is a form of a seismic fault creep that coincides closely in time with a large nearby earthquake, but is not along the primary rupture (Sylvester, 1986). By definition, triggered slip occurs on primary and secondary tectonic faults, and thus is not associated with non-seismogenic processes that can produce faults. Triggered slip on faults is similar to ground cracks and other phenomena produced by primary surface rupture. Therefore, if triggered slip is documented along a fault, the fault should be considered either a primary or secondary tectonic structure.

The causative mechanism and seismotectonic significance of triggered slip is not known. Williams et al. (1988) concluded that a variety of factors were involved in determining where triggered slip took place and how much offset was observed. The amount of slip, although controlled in part by the amount of elastic strain accumulation on the fault itself, also seems to be affected by the presumed level of strong ground motion. Based on observations from the 1968 Borrego Mountain earthquake, Allen et al. (1972) concluded that triggered slip on the San Andreas fault was not driven by the static strains imposed by the remote fault rupture, but more likely was dynamically triggered by the passage of transient seismic waves, and that it represented the release of shallow tectonic prestrain as a result of a slip deficit near the surface on creeping faults (*See* also Sharp et al., 1986a, b).

In central California, both triggered slip driven by static strain changes and dynamically triggered slip have been postulated. Simpson et al. (1988) have showed that coseismic static stress changes, together with a linear viscous creep rheology, can explain some observed triggered-slip creep. Bodin et al. (1994) proposed that seismic surface waves, perhaps amplified by sediments, generate transient local conditions that favor the release of tectonic strain to varying depths. They propose that the amplitude of triggered slip may be proportional to the depth of slip in the creep event and to the available near-surface tectonic strain that would otherwise eventually be released as fault creep. They further suggest based on synthetic strain seismograms from the Landers sequence that pore pressure during periods of fault-normal contraction may be responsible for triggered slip, since maximum dextral shear strain transients correspond to times of maximum fault-normal contraction. Various researchers also have observed that although some examples of slip appear to coincide with the passage of seismic waves from the mainshock, triggered slip may also develop slowly, or have delayed onset (e.g., Allen et al., 1972; Williams et al., 1988; McGill et al., 1989). Some triggered slip also is postulated to be a form of long-term afterslip following earthquakes that occurred many years or decades earlier (Hudnut and Clark, 1989). Numerous recent studies have modeled static stress changes associated with historical large magnitude earthquakes and the associated effects of triggering and inhibition of earthquakes (e.g., Harris and Simpson, 1992; Jaumé and Sykes, 1992; Stein et al., 1992; King et al., 1994).

Some of the best documented examples of this triggered slip occurred during historical earthquakes in Southern California. Bodin et al. (1994) noted the following with regard to these events:

(1) Triggered slip of as much as 25 mm has been observed, but offsets commonly are less.

(2) Triggered slip commonly is observed as surface cracks, which offer little or no information on the timing of triggered slip development.

(3) Prior to the 1992 Landers earthquake, triggered slip was observed exclusively on faults known to be creeping aseismically.

(4) The Landers sequence caused small amounts of slip on faults in the region that apparently did not have any deep displacement and were not known to have been creeping prior to the Landers sequence.

(5) There is no simple relation between the amount of triggered slip and its distance from a causal earthquake.

(6) Creepmeter observations indicate that triggered slip typically is initiated close to the time of arrival of seismic waves propagating from large nearby earthquakes.

(7) Triggered slip can occur repeatedly on a fault.

(8) Triggered slip commonly is reported from creeping faults that pass through sediment-filled basins.

(9) In the case of the southern San Andreas, Bilham and Williams (1985) noted and Williams et al. (1988) confirmed that triggered slip occurs only along reaches oblique to the inferred regional slip vector where fault-normal contraction occurs.

The above studies indicate that although triggered slip is a secondary process, the faults on which triggered slip occurred are themselves primary tectonic seismogenic structures.

There also exists abundant evidence that earthquake occurrence and associated ground deformation can be triggered by stress changes of several tens of bars (sometimes only several bars) induced by human activities, such as reservoir impoundment (Simpson, 1986; Roeloffs, 1988), deep-well injection, fluid extraction, salt solution mining (Wesson and Nicholson, 1987; Segall, 1989), and quarry off-loading (Pomeroy et al., 1976; Yerkes et al., 1983; Sylvester and Heinemann, 1996). Simpson et al. (1986) provides a summary of these types of triggered earthquakes and a discussion of the mechanism and influence of induced stresses responsible for their occurrence. Case histories for stress release faulting associated with such activities are described in Section 2.6.5. Surface deformation caused by subsidence due to fluid extraction is described in Section 2.6.6.

2.5.2 Secondary Tectonic Faults

Secondary tectonic faults are subordinate structures that accommodate deformation during slip events on primary seismogenic tectonic faults, but which do not contribute significantly to the seismic moment release during earthquake rupture (e.g., Yeats, 1986). According to this usage, secondary tectonic faults form as a "passive" mechanical response to non-uniform slip, and to the development of localized stresses during displacement on non-planar fault segments (e.g., Price, 1968; Segall and Pollard, 1980; Bilham and King, 1989; Chester and Fletcher, 1997). Considerable information is available that describes secondary tectonic features and their mechanical significance with respect to the development of larger structures as observed in rocks and physical models and inferred from theory (e.g., Ramsay, 1967, Tchalenko, 1970; Friedman et al., 1976; Wilson, 1982; Ramsay and Huber, 1987; Nicolas, 1987; Price and Cosgrove, 1990; Twiss and Moores, 1992). Secondary tectonic faults that slip during moderate and large earthquakes have been recognized in association with all styles of active faulting. Coseismic ground ruptures on secondary faults also have been documented during post-earthquake investigations. A brief overview of the various types of secondary faults associated with historical and pre-historical earthquakes in compressional, extensional, and translational tectonic environments is provided below.

For each of these faults, we provide a table summarizing characteristics, grouped by: (1) regional context, (2) local context, and (3) fault-specific characteristics. We identify five elements of regional context that are relevant: tectonic, geologic, relation to the stress/strain field, geophysical, and seismologic characteristics. Similar elements of local context also are described. For fault-specific characteristics, the tables include spatial and temporal characteristics of the fault.

2.5.2.1 Compressional Tectonic Environments

Secondary tectonic faults observed in association with active compressional tectonic environments are flexural slip faults, bending-moment faults, chordal faults, relay (tear) faults, conjugate strike-slip faults, and hanging wall collapse faults. Flexural slip and bending-moment faults are the most common of these structures and are related directly to fold deformation associated with thrust and reverse faulting. Yeats (1986) provides an excellent review of these structures and their occurrence in active tectonic settings. **Flexural slip** includes reverse-displacement, bedding-parallel displacement between anisotropic, layered strata within fold limbs during contraction. Active flexural slip faults have been recognized in the Coast Ranges and Transverse Ranges of California (Cluff et al., 1981; Yeats et al., 1981; Lettis, 1985; Keller et al., 1982; Rockwell, 1983, 1988; Yerkes et al., 1983; Asquith, 1985; Treiman, 1995; Sylvester and Heinemann, 1996); Japan (Ota, 1969: Ota and Suzuki, 1979); the Gray-Inangahua basin, New Zealand (Suggate, 1957; Lensen and Suggate, 1968; Boyes, 1971; Lensen and Otway, 1971); the Zagros Mountains of Iran (Berberian, 1979); and northern Algeria (Philip and Meghraoui, 1983).

Bending-moment faults form in the hinge regions of folds during buckling of an effectively homogenous, isotropic rock layer (e.g., Price and Cosgrove, 1990). If flexural slip (or flow) cannot occur due to the isotropic nature of the rock (e.g., a very thick sandstone), then lengthening and extension will occur in the outer arc region of the hinge and a similar and opposite, contractional effect will occur in the inner arc region. Active bending-moment faults have been documented in the Transverse Ranges of California (Sarna-Wojcicki et al., 1976; Gardner, 1982); the Columbia plateau, Washington (Campbell and Bentley, 1981; West, 1996); the Santa Cruz Mountains, California (Cotton et al., 1990); the Reelfoot Scarp, Tennessee (Kelson et al., 1996a); El Asnam, Algeria (King and Vita-Finzi, 1981; Philip and Meghraoui, 1983); and Spitak, Armenia (Philip et al., 1992). Notably, sediments deposited in a graben bounded by bending-moment faults along the Reelfoot scarp in New Madrid Seismic Zone provided data on the timing and recurrence of post large earthquakes (Kelson et al., 1996a).

Other styles of secondary faulting in compressional tectonic environments that are not as widely observed (or cited) include: **conjugate** and **en echelon strike-slip faults** that form near the lateral termination of a thrust rupture (e.g., 1980 El Asnam [Philip and Meghraoui, 1983] and 1988 Spitak [Philip et al., 1992] surface ruptures); **hangingwall collapse faults** that form along the tip of an emergent thrust fault (Philip et al., 1992);

and strike-slip or **"chordal" faults** formed in the hanging wall of a thrust (e.g., the 1968 Meckering, Australia, surface rupture [Gordon, 1971]).

2.5.2.2 Extensional Tectonic Environments

Secondary faults produced during normal fault ruptures in extensional tectonic environments include both antithetic and synthetic normal faults, antithetic thrust faults, bending-moment faults, and cross faults. A wide variety of coseismic secondary faults that developed during normal fault ruptures were observed and recorded following the 1959 Hebgen Lake, Montana and 1983 Borah Peak, Idaho earthquakes. Crone et al. (1987) document a 100-m-wide zone of **antithetic** and **sympathetic secondary normal faults** within the hanging wall of the Borah Peak rupture. This zone is bounded on the east by the west-dipping primary normal fault and on the west by a low-angle **antithetic thrust** fault that dips to the east. A similar, secondary low-angle thrust fault rupture was observed in association with the Hebgen Lake earthquake (Crone et al, 1987). Both footwall and hanging wall **cross faults** are among the wide range of features described by Stewart and Hancock (1991) for active normal faults in the Aegean region in the northern Mediterranean. These authors also refer to minor faults formed in a stepover region between two active, en echelon normal faults. Wallace (1984) described similar secondary faults associated with for the Pleasant Valley, Nevada surface rupture.

Coseismic antithetic secondary faults also have been documented by paleoseismic studies conducted along the Wasatch fault in Utah (e.g., Swan et al., 1980, 1981: McCalpin et al., 1994). These investigations identified paleoearthquakes based on sediments deposited within a hanging-wall graben, which is bounded on the east by the main Wasatch fault strand and on the west by a secondary, coseismic antithetic fault. Swan et al. (1981) identified antithetic faults in the hanging wall as much as 350 m outboard of the primary, range front fault. In addition to the usefulness of identifying the coseismic or dependent nature of a secondary fault, these studies demonstrate that paleoseismic studies of secondary faults may be used to help characterize the associated primary seismogenic faults.

2.5.2.3 Strike-Slip Tectonic Environments

The map pattern and internal structure of strike-slip faults is very complex and a wide variety of kinematic styles and fracture geometries have been observed in both active, ancient, and physically modeled fault systems (e.g., Riedel, 1929; Tchalenko, 1968; Wilcox et al., 1973; Bartlett et al., 1981; Sibson, 1985; Woodcock and Fischer, 1986; Sylvester, 1988). The most common secondary features that form along active, primary strike-slip faults are:

(1) extensional faults and related "pull-apart" basins formed at releasing bends and stepovers;

(2) contractional faults and related uplifts formed at restraining bends and stepovers; and

(3) secondary strike-slip fault splays that emanate from the bends and tip regions of the primary fault.

The following discussion cites many examples of secondary faults and related features that are large enough in scale to be considered independent seismogenic sources, primarily because the larger features are better documented. Because of the scale-invariant nature of fault-related structures, however, descriptions of the larger features is applicable to local- and outcrop-scale in terms of geometry and kinematic relationship to a "parent" primary fault. The degree to which a fault splay may represent an independent seismogenic source, however, is largely a matter of scale and should be carefully considered during investigations of an active fault (e.g., Lettis and Hanson, 1991).

Secondary extensional faults and related basins along active, primary strike-slip faults are well documented (e.g., Crowell, 1974; Dibblee, 1977; Aydin and Nur, 1982; Mann et al., 1983; Biddle and Christie-Blick, 1985). Excellent examples of local-scale secondary normal faults along dilational jogs and bends were observed during post-earthquake investigations of the 1992 Landers (e.g., Zachariasen and Sieh; Sowers et al., 1994) and 1979 Imperial Valley strike-sliprupture (Yeats et al., 1997). Because pull-apart basins are areas of deposition, fault behavior (e.g., recency, magnitude, style, etc.) commonly is preserved in the stratigraphic record and, if accessible, can be used in paleoseismic studies (e.g., Di Silvestro et al., 1990; Williams, 1992).

Secondary contractional faults associated with contractional stepovers and bends along active strike-slip faults also have been widely recognized. Kilometer-scale examples are common along the San Andreas fault (e.g., Harding, 1974; Sylvester and Smith, 1976; McLaughlin, 1990; Hector and Unruh, 1992; Angell and Hall, 1993; Burgmann et al., 1994), the North Anatolian fault in Turkey (Kadinsky-Cade and Barka, 1988), the Alpine fault in New Zealand (e.g., Berryman et al., 1992) and the Gobi-Altai fault in Mongolia (e.g., Baljinnyam et al., 1993; Schwartz et al., 1996).

Post-earthquake observations of coseismic, small-scale secondary contractional faults have been documented by Tchalenko and Ambreyseys (1970), for several strike-slip earthquakes in Iran. The 1968 Borrego Mountain, California earthquake ruptured through a 2-km-wide contractional stepover that contains a structurally and geomorphically well-defined region of uplift and folding (Brown et al., 1991). A detailed study of local-scale contractional secondary structures within a restraining bend was conducted during post-earthquake investigations of the 1992 Landers, California earthquake rupture (Aydin and Du, 1995; Spotila and Sieh, 1995). Evidence for coseismic secondary contractional faulting also has been identified in paleoseismic trench investigations within an 8-m-wide contractional stepover along the 1906 rupture trace of the San Andreas fault in northern California (Angell et al., 1991; Niemi and Hall, 1992).

2.5.2.4 Fault Swarms

Fault swarms are a common feature associated with regional active fault systems as well as coseismic fault rupture (e.g., Tchalenko, 1970; Johnson et al., 1994) in all tectonic settings. Fault swarms are broad zones of discontinuous, relatively short, small-displacement faults that are concentrated in regions adjacent to or between primary fault ruptures. Fault swarms, for example, have developed within "stepover zones" between major active faults (Stewart and Hancock, 1991). Whether these smaller faults coalesce to form a single fault at depth, or represent distributed deformation *sensu stricto* is not well understood. Understanding the scale and kinematic framework of the faults swarm with respect to adjacent primary faults is critical for determining the origin and significance of the swarm.

A recent example of fault swarms produced during a large earthquake is the deformation within the Santa Cruz Mountains during the M_w 7.1 1989 Loma Prieta earthquake. A broad, northwest-trending zone of coseismic ground fissures formed near the range crest, and have been attributed to several causes:

(1) tectonic extension and shaking-induced gravitational ridge-top spreading (Ponti and Wells, 1991; Hart et al., 1990);

(2) secondary flextural slip faulting along bedding within Tertiary sedimentary rocks that underlie the ridge crest (Cotton et al., 1990; Zoback and Reches, 1990); and

(3) right-lateral shear within the "Summit Ridge shear zone", which Johnson and Fleming (1993) interpret as the surface expression of the fault responsible for the Loma Prieta earthquake.

The evidence supporting each of these proposed models is summarized by Cotton (Appendix A). In addition, the 1989 Loma Prieta earthquake produced several zones of ground deformation within alluvium along the northeast margin of the Santa Cruz Mountains (Haugerud and Ellen, 1990; Langenheim et al., 1997; Langenheim et al., 1997). Detailed geologic mapping and geomorphic analysis of these deposits by Hitchcock et al. (1994) shows that this deformation is coincident with pre-existing zones of lineaments and other potentially fault-related geomorphic features. In this case, it appears that coseismic secondary faulting occurred along an up-dip projection of the fault that generated the main earthquake.

Similarly, the proximity of several zones of prominent ground cracks (fault swarms) in the epicentral area of the 1994 Northridge earthquake to previously mapped late Quaternary faults suggest a secondary tectonic relationship (Hart et al., 1995; Stewart et al., 1995; USGS-SCEC, 1994). In particular, a 5-km-long, west-trending belt of discontinuous ground cracking is coincident with part of the Mission Hills fault trend in the towns of Granada Hills and Mission Hills (Angell et al., 1994; Hart et al., 1995; Ponti et al., 1996; Johnson et al., 1996). Deformation within this zone consisted of tension cracks and graben concentrated along the northern margin of the zone, and compressional buckles and small thrusts along the southern margin (Hecker et al., 1995; Holzer et al., 1996). Although displacements were small, the linear extent of these zone is consistent with what might be expected for a surface-faulting earthquake of similar magnitude (USGS-SCEC, 1994). Peterson (1994) initially reported that these ruptures may have been caused by sympathetic slip on the Mission Hills fault. Subsequent detailed investigations, including analysis of geotechnical borings, pre- and post-earthquake survey data, and distribution of sewer-line damage, suggest that no displacement occurred on the Mission Hills fault in 1994 (Ponti et al., 1996). Ponti et al. (1996) also noted the lack of clear evidence of Holocene offset associated with the fault. They speculate, however, that the fault may have controlled the distribution of ground cracks by influencing depths of groundwater and by focusing and/or amplifying seismic-wave energy. However, geologic and geomorphic evidence suggests that surface deformation has occurred along the Mission Hills and Northridge Hills fault trends over geologic time. Detailed surficial mapping and geomorphic analysis by Hitchcock and Kelson (1996) suggest that repeated late Quaternary displacement has produced ground deformation associated with these fault trends. Based on analysis of deep seismic and well data, Tsutsumi and Yeats (1996) suggest that surface ruptures coincident with the Mission Hills and Northridge Hills fault trends are the result of secondary faulting related to flexural-slip folding.

The regional, local, and feature-specific characteristics of secondary faults are summarized in Table 2.1.

Table 2.1 Summary of characteristics: secondary tectonic faults

Regional context	**Secondary tectonic faults**
Tectonic	Occur in interplate or intraplate tectonic settings; more common in interplate regions due to higher strain rates.
Geologic	Tectonically active geologic environments; may be better developed or more easily recognized in brittle, stratified rocks.
Relation to stress/strain field	No consistent direct relationship; may be controlled by local stresses during movement on primary structure in response to regional tectonic stress.
Geophysical	May be associated with regional anomalies that reflect fault-related structures.
Seismologic	Positive spatial correlation with contemporary seismicity.
Local context	**Secondary tectonic faults**
Topographic	More common in areas of young topographic relief due to tectonic activity (i.e., areas of active uplift).
Geomorphic	Associated with landforms that reflect active tectonic environments (e.g., faceted range fronts, pressure ridges, uplifted terraces, subsiding basins).
Stratigraphic	More common in moderately- to well-consolidated, bedded sedimentary rocks due to mechanical effects of anisotropy; preservation and/or ability to recognize features also may be controlled by properties of host material (e.g., induration, ductility).
Structural	Commonly associated with a larger, identifiable structural feature, such as an emergent fault system, anticline or fault-bounded basin, often with a predictable relationship to the larger structure.
Geophysical	May be associated with local geophysical anomalies that reflect fault-related structural and/or stratigraphic features.
Seismologic	Possible local association with microseismicity; focal mechanisms, if available, may be compared to observed and/or theoretical displacement patterns.
Hydrologic	Possible association with springs, seeps, ponded and/or deflected drainages related to fault strands.
Fault characteristics	**Secondary tectonic faults**
Spatial characteristics	
Morphology	Highly variable; both low-angle and steeply dipping faults may be present; scarps may be subtle due to low displacement; typically planar along strike.
Geometry	*Plan View* – variable, often linear, relatively short (<1 km), may occur as lineament or fault "swarms." *Cross section* – variable, rooted at shallow depth, often curved (low

Table 2.1, continued

	strain).
	Aspect ratio (H:V) - variable to large, reflecting low slip on a long feature controlled by size of primary structure.
Scale	Highly variable - centimeters to hundreds of meters; subsidiary to, but may approach the size of, the larger structure (usually in strike dimension).
Sense of deformation	Directly related to kinematics and mechanics of larger, primary structure; kinematic analysis may be extremely useful; secondary fault "swarms" with low total strain may exhibit highly variable slip characteristics.
Depth	Typically occur above dip-slip faults (in hanging wall) and within <500 m of strike-slip faults; rooted at shallow depth into primary structure.
Spatial associations	Associated with larger, primary tectonic structure.
Hydrologic	May be associated with springs, seeps, and ponded or deflected drainages.
Temporal characteristics	
Rate of deformation	Commonly low (< 1 mm/yr) to moderate (1 < 5 mm/yr) strain rates; may be higher depending on slip rate of primary structure.
Episodicity	Usually episodic, possibly continuous if related to creeping primary fault; short-lived, transient creep possible due to post-seismic strain.
Duration of deformation	Short-term, may be mechanically viable for only short interval of deformation history of primary structure.
Recurrent deformation	Commonly there is evidence of recurrent movement, but may be related to a single event.
Temporal associations	Associated with movement events on larger, primary structure, and therefore with other secondary features as well.
Investigative techniques	Local geologic and geomorphic mapping at a scale sufficient to identify the larger, primary structure, kinematic analysis of observed deformation features, thorough investigation of possible nontectonic gravitational origin.
Key diagnostic characteristics	Small-scale mechanical and kinematic relationship to larger structure, small (decimeter- to millimeter-scale) slip per event.
References	Segall and Pollard (1980), Yeats (1986), Ramsay and Huber (1987), Price and Cosgrove (1990), Aydin and Du (1995), USGS (1996), Yeats et al. (1997).

2.5.3 Features Caused by Strong Ground Motions

Many historical moderate and large earthquakes show that strong ground motion may produce considerable amounts of near-surface deformation both near and far from the fault responsible for the earthquake. Ground deformation resulting from the effects of strong ground motion may occur on or above a fault strand (on-fault, or near-fault structures), as well as away from or far above a fault strand (off-fault, or far-field structures) (McCalpin and Nelson, 1996). Consequently, features caused by strong ground motion may occur over widespread areas, and may result from a variety of local processes induced by ground motions from a nearby or distant earthquake. The extent and distribution of ground deformation is governed by factors controlling the amount of strong ground shaking (e.g., earthquake magnitude, distance from fault plane, directivity, attenuation topographic effects), as well as susceptibility due to local geologic conditions (e.g., depth to groundwater, slope, geologic materials and structure). McCalpin and Nelson (1996) note that on-fault and off-fault secondary structures may be produced either instantaneously (during an earthquake), or as a result of post-seismic processes. On the basis of this classification, McCalpin and Nelson (1996) identify several types of geomorphic and stratigraphic features that may be related to seismic shaking, including: liquefaction-related features (e.g., sand blows, sand dikes, filled craters), subsidence from sediment compaction, and landslides. All of these features also may be produced by nonseismic processes, and thus interpretation of the origin of these features and their significance as secondary evidence of paleoearthquakes may be difficult in some cases. This section provides a brief summary of features caused by strong ground motion as a result of seismically induced soil liquefaction, compaction, and slope failure, with an emphasis on differentiating features related to strong ground motions from those produced by primary surface rupture.

The process of liquefaction is the transformation of a granular material from a solid state into a liquefied state as a consequence of increased pore-water pressure (Youd, 1973). This process may occur as a result of cyclic seismic shaking, during which the pore-water pressure within loosely packed, cohesionless sediments becomes equal to or greater than the static confining pressure, and the material becomes fluidized (Obermeier, 1996). This process occurs only within saturated cohesionless sediments, and generally within a depth range of a few meters to about 10 m. Liquefaction-induced features include **sand blows**, which are conical accumulations of sand at the ground surface that are deposited by the venting of the sand/water mixture to the surface (Dutton, 1889; Fuller, 1912; Youd and Hoose, 1978; Obermeier, 1989; Tuttle, 1994). The sand is extruded to the surface via fissures that are subsequently filled with sand and blocks of material adjacent to the fissure. These **clastic dikes** also may form by fracturing of upper soil layers and intrusion of liquefied sand. **Sand-filled craters** also may be produced by strong ground motions (Dutton, 1889; Amick et al., 1990; Obermeier, 1996). The identification of sand-blow deposits, clastic dikes, and sand-filled craters has been used in many locations to interpret the occurrence of prehistoric ground motions (e.g., Amick et al., 1990; Saucier, 1991; Tuttle and Schweig, 1996). Liquefaction of near-surface granular materials also may decrease the strength of the materials and result in **lateral spreads**, in which a surface soil layer moves laterally down a gentle (less than 5%) slope. The blocks of surface soil generally move toward a stream bank or other free face, and may be tilted, fissured, and intruded by clastic dikes. Decreased strength of near-surface granular deposits also may produce **localized depressions** as a result of densification of liquefied sediment after extrusion of water (Tokimatsu and Seed, 1987; Obermeier, 1996). On slopes steeper than about 5%, liquefied sediments may cause landslides termed **flow failures**, that may be laterally extensive and be very similar to nonseismic slope failures.

Surface fault fractures resulting from secondary effects due to strong ground shaking may be confused with coseismic fault ruptures. Case studies of surface deformation resulting from the 1994 Northridge earthquake provide good examples of the types of data and approaches that can be used to assess the tectonic significance of such surface deformation. Observations of compressional faults located near the up-dip projection of the thrust fault that caused the 1994 Northridge earthquake led to early speculation that the ground deformation seen there may have been the result of primary tectonic faulting (Rymer et al., 1995). Rymer et al. (1995) describe fracture zones around the margins of Potrero Canyon that exhibited vertical displacements of a few centimeters to greater than 61 cm. Based on the results of detailed surface and subsurface investigations, they concluded that these features formed by differential settlement and lurching during strong ground motion caused by the earthquake. They cite as evidence of a non-fault origin the development of fractures near the alluvium-bedrock contact that dip toward the canyon center from both sides of the valley. These fractures, therefore, are consistent with a net vertical settlement of the alluvial fill. They also note the presence of localized liquefaction, pipe breaks, and better developed crack sets on the down-gradient sides of ridge spurs that are consistent with this model. Numerous other coseismic ground failures that occurred in the Mission Hills and Granada Hills area during this earthquake (Hall, 1995; Hart et al., 1995; Stewart et al., 1995; Angell et al., 1994) exhibited orientations and displacements consistent with liquefaction-induced lateral spreading and compaction of loosely consolidated sediments (USGS-SCEC, 1995). Hecker et al. (1995) and Holzer et al. (1995) interpret that liquefaction at depth produced a linear series of extensional fractures and graben, as well as a parallel zone of contractional features further down slope in these areas. In contrast, Johnson et al. (1996) map the faults and interpret them as related to primary surface rupture on the Mission Hills fault, rather than as secondary, liquefaction-related faults.

In many cases, faults produced by seismically induced liquefaction are comparable to features produced by increased pore-water pressures from flooding, and by rapid sedimentation (depositional loading), permafrost, and other physical and biological processes. Obermeier (1996) provides five criteria from which to identify seismically induced liquefaction. These are:

(1) evidence of an abrupt, short-lived, upward-directed hydraulic force

(2) sedimentary characteristics similar to those documented from historic liquefaction

(3) shallow groundwater conditions favorable for seismically induced liquefaction

(4) presence of features at multiple locations such that the regional pattern of the size and abundance of features is consistent with a strong ground motion source

(5) ages of features are consistent with one or more discrete, short episodes of deformation such that the temporal pattern of the features is consistent with single or recurrent strong ground motion

Obermeier (1996) also presents descriptions of features that may be misinterpreted as resulting from paleoliquefaction, including artesian conditions (Kolb, 1976; Holzer and Clark, 1993; Li et al., 1996), tree-throw craters, mima mounds (Berg, 1990), syndepositional sedimentary structures, soil weathering features (Obermeier et al., 1990), and periglacial features. These features are described in detail by Obermeier (1996) and references cited therein.

Strong ground motion also may produce compaction of unconsolidated sediment and differential movement at the ground surface, producing small scarps and fissures on relatively level ground. As noted above, this settlement may occur as a result of densification related to liquefaction of saturated sediments, although unsaturated sediments also may undergo some settlement due to strong ground motion. The orientation and lateral continuity of features at the ground surface may be controlled by near-surface stratigraphy, topography, and depth to groundwater, and thus may be readily differentiated from possible tectonic surface rupture. However, some scarps or fissures may have orientations that are consistent with tectonic surface deformation, and thus it may be difficult to difficult to determine the origin of some features. In particular, evaluating the origin of prehistoric, seismically induced settlement cracks or fissures may be difficult if there are no data on the likely orientations and locations of possible seismogenic sources. Tokimatsu and Seed (1987) provide data on the amount of settlement that may be expected during strong ground motion, with volumetric strains being generally less than 5% in unconsolidated sediments. Assessing the stratigraphy of sediments beneath a site may allow a comparison of displacements observed at the surface and the expected amount of settlement. Strong ground shaking also may produce slope failures at scales ranging from shallow soil slips to large-scale gravitational spreading features. Descriptions of surface deformation, including faulting that may be associated with landslides and large-scale gravitational spreading are provided by Cotton (Appendix A), McCalpin (Appendix A), and Jibson (1996). Clearly, earthquakes can trigger all types of landslides, and all types of landslides triggered by earthquakes can occur without seismic triggering (Jibson, 1996). Keefer (1984) notes that, in general, the more disrupted types of landslides (e.g., rock falls, rock slides) are much more abundant during large earthquakes than the more coherent landslides (e.g., slow earth flows, rock block slides). In addition, Keefer (1984) suggests that slope materials that are weathered, sheared, intensely fractured or jointed, or saturated are particularly susceptible to landsliding during earthquakes.

Regional, local, and feature-specific characteristics of faults caused by strong ground motion are summarized in Table 2.2.

Table 2.2 Summary of characteristics: faults caused by strong ground motions

Regional context	Strong ground motion faults
Tectonic	Faults occur in both interplate and intraplate tectonic settings; more common in interplate regions due to higher strain rates.
Geologic	Present in tectonically active geologic environments; may be better developed and/or more abundant in areas containing extensive unconsolidated or semiconsolidated surficial deposits.
Relation to stress/strain fields	No consistent relationship to regional stress field.

Table 2.2 (continued)

Geophysical	Not associated with a particular geophysical environment; may be associated with regional anomalies that reflect fault-related regional structures.
Seismologic	May be correlative with regional contemporary seismicity; strong ground motions are required for formation; presence of similar features in areas lacking contemporary seismicity may indicate occurrence of paleoearthquakes.
Local context	**Strong ground motion faults**
Topographic	Slope failure susceptibility of fault and scale are dependent on topographic relief. Liquefaction-related features and ground settlement generally occur in areas of low topographic relief. Topographically-induced amplification of strong ground motions (e.g., ridges, basins) may localize features.
Geomorphic	Common in areas containing geomorphic evidence of nonseismic slope failures; liquefaction-related and settlement features tend to be associated with young geomorphic surfaces underlain by unconsolidated surficial deposits.
Stratigraphic	Slope failures commonly are associated with previous landslide deposits. Liquefaction-related and settlement features tend to be associated with shallow, unconsolidated sand-rich deposits (e.g., fluvial, lacustrine, or eolian).
Structural	Not necessarily associated with geologic structure except for those contributing to slope failure in general (e.g., dip-slip discontinuities) in areas where distribution of strong ground motions is influenced by directivity of earthquake waves.
Geophysical	May be associated with local geophysical anomalies that reflect fault-related structural and/or stratigraphic features
Seismologic	Possible local association with seismicity; conversely, presence of features may indicate paleoseismicity.
Hydrologic	Slope failures commonly associated with saturated soil conditions, although seismically induced rockfalls may be unrelated to subsurface water conditions; liquefaction-related features associated with shallow groundwater (<13 m).

Table 2.2 (continued)

Feature Characteristics	Strong Ground Motion Faults	
Spatial characteristics		
Morphology	Morphology of seismically induced slope failures is indistinguishable from nonseismic slope failures. Arcuate headscarps and bulging toe, typically concave toward center of slide mass. Liquefaction-related features include linear sand ridges, sand-filled craters, lateral spreads, and sand blows. Minor scarps and local depressions due to settlement/compaction also are common.	
Geometry	*Plan view -*	Laterally discontinuous; commonly arcuate headscarp.
	Cross section -	Basal slide plane, commonly listric (rotational) or planar (translational); depression at head, relief at toe.
	Aspect ratio (H:V) -	Variable; typically low to very low.
Scale	Scale of seismically induced slope failures is indistinguishable from nonseismic slope failures. Variable - < 0.1 km to 1.0 km; rarely larger than one kilometer Liquefaction-related features: linear sand ridges may be up to a kilometer long; sand-filled craters are as large as one meter in diameter; lateral spreads may be kilometers in extent and have displacements of hundreds of meters; and sand blows are as much as tens of meters in diameter. Closed depressions typically are tens of meters wide and a few meters deep, although larger features are possible.	
Sense of deformation	Sense of deformation in seismically induced slope failures is predominantly extensional and indistinguishable from nonseismic slope failures. Extensional at head scarp, right-lateral at right margin, left-lateral at left margin, and reverse at toe (all or none of which may be preserved); overall consistency with downslope movement (local variations in displacement vector can occur). Liquefaction-related features: linear sand ridges, sand-filled craters, sand blows, and settlement depressions may be associated with shallow subsurface extensional faults having linear or semicircular geometries.	
Depth	Depth of deformation in seismically induced slope failures comparable with depths of nonseismic slope failures. Generally shallow depth/thickness (estimated maximum ~ 200 m). Liquefaction-related features: liquefaction-related and seismically induced settlement features generally are shallow (<13 m).	

Table 2.2 (continued)

Spatial associations	Proximity to seismogenic fault is required, but presence of features is strongly influenced by local and regional attenuation. Seismically induced slope failures commonly are associated with nonseismic slope failures. Different liquefaction-related features may or may not occur together within a local area, and may be associated with evidence of tectonic surface rupture.
Hydrologic	Slope failures commonly associated with saturated soil conditions, although seismically induced rockfalls may be unrelated to subsurface water conditions. Liquefaction-related features associated with shallow groundwater (<13 m); absence of present-day shallow groundwater may not preclude possibility of paleoliquefaction if groundwater conditions have changed.
Temporal characteristics	
Rate of deformation	Generally high short-term strain rates, with moderate to large displacements occurring during and following strong ground motions on the scale of seconds to hours. Long-term strain rates are considerably less if dependent on recurrence of strong ground motions.
Episodicity	Time series is commensurate with episodicity of strong ground motions: features at a given site may be formed episodically if strong ground motions are episodic; features may be formed as a result of ground motions generated from multiple sources and thus reflect the episodicity of sources in a regional context.
Duration of deformation	Features commonly are produced during generally short intervals of strong ground motions (duration of seconds). An exception is the continued venting of sand blows for minutes after cessation of strong ground motions.
Recurrent deformation	Although features may result from a single earthquake, recurrent strong ground motions generally produce additional movement on or augmentation of features.
Temporal associations	Seismically induced slope failures form instantaneously at geologic time scales; demonstration of contemporaneity of several slope failures and/or liquefaction-related features with known tectonic features may indicate a seismic origin. However, multiple slope failures also may form simultaneously during nonseismic (climatic) events, so that many independent lines of evidence are required to show a seismic origin of slope failures. Attributing other features to strong ground motions also requires demonstration of contemporaneity with active tectonic features.

Table 2.2 (continued)

Investigative techniques	Slope failures: Regional, local, and detailed geologic and geomorphic mapping; trench and test-pit excavation; small-diameter boreholes; large-diameter boreholes; installation of inclinometers; Slope stability back-calculations; radiometric and other dating techniques. Liquefaction-related features: Air-photo interpretation; local and detailed geologic and geomorphic mapping; analysis of groundwater data; excavation of trenches and test pits; analysis of liquefaction susceptibility; radiometric and other dating techniques.
References	Jibson (1996), Keefer (1984), Cotton (this volume), McCalpin (this volume), McCalpin and Nelson (1996), Obermeier (1996), Tuttle and Schweig (1996), Holzer and Clark (1993), Holzer et al. (1995), Li et al. (1996), Bull (1996).

2.6 Nonseismogenic-Nontectonic Faults

2.6.1 Landslide Phenomena

Mass wasting phenomena are classified by material types and type of movement (Varnes, 1978). Landslides include falls, topples, slides, lateral spreads, flows and combination of two or more principal types of movement. Of these, only **slides, lateral spreads** and **flows** produce discrete shearing of earth materials, and the development of rupture surfaces both within the body and at the boundaries of the landslide (*See* Figures WC-1 and WC-2, Cotton, Appendix A). A discussion of the characteristics of landslide faults that may be confused with tectonic faulting is provided by Cotton (Appendix A).

Sliding develops stresses in brittle earth materials that produce faults of normal-, reverse-, and strike-slip displacements, and their associated geomorphic features at a variety of scales. The worldwide geologic record contains excellent examples of faulted rock materials that are the product of large gravity sliding events along detachment surfaces that are regional in extent, some of which have subsequently undergone deformation and deep erosion. A description of the characteristics of large, ancient bedrock landslides in the Appalachian Valley and Ridge province of eastern North America is provided by Schultz (1986) and Schultz and Southworth (1989). Only with extensive field studies can these large fault surfaces of nontectonic origin be accurately recognized. On a more local scale, landslides commonly are localized along preexisting shear zones or zones of weakness, making it difficult to differentiate between superimposed tectonic and landslide features. Also, landslides may be caused by earthquakes (Keefer, 1984).

The boundaries of a landslide may be delineated by one or more types of faults. In the head area of a landslide, where the slide mass is moving downslope and away from stable ground, tensional stresses produce extensional features such as horsts and graben bounded by normal faults. Alternatively, in response to compression, the toe area is associated with contractional structures, including thrust faults and sinuous, bulging ridges (commonly antiformal) that overrun the intact hillside or a valley edge below the slide. Left-lateral and right-lateral strike-slip faults may define the left and right margins (looking downslope) of the landslide, respectively. The sense of slip changes by gradual transition from normal slip at the head to left- and right-lateral strike-slip along the flanks to reverse slip at the toe of the landslide (*See* Figure 3, Cotton, Appendix A). Landsliding may be expressed geomorphically by a number of features including arcuate headscarps, hummocky terrain, ponded water, bulging toe, and displaced drainages.

Many geomorphic features (e.g., ridges, sag ponds, deflected drainages) and microstratigraphic deposits (e.g., colluvial wedges, juxtaposed strata, buried soils) that are considered by paleoseismologists to be indicators of young tectonic faulting also can be produced by sliding processes. Subsurface investigations (i.e., large-diameter borings and trench excavations), however, commonly encounter basal rupture surfaces unique to landslides. These surfaces may be characterized by subplanar slickensided and striated shear surfaces, and accumulations of clay gouge and crushed fabric of adjacent earth materials. Anomalies in the orientation and juxtaposition of microstratification, and the morphology of the faulted rock and sediments may result from sliding.

Very large active slides, such as those associated with both subaerial and submarine volcanic centers, have produced seismic activity (Eissler and Kanamori, 1987). However, the density, pattern and historical record of this activity is not as high as active tectonic fault systems. A long-term record of historical seismicity, microseismicity, geophysical and geodetic anomalies common to active fault systems, is not commonly characteristic of slide processes.

Regional, local, and feature-specific characteristics of landslide features are summarized in Table 2.3.

Table 2.3 Summary of characteristics: landslide faults

Regional context	**Landslide faults**
Tectonic	Not restricted by regional tectonic setting; more common in active tectonic environments due to strong ground motion and tectonic uplift.
Relation to tectonic stress/strain fields	No consistent relationship with regional stress or strain fields.
Geologic	Not restricted by regional geologic setting.
Geophysical	Not associated with any particular geophysical setting.
Seismologic	Not associated with any particular seismologic setting; may be more common in seismically active regions due to strong ground motions and active uplift.
Local context	**Landslide faults**
Topographic	Common in mountainous or high-relief terrain, but may occur in moderate- to low-relief areas where geologic and/or topographic conditions favor instability. Topographic relief sufficient to create the gravitational potential energy is required to overcome geometric and physical resistance to internal sliding (i.e., lower for weaker materials, higher for stronger materials).
Geomorphic	Hummocky terrain; closed depressions on hillslope; headscarp at trailing edge, commonly arcuate; bulged region at leading edge; displaced drainages.
Stratigraphic	Presence of unstable geologic strata (e.g., non-lithified, highly fractured) and/or weak (detachment) horizon(s).

Table 2.3 (continued)

Structural	May be associated with one or more sets of planar discontinuities in a rock mass, the surfaces or intersections of which dip greater than 5°-10° and less steeply than hillslope (e.g., dip-slope failure); may have complex internal structure not consistent with adjacent rock (e.g., slope reversals on dip slopes, abrupt angular formation contacts, truncation of formation boundaries, and unusual changes in strike and dip of bedrock).
Geophysical	Not associated with any particular local geophysical context.
Seismologic	Generally non-seismogenic, although rare cases of seismogenic landslides have been proposed (identified based on single-couple focal mechanism).
Hydrologic	Geologic conditions that result in perched or impounded water (e.g., aquitards, faults).

Feature characteristics	**Landslide faults**
Spatial characteristics	
Morphology	(*See* Table 2.2)
Geometry	(*See* Table 2.2)
Structure	Have basal slide plane or rupture surface that commonly is listric. Normal fault(s) at upslope margin; left-lateral and right-lateral faults along margins; and arcuate thrust faults and folds at toe. Internal faulting (e.g., brecciation) may be present. Fault rock developed in zones of shearing.
Scale	(*See* Table 2.2)
Sense of deformation	(*See* Table 2.2)
Depth	(*See* Table 2.2)
Spatial associations	May be localized along structural zones of weakness and/or within unstable strata.
Hydrologic	Common association with springs and/or shallow groundwater.
Temporal characteristics	
Rate of deformation	Variable: Commonly high to very high strain rates (mm/year to m/day); catastrophic failures may have higher rates.
Episodicity	Episodic and continuous; closely related to strain rate and climatic/tectonic forcing conditions.
Duration of deformation	Generally less than tens of years; although preexisting slides may be partially reactivated.
Recurrent deformation	Common evidence of recurrent movement, but may be a single event.
Temporal associations	May be triggered by strong ground motion or extreme climatic event resulting in regional, coeval slope failures.

Table 2.3 (continued)

Investigative techniques	Geomorphic/geologic mapping ("Landslide Inventory Map"); kinematic analysis; exploratory trenching; instrumentation; drilling investigations; large-diameter boreholes.
References	Varnes (1978); Fleming and Johnson (1989); Voight (1978); Cotton (Appendix A).

2.6.2 Sackungen

The term "sackungen' (from the German verb "to sag") describes a family of landforms that occur in mountainous areas, particularly near or on ridge tops, that include **ridge-crest troughs, antislope scarps**, and **closed depressions** (*See* McCalpin, Appendix A). A deep-seated rock-failure origin for sackungen is implied by their topographic occurrence (high on ridge flanks, at ridge crests) where an erosional origin is unlikely. The most common geomorphic features of sackung can be divided into four categories: downhill-facing scarps, double-crested ridges, uphill-facing scarps, and notches on ridge axes (*See* Figure JM-1, McCalpin, Appendix A). Double-crested ridges (the *doppelgrat* of Zischinsky, 1969) are a classic sackung landform. The axial depression typical of *sackungen* is difficult to explain by a fluvial erosional process. Streams required to erode an axial trough are unlikely to flow down the crest of a ridge, and closed depressions could not have been excavated by running water. Sackungen typically are short, discontinuous, arcuate, and occur in swarms of multiple parallel scarps. Uphill-facing scarps and sidehill benches (degraded scarps?) are the most common sackung landform (e.g., Varnes et al., 1989). Along strike, sackungen may grade from antislope scarps to benches, and grabens may grade into irregular closed depressions; along-strike changes in height and morphology are common.

An inventory of published sackungen scarp dimensions (Woodward-Clyde Consultants, 1978) yields these ranges as being typical: scarp lengths of 15 to 300 m; scarp heights of 1 to 9 m; slope heights of 400 to 1200 m; and slope gradients of 25° to 50°. More recently, Salvi and Nardi (1995) interpret a trough 100 m deep, 700 m wide, 9 km long in the Apennines (Italy) as an earthquake-induced sackung. Almost all of these landforms are found at or near the crest of slopes, in a zone of tensional failure for slope-failure processes. In contrast, very few distinctive landforms have been observed on the lower slopes, except where authors have postulated that lower slopes have been "oversteepened" or bulged outward by compressive forces at the "toe" of a creeping rock mass.

Varnes et al. (1989) distinguish three types of sackungen:

(1) spreading of rigid rocks overlying soft rocks (Radbruch-Hall, 1978; Radruch-Hall et al., 1976)

(2) sagging and bending of foliated phyllites, schists, and gneisses (true *Sackungen* of Zischinsky, 1969)

(3) differential displacements in hard but fractured crystalline igneous rocks

Two opposing hypotheses have been proposed for the subsurface geometry (and hence mechanism of failure) of sackungen in massive competent rocks. One, held by Zischinsky (1969) and other workers, proposes that "...a well-defined slide plane near the headscarp passes downward into a broader zone of rock creep." Consequently, the lower portion of this type of failure bulges out into the valley (Morton and Sadler, 1989). Radbruch-Hall (1978) claims that rock creep can extend to depths of several hundred meters. However, there are no well-documented examples where the depth or shape of the failure plane has been observed and measured directly.

The most detailed study of subsurface deformation features (folds, faults) associated with rock creep is that of Chigira (1992), who characterized the micro- and meso-scale characteristics of the causative shear zones that underlie sackungen at depths of tens to hundreds of meters. In densely foliated rock, the phyllitic zone may form by microscopic slip along preexisting foliations. In sparsely foliated and massive rock, the brecciated zone may form by coalescence of newly formed and existing tension fractures and random crushing. Subsurface shear zones created by deep-seated gravitational creep are consistently asymmetrical, with a sharp upper contact and a transitional lower contact (*See* Figure JM-2, McCalpin, Appendix A).

A second hypothesis is that sackungen are shallow surface manifestations of toppling and flexural slip along discontinuities that dip steeply into a mountain mass, but which do not penetrate to any great depth (Jahns, 1964; Beck, 1968). Bovis (1982) termed this process *flexural toppling* and cited model studies (Barton, 1971) and studies in quarries (Goodman and Bray, 1976) as support for this non-penetrative mode of extensional deformation. Given that different (shallow and deep) mechanisms for sackung formation exist, combinations or even a continuum of mechanisms also may exist.

Although many authors have concluded that sackungen result from slow rock creep (e.g., Chigira, 1992), linear sackungen resemble tectonic fault scarps (Bovis and Evans, 1995; Thompson et al., 1997). In addition, the formation or rejuvenation of sackungen-like landforms during historic earthquakes is well documented (Dramis and Sorriso-Valvo, 1983; Wallace, 1984; Cotton et al., 1990; Ponti and Wells, 1991; Nolan and Weber, 1992). Sackungen, therefore, may have formed by: (1) displacement on tectonic faults, (2) gravity failures caused by earthquake shaking, or (3) gravity failures unrelated to tectonic activity.

Regional, local, and feature-specific characteristics of sackungen are described in Table 2.4.

Table 2.4 Summary of characteristics: sackungen

Regional context	Sackungen
Tectonic	Not associated with any particular tectonic setting; may be more common in active tectonic environments due to close association with mountainous terrain.
Relation to tectonic stress/strain fields	Orientation primarily controlled by topography; may be inconsistent with contemporary tectonic stress and strain fields.

Table 2.4 (continued)

Geologic	Sackungen are observed in a variety of geologic terranes including sedimentary rocks, foliated metamorphic rocks and crystalline igneous rocks. Predominantly restricted to mountainous physiographic regions (e.g., Pacific Border, Rocky Mountains, and Basin and Range Provinces). Commonly associated with glaciated terrain due to oversteepened topography and shallow or exposed bedrock.
Geophysical	Sackungen are not characteristically associated with regional geophysical anomalies.
Seismologic	The sackung process is nonseismogenic, but some sackungen may originate as a result of strong ground shaking.

Local context	**Sackungen**
Topographic	Regions of high topographic relief; sackungen form high on ridge flanks and at ridge crests; commonly restricted to one side of, or adjacent to, ridge crests.
Geomorphic	Downhill-facing scarps, double-crested ridges, uphill-facing scarps with graben, and erosional notches on ridge axes.
Stratigraphic	Can occur in igneous, sedimentary or metamorphic rocks; unconsolidated fine-grained colluvial sediments may be deposited in sackungen troughs, and may be tilted adjacent to the fault plane. Deep-seated mechanisms may be accommodated by distributed shearing at depth.
Structural	Two hypotheses: (1) sackung have a well-defined slide plane that extends from a headscarp into a zone of rock creep (deep-seated feature); and (2) sackung are shallow surface manifestations of toppling and flexural slip along shallow, steeply dipping discontinuities. Both have asymmetric shear zones. Deep-seated mechanisms may be accommodated by distributed shearing at depth.
Geophysical	Sackungen are not characteristically associated with local geophysical anomalies.
Seismologic	The sackung process is nonseismogenic, but some sackungen may originate as a result of strong ground shaking.
Hydrologic	Sackungen are not restricted to particular local hydrologic settings; as a slope failure mechanism, anomalous groundwater conditions (e.g., shallow and/or pressurized) may contribute to sackung development.

Feature characteristics	**Sackungen**
Spatial characteristics	
Morphology	Four most common geomorphic features: downhill-facing scarps, uphill-facing scarps, double-crested ridges, and erosional notches on ridge axes.

Table 2.4 (continued)

Geometry	*Plan view -*	Short, high, discontinuous, arcuate scarps; commonly multiple and clustered.
	Cross section -	In documented trenches and natural exposures the shear zones are moderately steep to vertical (45-90°). Down-dip geometry at depth uncertain, possibly distributed shear if deep-seated.
	Aspect ratio -	Typical dimensions: scarp length 15-300 m; scarp height 1-9 m. Sackungen scarp length/height ratios ($\leq 10^2$) are significantly less than tectonic fault scarp length/height ratios ($\geq 10^4$).
Structure	Internal fault zone morphology is similar to tectonic faults, including a discrete pulverized zone with fault gouge and a phyllitic or brecciated zone (depending on host-rock lithology). In densely foliated rock, a phyllitic zone may form by microscopic slip along foliations. In massive rock, a brecciated zone may form through networks of tension fractures. In sparsely foliated rock, the brecciated zone may form by random crushing. Fault zones may have an asymmetric sharp upper contact and a transitional lower boundary.	
Scale	Documented scarp lengths < 300 m and heights <9 m. Generally on a scale of a few km long (<10 km).	
Sense of deformation	Extensional, with valley-side down and ridge-side down both common.	
Displacement per event	Aseismic displacement may be continuous; both secondary coseismic and strictly gravitational displacements may be large (> 2 m).	
Depth	Extent and geometry of sackungen at depth generally are not well known and both shallow and deep mechanisms are hypothesized. Mass rock creep may extend to depths of several hundred meters. The maximum depth of the deep-seated failure plane is limited by local relief; the minimum depth of failure surface associated with flexural slip along pre-existing discontinuities and toppling is unconstrained.	
Spatial associations	Sackungen may have spatial associations with local stress fields due to: (1) ice wedging, (2) gravity forces, (3) stored forces from prior loading (e.g., glaciation), (4) seismic shaking, and (5) tectonic displacement connected to deep-seated seismogenic faults.	
Hydrologic	No known, consistent association with hydrologic conditions. As a shallow slope failure process, may be influenced by anomalous groundwater conditions (e.g., shallow and/or pressurized).	
Temporal characteristics		
Rate of deformation	Aseismic, short-term slip rates of up to 10 mm/yr have been observed, long-term rates not well documented.	
Episodicity	Stratigraphic evidence suggests slow, continuous deformation is most common; less common presence of colluvial wedge deposits across some sackungen suggest episodic deformation.	

Table 2.4 (continued)

Duration of deformation	Duration characteristics not documented, although thick depositional sequences suggest that activity may span at least thousands of years.
Recurrent deformation	Patterns of recurrent deformation may be indicated by buried soils within sackung depressions.
Temporal associations	May be reactivated by strong ground shaking, and thus have ages similar to other nearby coseismic features (e.g., landslides, fault scarps). May be associated with periods of fluvial and/or glacial incision and erosion that result in oversteepend slopes.
Investigative techniques	Regional and local geologic and geomorphic mapping. Drilling and geophysical investigations may provide data on presence and geometry of basal slide plane. Small-scale geophysical methods also may be appropriate for delineating deformation in Holocene deposits. Trenching useful in determining origin of feature (tectonic or gravitational) and rate of deformation. Geodetic surveying can be useful in determining mode (episodic or continuous) and rate of slip, as well as the presence and character of displacement in the "toe" region.
References	*See* summary and references cited in McCalpin (Appendix A).

2.6.3 Diapirs and Large-Scale Gravity Glide Structures Above Salt or Shale

2.6.3.1 Salt/Shale Diapirism Structures

Rock salt and overpressured shale are weak, mobile rocks that can deform and flow under low tectonic or gravitational stress levels. Due to their very weak nature and ability to flow and spread, both types of rocks commonly act as basal lubricating layers above which overlying rocks can glide, spread, extend, and be translated over large distances. **Diapirs** and **growth faults** typically result from deformation of salt or shale. Geomorphic features common to areas of active salt tectonism include **sinkholes, fissures, arcuate fault scarps, slump structures**, and **chaotic tilting** of adjacent bedrock blocks. Rising salt diapirs commonly produce positive topography in the arid west (*See* Figures PH-2, PH-3, and PH-4, Huntoon, Appendix A). **Caprocks** comprised of dissolution residues such as gypsum, limestone, shale and other clastics may protrude above the land surface. Surface and near-surface faults associated with diapiric salt structures may exhibit structural and geomorphic characteristics common to faults having tectonic origins. In contrast, humid environments produce collapse or subsidence features *(See* Section 2.6.6*)*, such as sinkholes or circular lakes, under which the cores of salt diapirs are dissolving. Regional- and local-scale subsidence also may occur in arid environments due to salt migration and dissolution (Kirkham and Streufert, 1996).

Salt or shale-cored **anticlines** and **domes** above actively rising **diapirs** commonly exhibit youthful, high-gradient, steep-walled drainages that appear to be out of character with other drainages in the surroundings (Huntoon, Appendix A). In extreme cases, hill slopes will be at the angle of repose and barren of climax vegetation because of high rates or erosion. Sedimentation in graben occurs where through-flowing drainage has been captured by faulting, or where graben subsidence exceeds channel incision rates. Careful geologic mapping that emphasizes the planimetric distribution and subdivision of Quaternary deposits, delineation of caprock exposures, and trends and closures along fold axes is required to identify and characterize areas of deformation related to salt/shale diapirs (Coleman, 1983).

Diapirs and growth faults are observed in two types of tectonic settings (*See* Figure BV-1, Vendeville, Appendix A):

(1) Diapirs commonly occur within divergent continental rifts or basins affected by thick-skinned extension where salt was deposited before episodes of rifting began (Figure BV-2, Vendeville, Appendix A). Seismic data, experimental results, and theoretical considerations suggest that the locations of salt structures and overlying faults are mostly independent of underlying basement faults (Vendeville et al., 1995).

(2) Diapirs and growth faults also occur in basins and on passive continental margins where the salt or shale layer was deposited after rifting ended (*See* Figure BV-1c and d, Vendeville, Appendix A). Deformation in these areas is thin-skinned, affecting only the salt/shale layer and overlying rocks. Examples include the Gulf of Mexico (from SW Texas to NE Florida) and the south Atlantic margins. In these basins, early thin-skinned extension and salt flow are shown by stratigraphic thinning and thickening, and onlaps and truncations of seismic reflectors across growth faults and above rising salt diapirs (*See* Figure PH-1, Huntoon, Appendix A). In this type of setting, salt and shale structures commonly are caused by gravity gliding and/or gravity spreading (Ramberg, 1981) (*See* Figure BV-3, Vendeville, Appendix A).

Regional, local, and feature-specific characteristics of salt flowage structures (anticlines, salt domes, salt diapirs, and associated faults) are summarized in Table 2.5.

Table 2.5 Summary of characteristics: salt/shale diapirism structures

Regional context	Salt/shale diapirism structures
Tectonic	Associated with both thick-skinned (e.g., rift) and thin-skinned (e.g., margin) extensional tectonic settings (modern and ancient).
Relation to tectonic stress/strain fields	Geometry and structure may be inconsistent with contemporary stress/strain fields in tectonically inactive regions. In regions of active tectonic extension, active structures likely are consistent with contemporary stress/strain fields. Salt-related structures related to previous extension that have been reactivated in compressional environments also may be consistent with contemporary tectonic stress/strain fields.
Geologic	Thick source layers of salt and/or shale capped by sedimentary deposits.
Geophysical	Gravity anomalies associated with buried shale and/or salt bodies; strong seismic reflectors associated with salt.
Seismologic	Little or no direct association with moderate to high levels of seismicity. Possible indirect association with moderate to high levels of regional seismicity in active tectonic environments.

Table 2.5 (continued)

Local context	Salt/shale diapirism structures
Topographic	Structures are associated with low to moderate relief; produces positive topography in arid west. Regional collapse (km^2- to 100 km^2-scale) may result from evaporite flowage and dissolution.
Geomorphic	Linear, stepped extensional scarps (early stage); circular or oval scarp complexes in plan view (middle stage); emergent ,circular or oval, dissected salt or shale body (advanced stage). Anomalous, high-gradient steep-walled drainages; anomalies in stream gradient that suggest localized folding and anticlinally folded fluvial terraces that dip away from axial stream. Salt dissolution features (hummocky ground, closed depressions, etc.) may develop in areas of salt diapirism with shallow groundwater (See Section 2.6.6.2).
Stratigraphic	Presence of salt or shale is required; stratigraphic thickening and thinning of salt/shale deposits. Localized sedimentation within collapsed blocks due to disruption by faulting, or where graben subsidence exceeds channel incision rates.
Structural	Structures are associated with three piercement modes for salt diapirs: (I) Reactive - stairstep normal faulting above buried wall; (II) Active - anticlinally folded bedrock, radial or subparallel normal faults, monoclines and thrust faults; (III) Passive - cap rock or glacier of salt, contact drag (highly contorted and folded bedrock) in wall rock of diapir. The spacing of normal faults associated with diapirs typically is less than the spacing associated with tectonic normal faults. Diapir-related deformation is restricted to rocks that overlie the source layer, resulting in more localized rollover folds and footwall uplifts than those associated with basement-involved normal faults.
Geophysical	Gravity anomaly for salt bodies; strong seismic reflectors associated with salt; truncation of stratigraphic reflectors.
Seismologic	Low levels of seismicity due to failure of rock within and above salt structures, usually extensional.
Hydrologic	No hydrologic characteristics for salt diapirs. Shale mobility typically associated with high fluid pressure.

Feature characteristics	Salt/shale diapirism structures	
Morphology	Linear, ridge-like scarps and landforms during early stage; circular ridges and depressions (associated with ring faults) during later stage.	
Geometry	*Plan view -*	Linear, ridge-like or circular.
	Cross section -	Diapirs form upwelling “tear” shape.
Structure	Graben above diapir crest. Ring faults in suprasalt rocks; faults dip toward diapir; older faults die out upward; younger faults closer to diapir; fault traces perpendicular to maximum extension.	
Scale	Diapirs typically less than 2 km in diameter, associated faults may be smaller. Features related to salt migration may extend for several km to 10s of km.	
Sense of deformation	Predominantly extensional.	

Table 2.5 (continued)

Depth	Faults restricted to rocks that overlie source deposits of salt/shale and generally involve only the uppermost section (< 7 km) of the crust. Typically are less than 3-5 km deep.
Spatial associations	Associated with mobile salt or shale; diapirs generally rise from the cores of salt anticlines.
Hydrologic	Elevated fluid pressures associated with shale/mud diapirism.
Temporal characteristics	
Rate of deformation	Strongly controlled by rate of extension (i.e., rapid extension promotes rapid growth of diapirism).
Episodicity	Episodic and continuous.
Duration of deformation	Regional deformation due to salt diapirism can span tens to hundreds of million years. Shale diapirism determined by duration of high fluid pressures.
Recurrent deformation	Recurrent or continuous deformation evidenced by continuous sedimentary thinning. A thickness increase in roof strata indicates diapiric rise ceased or slowed down significantly.
Temporal associations	
Investigative techniques	Regional geologic mapping to discern thickening and thinning relationships; seismic reflection and gravity studies, combined with subsurface drilling data.
References	See summary and references cited in Huntoon (Appendix A) and Vendeville (Appendix A); Kirkham and Streufert (1996).

2.6.3.2 Gravity Glide and Gravity Spread Features

Gravity gliding is the transport or translation of rock masses along a weak detachment plane or décollement, at the kilometer scale, composed of salt and/or shale. Gravity gliding requires that both the base and the topographic surface of the gliding unit be inclined. Gravity gliding typically results in normal faults in the upper part of the slope separated from folds or thrusts in the lower part of the slope by long, undeformed gliding blocks. Near-surface glide structures that result from sliding of the rocks above the salt are not commonly observed. Huntoon (1982) identified the causative stresses and classified the resultant strain features associated with one well documented salt-floored gravity glide structure, the Needles fault zone in Canyonlands, Utah (*See* Figure PH-9, Huntoon, Appendix A).

Gravity glide structures above salt or shale exhibit geomorphic features similar to that of low-gradient landslides. Deformation at the leading edge of the plate is compressional in style and may result in the formation of anticlines and reverse faults. In contrast, extension predominates at the trailing edge and within the moving plate. A detachment fault typically underlies the plate, and the plate itself commonly is deformed internally by high-angle normal faults, producing a horst-graben complex with faults that strike perpendicular to the direction of sliding. Active glide plates may exhibit open fissures and sinkholes along buried extended fissures. Older fissures commonly are filled with clastic sediments.

Gravity spreading is the vertical thinning and horizontal widening of rock masses and is regarded as the main trigger for growth faulting and diapirism along passive margins. Gravity spreading tends to lower topographic highs and thicken topographic lows by (1) extending the continental shelf and the upper slope, (2) translating the middle slope seaward, and (3) shortening sediments or salt on and in front of the lower slope (Figure BV-5, Vendeville, Appendix A). Gravity spreading may occur in the absence of a basal slope, but requires a surface slope, and space on the lower slope or in front of the lower slope to allow for lateral spreading. This space can be provided either by the absence of sediments in front of the spreading sediment wedge, formation of folds or thrusts in thin sediments on or in front of the lower slope, or by shortening of pre-existing diapirs or salt/shale massifs on or in front of the lower slope.

In tectonic settings subjected to gravity spreading or gliding, normal faulting of the overburden promotes diapir rise regardless of the thickness and density of the sedimentary rocks. Vendeville (Appendix A) provides a description of the evolutionary stages of development and the different types of faulting associated with extension-induced diapirs.

Where sedimentation is fast enough to compensate for thinning by normal faulting, extension is accommodated by **growth faulting** rather than diapirism. The best examples of large growth faults detaching on a thin décollement layer are along the Texas portion of the northern Gulf Coast (*See* Figures BV-4 and BV-13, Vendeville, Appendix A) and have been described in detail in Worral and Snelson (1989) and Nelson (1991). Growth faults in these environments initiate at or near the shelf break, and their formation progresses seaward in response to seaward sediment progradation (*See* Figure 14, Vendeville, Appendix A). In vertical section, these faults have a listric profile, dip seaward, and their dip decreases with depth down to a décollement layer of salt or shale generally 7 to 8 km deep. Commonly, these growth faults have accumulated large amount of slip (thousands) of meters. Strata deposited during faulting are **growth strata** that thicken toward the fault plane and are anticlinally deformed by rollover folding .

Where the décollement layer is thicker (*See* Figure BV-15, Vendeville, Appendix A), as along the Louisiana portion of the Gulf Coast, growth faulting and vertical rise of salt domes combine to produce more complex structures. Fault traces commonly are arcuate (concave seaward), a few kilometers long, and terminate against diapirs. Fault orientation and vergence can vary greatly, especially where the prograding clastic sediments loading the salt were not deposited as a linear, continuous front along the entire shelf edge, but more locally in circular or hemi-circular delta lobes. In the latter case a complex set of radial and concentric faults may form.

The limited published literature on seismicity associated with salt structures suggest that they lack moderate and high levels of seismicity, and generally are not capable of producing significant ground motion (i.e., moderate to large magnitude earthquakes). Based on an analysis of microseismicity data collected in the Paradox Basin, Utah, Wong et al. (1987) note the following characteristics of seismicity in regions underlain by salt: (1) failures in rocks above active salt structures usually are caused by extension, and generally have low rates and magnitude of seismicity; (2) hypocentral location data reveal a pattern of earthquake foci that are predominantly restricted to the salt and overlying rocks; and (3) deep-seated crustal tectonism is indicated if the earthquakes occur at depths below the salt.

Seismicity of the central Gulf of Mexico, a region that includes many actively growing salt-related structures, is very low (Frolich, 1982). Although a moderate magnitude (magnitude 5.0) earthquake occurred in this region in 1978, the hypocentral depth of this event was about 15 km, deeper than the crustal region currently involved in salt tectonics and gravity gliding (i.e., 7-8 km) (Frolich, 1982). Recent analysis of the 1984 sequence of small (M 1.9 to 3.2) earthquakes in the Carbondale, Colorado area suggests that the earthquake-generating deformation may have been associated with slip within or along the interface of an evaporitic layer at depths of about 3-6 km (Goter and Presgrave, 1986; Goter et al., 1988).

In areas of regional compression, such as the Salt Ranges and the Potwar Plateau in Pakistan, the Zagros Mountains in Iran, or the Kuylab area in Tadjikistan, studies provide conflicting interpretations of the association between salt structures and seismicity. Yeats and Lillie (1991) and Seeber et al. (1981) suggested that movement along an evaporite detachment was unlikely to generate significant seismicity and attributed earthquakes observed in the region to folding. Leith and Simpson (1986), however, reported moderate earthquakes (M>5) clustered around emergent and buried diapirs, and attributed the earthquakes to active rise of the salt diapirs. It seems more likely that this seismicity reflects folding and/or thrust faulting of the sedimentary rocks above and between the salt diapirs, because these rocks are able to store (and release) greater amounts of elastic strain energy.

Regional, local, and feature-specific characteristics gravity glide features are summarized in Table 2.6.

Table 2.6 Summary of characteristics: gravity glide and spread structures

Regional context	Gravity glide and spread structures
Tectonic	Associated with active and inactive extensional tectonic settings.
Relation to tectonic stress/strain fields	No consistent relationship to regional tectonic stress/strain fields.
Geologic	Regional detachment layer or décollement composed of salt or overpressured shale body overlain by sedimentary deposits.
Geophysical	Regional gravity anomalies associated with large salt bodies and basin formation.
Seismologic	Little or no association with moderate to high levels of seismicity. Possible association with microseismicity.

Local context	Gravity glide and spread structures
Topographic	Low to moderate relief; gravity glides require the topographic surface and base of the gliding unit be inclined in the same direction; gravity spreads – vertical thinning and horizontal widening of rock masses driven by topographic slope; tend to lower topographic highs and raise topographic lows.

Table 2.6 (continued)

Geomorphic	Geomorphic features similar to that of low-gradient landslides: headscarps, closed depressions, graben, hummocky terrain due to extensional faulting and fissuring at the trailing edge and within the glide block; non-cylindrical anticlinal uplifts, overhanging scarps due to contractional faulting and folding at the leading edge.
Stratigraphic	Salt or shale layer within basal décollement is required. Décollement is overlain by sedimentary deposits, commonly including conspicuous "growth strata." Internal closed depressions and older fissures filled with clastic sediments.
Structural	Normal faults are present in upper slope and thrust faults and folds in lower slope. These faults sole into a basal detachment. Internal blocks may undergo horizontal and vertical rotation accommodated by internal faults.
Geophysical	Abundant discordant geologic relationships (e.g., shallow faults, internal block rotations, detachment) are discernible on seismic reflection data. Local gravity anomalies are associated with salt bodies.
Seismologic	Not associated with moderate or high levels of seismicity. Seismicity may be associated with faulting within the salt/shale detachment and overlying rocks.
Hydrologic	Shale-gravity glide structures are controlled by abnormally high fluid pressures within the shale formation. Salt-gravity glide structures are independent of hydrologic setting.

Feature characteristics	**Gravity glide and spread structures**	
Spatial characteristics		
Morphology	Similar to low-gradient landslides: fault scarps, troughs, and graben occur at the upper, trailing edge; creep indicators (e.g., oversteepened slopes, tilted trees, hummocky landscape) in internal region and leading edge; youthful, positive topography and anticlinal uplifts occur at leading edge. Very active glides will exhibit open, emergent fissures, and sinkholes along buried fissures.	
Geometry	*Plan view -*	Arcuate, commonly en echelon, and normal faults concave to glide direction located upslope, rotated blocks within center of glide, reverse faults and folds (convex to glide direction) downslope.
	Cross section -	Series of rotated fault blocks and overlying half-graben bounded by listric normal faults (growth faults), commonly imbricate systems, that merge at depth into salt or shale décollement layer.
Structure	Ductile shear zone (décollement) occurs at base of glide; internal blocks commonly are back rotated on listric faults that dip in the direction of plate movement. Half-graben commonly are present in hangingwall; normal faults are present in upper part of slope, and folds or thrusts are located in lower part of slope. Internal minor faults are typical within the glide block.	
Scale	<10 km lateral ; <8 km vertical; faults may terminate against diapirs.	

Table 2.6 (continued)

Sense of deformation	Downslope movement is due to gravitational force: extension occurs in the upper region block; contraction in lower parts of the mass; possible strike-slip displacement occurs along margins of slide mass.
Depth	Basal detachment faults (décollement) are limited to depths of less than 8 km.
Spatial associations	Salt or shale diapirs.
Hydrologic	Elevated fluid pressure closely associated with shale detachments.
Temporal characteristics	
Rate of Deformation	Rates of deformation are variable. Moderate to potentially high strain rates (mm/yr) are common; total displacements of as much as thousands of meters are common.
Episodicity	Variable, episodic and continuous (closely related to local geologic conditions such as sediment loading, slope, strain rate, etc.). Shale-gravity slides may respond to fluctuating fluid pressure. Salt flows continuously, but strain in overlying rock may be episodic.
Duration of deformation	Process may occur continuously for time periods of up to several million years as recorded by growth strata; duration of shale gravity glides are dependent upon duration of high fluid pressure, which may be transient.
Recurrent deformation	Evidence of recurrent deformation commonly is recorded in sedimentary record.
Temporal associations	Movement of shale-gravity glides is linked to periods when shale is overpressured (e.g., climatic, sediment loading, seismic activity) but also can be continuous.
Investigative techniques	Regional and local geologic mapping; shallow and deep seismic reflection profiling, combined with drilling, and local geologic "paleoseismic" investigations of fault features.
References	See summary and references cited in Huntoon (Appendix A) and Vendeville (Appendix A).

2.6.4 Faults in Areas Affected by Glaciation

Identifying tectonic faulting in glaciated regions is complicated by the many glacial and periglacial processes that create geomorphic features that mimic surface fault rupture (e.g., ice-push or basal drag, freeze-thaw heave, ice-cave collapse, cryoturbation, and meltwater expulsion). In addition, glacio-isostacy produces transient loading stresses that may result in differential uplift and reactivation of tectonic faults. Recent research regarding postglacial rebound and its capacity for generating earthquakes is summarized in a compilation volume by Gregersen and Basham (1989). This publication addresses several topics including: (1) the influence of tectonic versus postglacial isostatic recovery stresses in earthquake generation; (2) spatial variations and the nature of regional and more localized differential uplift; (3) triggered slip and reactivation of preexisting structures; and (4) the implications of temporal and spatial patterns of postglacial deformation for assessments of neotectonic activity and seismic hazard assessment.

The various types of faulting observed in glaciated or recently glaciated regions may be classified into the following categories:

(1) **Glacio-isostatic** (commonly referred to as **postglacial)** faulting that occurs in regions of ice cover in response to changes in the glacial load, either as a result of deglaciation (crustal unloading) or glacial advance (crustal loading).

(2) **Glaciotectonic** faulting used to denote any deformation resulting from ice movement (ice push or ice drag).

(3) **Periglacial** faulting resulting from freeze-thaw processes, and

(4) **Shallow stress relief** faulting resulting in the formation of **pop-up** structures. Shallow stress-relief faulting due to glacial loading/unloading will be associated spatially and temporally with the extent and timing of glaciers. Shallow stress-relief faulting also can result from non-glacial unloading mechanisms, both natural (e.g., erosion) and cultural (e.g., quarrying). Both mechanisms result from the relief of shallow stress in the regional compressive stress regime.

Fenton (Appendix A) provides a summary of the characteristics of these types of faults, and gives criteria for distinguishing between glacio-isostatic (postglacial) faults and nontectonic deformation features in glaciated regions. He notes that differentiating between faulting produced by transient loading stresses (i.e., postglacial faulting) and tectonic faulting that results from long-term regional lithospheric stresses is more problematic. Regional, local, and feature-specific characteristics of glacio-isostatic (postglacial) faults and glaciotectonic features are provided in Tables 2.7 and 2.8, respectively.

2.6.4.1 Glacio-isostatic Faulting

Postglacial faults of varying scales have been described at numerous locations in glaciated regions of North America as well as in Fennoscandia and Scotland (see Fenton, Appendix A and references cited in Table 2.7). Small-displacement postglacial faults are thought to form in the brittle, jointed upper half-km crust in response to transient flexural stresses during deglaciation (Adams, 1989). Oliver and others (1970) favor a model of differential movement on planes of weakness resulting from expansion due to hydration or unloading. Postglacial faults generally are dip-slip (mainly thrust and reverse) and occur on pre-existing planes of weakness (e.g., bedding planes, cleavages, joints). Zones comprised of numerous small-throw, closely-spaced faults are common in steeply-dipping slaty rocks.

Large-displacement postglacial faults are recognized in Fennoscandia and Scotland. These predominantly steeply-dipping, reverse faults also reactivate pre-existing faults, fractures, and shear zones. Postglacial movement on the larger of these faults, which may extend for a distance of up to 200 km have produced scarps up to 15 m high. Movement on the larger of these reactivated faults may be seismogenic, as shown by the spatial and temporal association with seismically induced deformation including liquefaction and landsliding. Large-scale postglacial faulting and significant earthquakes at the time of deglaciation may occur in response to sudden release of stresses that accumulated deep in the crust during the period of ice loading. Limited exploratory trenching shows that postglacial fault scarps are the result of single events, at or immediately following the time of most recent deglaciation. In some cases, the reactivated structure is a ductile shear zone that shows little or no evidence of previous brittle deformation since initiation of the contemporary stress regime.

Table 2.7 Summary of characteristics: glacio-isostatic (postglacial) faults (modified from Fenton, Appendix A)

Regional context	Glacio-isostatic (postglacial) faults
Tectonic	Faults are commonly observed in intraplate/craton environments and passive margin or failed rift settings undergoing rapid glacio-isostatic uplift. To date, no examples have been reported from active tectonic settings.
Geologic	Faults are most commonly observed in Precambrian shield and early Paleozoic fold belts (intraplate settings). Faults are not confined to any particular rock type.
Relation to Stress/Strain Fields	Small-displacement faulting is thought to represent the response of the brittle, jointed upper half kilometer of crust to transient flexural stresses during deglaciation. Near the surface, the magnitude of the flexural stress may have been similar to the regional, continent-wide stresses, and being radial to the ice front would sometimes reinforce and sometimes oppose and override the contemporary stress field. Thus, some small-displacement faults may not show a relation to the regional stress field. Large-displacement faults more commonly are oriented orthogonal to the direction of maximum horizontal compression.
Geophysical	Faults occur in recently deglaciated regions that are undergoing glacio-isostatic rebound. Faults are typically associated with thickened continental crust with low heat flow (cratons) or moderately attenuated crust with low to moderate heat flow (passive margins and failed rifts).
Seismological	Regions of presently low to moderate seismic activity.
Local context	**Glacio-isostatic (postglacial) faults**
Topographic	Commonly low topographic relief, although examples have been observed in regions of alpine relief.
Geomorphic	Evidence of recent glaciation (glacially polished/scoured bedrock, striated pavements, glacial constructional landforms, etc.).
Stratigraphic	Not restricted to any particular local stratigraphic setting. Small faults in eastern Canada and northeastern U.S., however, are better expressed in fissile slatey rock types with steep to vertical cleavage planes.
Structural	All reported examples of postglacial faults have reactivated pre-existing faults, fractures, and shear zones. Displacement is predominantly reverse, although strike-slip components of displacement have been noted along lateral ramps and minor accommodation splays. Minor normal faulting has been reported from the hangingwall blocks of larger faults in Fennoscandia.
Geophysical	Associated with regions of high bending strain as a result of glacio-isostatic rebound.

Table 2.7, continued

Seismologic	Spatial and temporal association with widespread liquefaction and low-angle landsliding observed in some areas indicate that postglacial faulting was associated with large magnitude seismicity. Contemporary seismicity is low.
Hydrologic	Evidence from trench exposures across the Lansjärv fault in northern Sweden indicates expulsion of groundwater during faulting. Subglacial fluid recharge may be important in triggering postglacial faulting. Models for triggering postglacial faulting indicate that crustal fluid overpressuring may be required to initiate faulting.

Feature characteristics	**Glacio-isostatic (postglacial) faults**
Spatial characteristics	
Morphology	Steeply-dipping, reverse fault scarps, generally in bedrock or shallow glacial and postglacial deposits. Closely-spaced scarps generally show consistent sense and direction of displacement. Small normal faults sometimes present in the hangingwalls of larger faults. Small strike-slip components observed on small accommodation structures. Scarps range between a few millimeters to 15 m high. Fault lengths range from a few tens of meters, up to 200 km.
Geometry	*Plan View* – Variable; continuous, roughly linear to slightly arcuate reverse fault scarps to highly angular, irregular scarps utilizing pre-existing fracture sets. *Cross Section* – Subsurface geometry of these faults are unknown, although sparse seismicity data does suggest that they may become shallower with depth. Aspect ratio (H:V) < 1:11,000, generally less than similar tectonic reverse faults having the same displacement.
Scale	Fault lengths vary from 1 km to 100+ km. Smaller faults (<10 m length in slatey units in eastern North America. Scarp heights (offsets) vary between 1 mm and 15 m.
Sense of Deformation	Reverse faulting, with rare secondary normal faulting. Small components of strike-slip faulting also have been reported on small accommodation splays.
Depth	Larger faults extend through the entire seismogenic crust (up to 40 km in Fennoscandia). Smaller faults (in slatey units) may extend no more than a few meters.
Spatial Associations	Faulting confined within regions of former ice cover. Postglacial faulting also occurs along pre-existing faults, fractures, or shear zones.
Hydrologic	Occasional association with groundwater expulsion.
Temporal characteristics	
Rate of Deformation	Faults are characterized by high rates of deformation, followed by lengthy periods of inactivity.

Table 2.7, continued

Episodicity	Faults commonly are associated with single ruptures.
Duration of Deformation	Fault movement may occur during a short term around the time of deglaciation.
Recurrence Deformation	There is no conclusive evidence for formation of faults as a result of multiple events.
Temporal Associations	Many faults formed soon after deglaciation. Faults may be triggered by deglaciation and consequent glacio-isostatic rebound.
Investigative Techniques	Geologic and geomorphic mapping; exploratory trenching; age-dating; structural analysis; drilling.
References	Adams (1981, 1989); Bäckblom and Stanfors (1989); Fenton (1991, 1994); Grant (1990); Lagerbäck (1988, 1992); Lundqvist and Lagerbäck (1976); Mather (1843); Mörner (1978, 1981); Muir Wood (1989, 1993); Olesen (1988); Oliver et al. (1970); Ringrose (1987, 1989a, 1989b); Sissons and Cornish (1982a, 1982b); Tanner (1930).

2.6.4.2 Glaciotectonic Deformation

Glaciotectonics, is the process by which glacier-ice movement or loading results in the deformation and dislocation of underlying substrate, including shallow bedrock as well as unconsolidated deposits. Glaciotectonic structures and subglacially deformed material have been studied to provide information on subglacial processes, glacial dynamics, the long-term behaviour of large ice masses, and the local ice-movement direction for surficial mapping, indicator tracing, and glacial geology studies (e.g., Hicock and Dreimanis, 1984; Benn and Evans, 1996; Croot, 1988). Benn and Evans (1996) provide a general classification of subglacially deformed materials based on sediment properties and their relationship to styles of subglacial stain and drainage conditions. They use geologic evidence to infer former glacier-bed conditions.

Studies of glaciotectonic structures produced by ice push and basal drag have been used to develop concepts and models of thin-skinned tectonics (see discussions in Croot, 1987, and Banham, 1988). In these models a front-end, compressional regime dominated by folds and listric thrusts, is structurally and mechanically linked with a trailing end, extensional regime characterized by listric normal faults. Commonly, a shallow, basal thrust directly links the proglacial (contractional) and subglacial (extensional) features formed. The majority of glaciotectonic deformation is the result of ice-push at the leading edge of advancing ice fronts (Croot, 1988) and this deformation is more widely recognized and reported in the literature. However, recent studies also have addressed the style and nature of the extensional features formed as a result of ice push and e.g., Hicock and Dreimanis, 1984; Croot, 1988; Zelčs and Dreimanis, 1997).

Hydrostatic jacking of bedrock blocks resulting from fluid overpressuring during and after deglaciation (Zotikov, 1986) is a related process that also may generate substantial vertical displacements (Talbot, 1990; Fenton, 1992). Such dilational delamination of the shallow crust would likely result in vertical faulting with complementary compressional and extensional faulting at the leading and trailing edges of the block (Adams et al., 1993).

Table 2.8 Summary of characteristics: glaciotectonic faulting

Regional context	Glaciotectonic faulting
Tectonic	Faults are not restricted to a specific tectonic setting; faults may occur in interplate or intraplate tectonic settings.
Relation to tectonic stress/strain fields	There is no consistent relationship to contemporary stress/strain fields. Orientation of faults may be controlled by direction of ice movement and local subglacial strain conditions.
Geologic	Faults occur in areas subject to ice-cap or continental scale glaciation (e.g., Fennoscandia and Canada), and areas subject to valley or alpine glaciation.
Geophysical	Faults are not restricted to any particular geophysical environment.
Seismologic	Faults result from nonseismogenic processes, and there is no correlation to present seismicity.
Local context	**Glaciotectonic faulting**
Topographic	Larger faults are present in low relief areas, smaller, less extensive faults in alpine areas.
Geomorphic	Associated with evidence of recent glaciation (glacially polished/scoured bedrock, striated pavements, rôche moutonées, flutes, etc.), glacial landforms (eskers, drumlins, terminal moraines), and associated extrusion of meltwater (e.g., meltwater channels).
Stratigraphic	Associated with glacial deposits (tills) and evidence of water-saturated sediments. May be modified by glaciofluvial processes or buried by outwash deposits.
Structural	Heterogeneous deformation; broad deformation zones; commonly displays both compressional and extensional deformation in close proximity or within the same outcrop. Compressional glaciotectonic deformation commonly involves the formation of (disharmonic) folds. Lack of consistency in the sense and amount of throw along strike. Associated soft-sediment deformation (e.g., diapirs, convolute bedding, ball and pillow structures, syn-diapiric faults, boudinaged structures) and clastic dikes comprising fluid-escape structures, glaciogenic injections, and fluidal and viscous hydraulic expulsions.
Geophysical	No association to specific geophysical anomalies.
Seismologic	No association to present or past seismicity.
Hydrologic	Influenced by basal and englacial hydrologic conditions during glaciation.

Table 2.8, continued

Fault characteristics	Glaciotectonic faulting
Spatial characteristics	
Morphology	Faults commonly are expressed as broad, discontinuous zones of deformation, comprised of multiple fault planes, with both normal and reverse displacements; faults may form geomorphic scarps; glacial plucking at base of the ice sheet/glacier along preexisting faults, fractures, or bedding/schistosity planes can result in bedrock steps.
Geometry	*Plan View* – Broad, discontinuous zones of deformation; parallel to former ice front. *Cross section* – Variable showing wide range of dips from near vertical to subhorizontal, commonly along a single fault plane; shallow décollement, typically at stratigraphic contact (e.g., till-bedrock contact). Faults typically do not propagate up from bedrock into overlying unconsolidated deposits as discrete planes; deformation in unconsolidated sediments is more commonly expressed as folding.
Scale	Generally few tens of mm to few tens of m. Can be up to km size if ice-rafted blocks are included.
Sense of deformation	Dominant style of deformation resulting from ice-push at the leading edge of advancing ice fronts is compressional, producing curviplanar thrust faults, reverse faults, shear planes, fractures, and minor folds. Extensional features also form as a result of bending moment stresses or stress relaxation. Basal drag also results in compressional deformation at the leading edge, but along trailing edges extensional deformation is commonly displayed. Subglacially-formed extensional faults show down-ice dip direction.
Depth	Shallow phenomenon; faulting tends to shallow with depth, sometimes dying out in shallow décollement.
Spatial associations	Faulting confined within regions of former ice cover.
Temporal characteristics	Generally occurs during glacial advance.
Rate of deformation	Generally slow continuous strain rate. (Examples from Svalbard show formation in one winter season).
Episodicity	Single or repeated events that may result from pulses or surges of glacial movement.
Duration of deformation	Can only occur while ice is present and actively advancing (less likely during ice retreat).
Recurrent deformation	Generally not likely at specific locality, but could result from repeated glacial advances.
Temporal associations	Occurs contemporaneously with active advance of ice sheets.

Table 2.8, continued

Investigative techniques	Local geologic and geomorphic mapping at a scale sufficient to identify regional patterns of ice movement; kinematic and mechanical analysis of observed deformation features, geochronologic studies to constrain timing of movement with respect to recent glaciation; geophysical and subsurface investigations (drilling, trenching) to evaluate geometry and extent of faulting at depth.
References	Schroeder et al. (1986); Croot (1987, 1988 and references therein); Hicock and Dreimanis (1984); Broster (1991); Adams et al., (1993); Benn and Evans (1996); Zelčs and Dreimanis (1997); Dreimanis and Rappol (1997).

2.6.4.3 Ice-Contact Deformation

Other processes associated with mass wasting of glaciers, such as collapse of sediments deposited on top of remnant ice masses (Shilts and Farrell, 1982) or from sediment instability associated with rapid sediment accumulation during deglaciation (Shilts, 1984) also can produce surface faults or sediment slumping in lake basins, that could be confused with tectonic faults or seismically triggered deformation.

2.6.4.4 Periglacial Faulting

In areas with mean annual temperatures ≤ 5°C (i.e., arctic climate) the upper meter or so of sediment undergoes repeated freezing and thawing on an annual basis. Fluid overpressuring associated with these processes result in bedrock heave and frost bursting (Michaud et al., 1989). Thermal expansion and contraction commonly results in a surface expression known as polygon ground or sand/ice wedge casts (Hamilton et al., 1983; Svensson, 1989). In outcrop, frost-wedge features typically are confined vertically to the uppermost 1-2 m, the zone of seasonal temperature change. Features of uncertain origin that are similar in appearance, but more continuous have been observed in the Canadian Arctic (Dyke et al., 1991). Lineaments that resemble ice wedge troughs extend across raised marine sediments and continue onto till where they are expressed as lines of mudboils (although the precise cause of this relationship is uncertain).

2.6.4.5 Shallow Stress Release Features

Shallow stress-relief features, pop-ups, develop at the surface in response to high horizontal compressive stress. In Southern Ontario, Canada, pop-ups occur primarily in flat-lying limestone and dolomite (White and Russell, 1982). They appear to be generated by decoupling of the topmost few meters of flat-lying sedimentary rock along some bedding plane or shale layer (a décollement) in response to ambient horizontal stresses of 5-10 MPa found in surface bedrock throughout the region. Wallach and others (1992) state that they have been identified in a broad belt that extends in a generally southwest direction from southeastern Canada into the east-central United States. They generally trend northwest, or perpendicular to the prevailing orientation of the greatest principal stress in eastern North America, however they display other orientations as well. Rutty and Cruden (1993) suggest that pop-up structures formed in a region of southern Ontario in order to relieve high near-surface tectonic stresses that were accumulated beneath the Laurentide Ice Sheet. They note that ice removal is similar to overburden removal in the case of quarry floor buckles and is predicted by Roorda and others (1982) to promote pop-up formation. They also note that pop-up structures appear to have nucleated on favourably-oriented pre-existing joints in bedrock.

Typically, pop-up features are elongate, up-arched, brittle folds up to several meters high and hundreds of meters to more than a kilometer long. Fakundiny and others (1978) provide descriptions and diagrammatic cross-sections of various types of pop-ups they have observed in western New York and Lake Ontario. Their observations indicate that most pop-up structures are elongate, anticlinal folds that have no corresponding synclinal neighbors. Their axes are commonly coincident with a prominent joint set or multiplex joint and the crest of the fold is usually along one of these joint planes. Unbroken versions also exist. Stress relief pop-ups, when observed in section, die out into shallow décollements (Fenton, 1994; Fakundiny and others, 1978; Wilson, 1902).

2.6.5 Stress Release Faults

The scale of faulting triggered by unloading spans a wide range from small-scale displacements resulting from mining or quarrying operations, to large-scale faulting related to crustal-scale stresses caused by formation and melting of continental ice sheets. The characteristics of large-scale crustal movement in areas influenced by glacial loading and unloading are described in Section 2.6.4. A summary of the characteristics of faulting resulting from smaller scale unloading, chiefly resulting from human engineering activities, is provided in this section. Simpson (1986) provides an overview of the influence of engineering works on local stress and a discussion of the three main types of triggered seismicity: injection, mining, and reservoir loading. Summaries of triggered seismicity related to human engineering activities also are presented in Judd (1974) and Milne (1976).

Seismicity may be triggered by changes in elastic stress caused by the removal of large masses of rock in mining and quarrying operations (Cook, 1976). Studies show that two types of triggered earthquakes have resulted from the excavation of rock from the crust: those in the local working area of deep underground mines, and those triggered at depth beneath shallow mines and large surface quarries. Rockbursts resulting from shear failure, especially in deep mines, have many characteristics in common with natural earthquakes (Cook, 1976; McGarr et al., 1975; Spottiswoode and McGarr, 1975). These researchers report events of up to magnitude 5 at depths of nearly 3 km and the initiation of new shear failures within a few hundred meters of the active mine face in deep mines. Removal of the surface load in quarrying operations results in a decrease in vertical stress, and thus the greatest effects are in regions of thrust faulting, where the vertical stress is the minimum (e.g., Pomeroy et al., 1976; Yerkes et al., 1983; Sylvester and Heinemann, 1996).

The characteristics of stress release structures and seismicity resulting from deep underground mining and large-scale surface quarrying are exemplified by a number of case studies, as described below. Because a key characteristic of such deformation is its spatial and temporal association to human engineering activities, rather than a specific geologic or tectonic environment, we do not attempt to describe the regional or local context of such features in a summary table.

- An instance of faulting triggered by unloading comes from a lignite mine near Peissenberg (southern Germany) on the northern edge of the Bavarian Alps (Illies and Greiner, 1978). In a west-trending adit located at a depth of 1120 m, a southward directed low-angle reverse fault, striking parallel with the tunnel, ruptured into the void, producing a meter-high 'fault-scarp' in the tunnel wall.

- Another example comes from the western Transverse Ranges in central California, where in 1981 the removal of 44 m of diatomite overburden at a large quarry triggered a bedding-parallel right-reverse oblique fault rupture (Yerkes et al., 1983; Sylvester and Heinemann, 1996). The fault plane dips between 39-59°, and created a scarp 575 m long with maximum vertical and horizontal displacements of 23 cm and 9 cm, respectively. A small (M_L=2.5) earthquake was associated with the production of this scarp (Yerkes et al., 1983). The net reduction in load in the rupture area was estimated as 5 bars. A similar size earthquake (M_D =2.3) and associated surface rupture (reverse slip; 670 m long, 25 cm maximum vertical separation) occurred in 1995 in the same vicinity, subsequent to removal of about 40 m of overburden (Sylvester and Heineman, 1996). Sylvester and Heinemann (1996) note that part of the faulting occurred along bentonite interbeds that have slipped previously during tectonic flexural folding and bending-moment bedding plane slip long before quarrying commenced. They agree with McGarr's (1991) speculation that 40 m thickness constitutes the critical threshold value that results in a decrease in normal stress of 3-4 bars, together with a 2-bar increase of shear stress with consequent slip failure on bedding surfaces that dip. Leveling data suggest that an undetermined part of the stored elastic strain also may have been released aseismically by isostatic uplift.

- Seismicity and displacements also have been associated with the excavation of a 45-m-wide tunnel at 420 m depth at the Atomic Energy of Canada Ltd's Underground Research Laboratory (Collins, 1994). The horizontal tunnel is situated in a relatively homogeneous, unfractured, granitic rock mass, and was excavated in 1 meter increments using a rock breaking technique. Displacements, notches similar to borehole breakout, formed to as much as 450 mm in the roof and floor of this tunnel. Much of the induced seismicity associated with the tunnel coincided spatially with the deformed regions. Events with the largest seismic moment and static stress drop typically occurred immediately after an excavation increment, and seismicity continued to occur in the particular excavation volume after a number of subsequent tunnel advances. The majority of events on the sides of the tunnel occurred at the tunnel face that was under shear failure conditions and evidenced a tensile component of failure (Collins, 1994).

- A review of mining-induced, stope-scale microseismicity data collected by the U.S. Bureau of Mines in a hard rock mine showed both spatial and temporal fractal characteristics (Kranz, 1994). The results of this study suggest that the physical processes responsible for rock mass relaxation following a sudden stress change are the same, regardless of progenitor or location. Differences were found in attributes that measure spatial extent, event rate, and cumulative event numbers and energy. Kranz (1994) concluded that the magnitude of the response, not the interactions of forces and local structure, distinguishes the aftershock sequences associated with rock bursts.

Surface fault rupture resulting from earthquakes that are assumed to be reservoir-triggered has occurred in a number of regions. As noted in a recent review of reservoir-triggered seismicity

(RTS) (U.S. Committee on Large Dams, 1997), almost all the largest RTS have occurred in areas where there is active Quaternary faulting. Packer and others (1981) noted that of 10 M>5 earthquakes that are generally accepted as triggered events, eight have occurred in areas of well documented active faulting, and the other two probably were located in such areas. Others note that RTS are probably more numerous in areas of normal and strike-slip faulting as opposed to areas of thrust faulting (Simpson, 1986; Packet et al., 1981; Jacob et al., 1979). From these studies it appears that there may be spatial and temporal associations with reservoir impoundment, but that the faults that ruptured exhibit evidence of prior tectonic surface rupture with similar characteristics in the existing tectonic environment.

2.6.6 Subsidence and Collapse Structures

Subsidence refers to the gradual downward settling or sinking of the Earth's surface due to consolidation or removal of underlying rock and is not restricted to rate, magnitude or area involved. Subsidence may be caused by either natural or man-made processes including the extraction of fluids (e.g., petroleum, water); solution of soluble subsurface deposits such as salt; thawing of permafrost; oxidation of organic-rich soils such as peat; desiccation and shrinkage of expansive soils; settlement of soils due to seismic shaking, hydrocompaction, and/or consolidation induced by increased loading; subsurface erosion via piping in poorly consolidated sediments; and tectonic movements. In contrast, **collapse** refers to the rapid and/or total failure of an overlying body by the force of gravity, due to the removal of an underlying support, as in the collapse of mines or natural caverns. Collapse may sometimes be preceded by subsidence. Holzer (1984a) and Borchers (1998) provide a number of review papers related to human-induced land subsidence and collapse.

Subsidence and collapse structures are produced by distinctly different processes. For instance, subsidence structures are produced by fluid (oil and groundwater) withdrawal, whereas collapse structures are produced by the dissolution of substratum through circulating groundwater as well as from underground mining. Preexisting fault planes and discontinuities may facilitate the formation of both of these structures. The processes and geomorphology of subsidence and collapse structures are discussed below, respectively.

2.6.6.1 Subsidence Structures

Fluid withdrawal is the most common and widespread cause of human-induced regional subsidence in the United States (Griggs and Gilchrist, 1983). Subsidence structures formed by the removal of subsurface mineral deposits, such as coal, are also prevalent, especially in the southeastern United States. Subsidence structures produced by these processes can range in size laterally from a few meters to tens of kilometers.

Excessive fluid (groundwater, hydrocarbon and geothermal) withdrawal may cause a reduction in pore pressure and consolidation of surrounding porous deposits that can result in subsidence (Martin and Serdengecti, 1984; Holzer, 1984b). The decrease in pore pressure increases grain-to-grain contact, which leads to sediment compaction and ground subsidence. The principal ground deformation includes local- or regional-scale subsidence basins that may be accompanied by earth fissures and faults limited to the area of subsidence (Holzer, 1984b).

Groundwater Extraction

Excessive groundwater withdrawal produces three types of deformation (Holzer, 1984b): (1) **subsidence basins**; (2) **aseismic ground fissures**; and (3) **aseismic faults**. The occurrence and characteristics of these deformation features are discussed separately below.

(1) **Subsidence basins** attributed to historic groundwater extraction typically experience subsidence ranging from 1 to 8.5 m, with the maximum recorded subsidence of approximately 10 m recorded at several locations in California, Japan, and Mexico (Holzer, 1984b; Johnson, 1995). Subsidence basins range in diameter from less than ten km to tens of kilometers. Subsidence typically is greater near the center of the groundwater basin, where the water-bearing sediments are the thickest, but these features may not be coincident for highly asymmetric basins. The development of a subsidence "bowl" causes the ground near the margins of the basin to extend and displace toward the basin center. This effect may be enhanced by hydraulic seepage forces where groundwater migrates toward extraction wells within the basin. As the groundwater elevation declines, an increase in the effective confining stress occurs due to a loss of buoyancy in the soils above the water table (or in the case of a confined aquifer, due to depressurization of the aquifer strata; e.g., see Lofgren, 1968). This activity will induce consolidation of compressible soil layers, permanently reducing their volume and causing the basin to subside. Subsidence usually occurs very slowly and can be traced to the initiation of groundwater extraction (Griggs and Gilchrist, 1983; Holzer, 1984b).

(2) **Aseismic ground fissures** induced by groundwater extraction have been documented for nearly twenty locations throughout the western and southwestern United States (e.g., Lofgren, 1968; Schumann and Poland, 1970; Holzer, 1976; Guacci, 1978; Holzer et al., 1979; Holzer and Pampeyan, 1981; Jachens and Holzer, 1982; Shlemon and Davis, 1992) and other locations worldwide (e.g., El Baruni, 1994). Observational data suggest that fissures typically occur at or near the points of maximum convex-upward curvature in a subsidence profile . The production of ground fissures is strongly influenced by localized differential compaction and horizontal contractions induced by capillary stresses in the zone above a declining water table (Holzer, 1984b). Fissures tend to develop where extensional strains are concentrated due to the behavioral properties of the sedimentary units involved. For example, fissuring may be localized where geologic conditions have juxtaposed different thicknesses of compressible layers, promoting differential subsidence. Such conditions might exist near the margins of a basin, across buried faults, bedrock highs, and/or buried stream channels--wherever alluvial deposition has created adjacent stratigraphic sections with markedly different compressibilities. In addition, aseismic fissuring can be localized by boundaries of different piezometric head or aquifer confinement conditions within the basin.

Field evidence suggests that the total strain accommodated by extensional separation across such fissures typically is small, typically in the range of about 0.05 to 0.2 percent (Holzer and Pampeyan, 1981; Guacci, 1978).

Two mechanisms have been suggested to explain the development of fissures and to predict where they are most likely to occur. One model likens the originally horizontal surface of a subsidence basin to a bending beam or plate. The extensional forces are greatest, and hence fissuring is most likely to occur where the convex upward radius of curvature is the greatest. This model predicts that fissures will occur along the shoulder of the subsidence basin where horizontal extension is greatest. For a homogeneous beam or plate, the theory predicts that fissures will initiate at the surface and propagate downward. In reality however, varying states of stress and strain occur throughout the areal and vertical extent of an alluvial basin, primarily due to nonuniformity of stratigraphic and groundwater conditions. However, the typically non-homogenous and anisotropic nature of basin geology may cause aseismic fissures to initiate at depth and subsequently propagate toward the surface or offset from the location of maximum convex-up curvature. The bending beam analogy should not be confused with the "draping" effect that is often observed in groundwater basins in the southwestern United States (Helm, 1992). This draping describes the landscape response to differential vertical movement at depth caused by stratigraphic heterogeneities or buried geologic structures, not migration/ extraction of groundwater (e.g., Fischer and Chervin, 1992).

An alternative model is the hydraulic seepage forces theory, which also may be used to explain and predict the occurrence of ground fissures. According to this model, the principal driving force on an aquifer (the saturated assemblage of solid particles in a groundwater basin) that is being pumped is the difference between the driving force on the solids and water together and the driving force on the water relative to the solids (Helm, 1992). The latter is the seepage force and is directly measurable as the gradient of hydraulic head. When a groundwater pump is turned on, both the aquifer and the water move radially inward toward the extraction well. A bulk hydraulic force allows such movement to occur in outlying areas near the margin of a groundwater basin even before drawdown occurs locally. Fissures are predicted to occur where geologic structures and heterogeneities impede this motion, particularly where preexisting planes or points of weakness allow a subsurface crack to be generated. Once formed, such a crack may propagate upward to the surface and evolve into a ground fissure.

This model is consistent with field observations that subsidence-related cracks migrate upwards from depth and express themselves at the surface as a final step in their development (e.g., Bell, 1981; Helm, 1992). Fissures are observed to occur not only as predicted along the shoulders of a subsidence bowl where the curvature of vertical movement is convex upward, but also beyond the outer perimeter of the bowl where no drawdown or subsidence has been observed. Helm (1992) also reports that fissures have been found near the center of a subsidence bowl where the curvature of subsidence is concave upward.

Ground fissures commonly are hundreds of meters long, although many are longer and fissures as long as 3.5 km have been observed (Holzer, 1984b). Cross-sectional form and dimension can vary along individual fissures and also between different fissures, with differences of size and shape primarily due to modifications by erosion, deposition, and the amount of tensile strain relieved by each fissure. Therefore, ground fissures may be expressed as hairline cracks (when they have not been modified by erosion), potholes, or large, collapsed, trench-like depressions (Holzer, 1984b; Jachens and Holzer, 1982). Fissure zones may consist of isolated, straight fissures, or as parallel fissures within a few zones. Open depths generally are less than 2 to

3 m, but have been documented up to 6.7 m (Holzer, 1984b). Measurements of the volume of void space and width of fractures formed suggest that the cracks extend to depths of tens to hundreds of meters (Holzer, 1984b).

Detailed logging of open fissures and associated fractures provides a general picture of the subsurface morphology of the open fissures. The subsurface expression of the Pixley fissure (Guacci, 1978) and fissures formed near Chino, (Stewart et al., 1998) are similar, consisting of a broad fissure within the upper 2.5 to 3 m of the ground, and an essentially vertical, narrow constant-width infilled fissure at greater depths. The Pixley fissure was reported to extend at least 17 m below ground surface, and fissure width deeper than about 3 m varied between approximately 13 and 50 mm. Trench exposures across the features in the Chino area (Geomatrix, 1994; Stewart et al., 1998) showed a complex zone of fractures, voids, and filled fissures that is 0.6 to 0.9 m wide in the upper approximately 3 m and < 12 to 37 mm wide at depth. In some areas the voids had propagated to the surface and formed troughs or grabens; in other areas partially or completely filled voids are present below the surface. Voids within the upper 3 m typically are filled with blocks of soil, loose soil, vegetation, and particles of soil rounded by subsurface water flow. Below this depth the fissures are filled with fine-grained, highly dispersive (i.e., readily transported) materials such as silt and very fine sand. Due to void space created by the significant depth of the fissures and the erodibility of the soils in which they form, the fissure near the surface is frequently enlarged into the dramatic surface manifestation (i.e., wide, deep gullies) by surface water infiltration and erosion.

The initiation and growth of fissures generally correlates with the timing and magnitude of human-induced fluid-level changes. Although evidence for recurrent movement is common, single event features also are observed. Fissure systems may remain intermittently active for more than 30 years. Strong ground motion or extreme climatic events may trigger events.

(3) **Faults related to groundwater extraction.** Excessive groundwater extraction can produce high-angle normal faults that creep aseismically at rates ranging from 4 to 60 mm/yr (Holzer, 1984b). The initiation and growth of these faults generally corresponds with the timing of groundwater extraction. They may begin as new faults or propagate along pre-existing faults and discontinuities. The locations of groundwater extraction faults in the Houston-Galveston area commonly coincide with the locations of pre-existing faults mapped on deeply buried stratigraphic horizons.

Fault scarps associated with groundwater extraction faults have similar geomorphic expression to tectonically produced scarps along normal faults. Typically, scarps are about 1 km long and less than 0.5 m high (Holzer, 1984b). The highest modern scarp reported is 1.12 m high in the Houston-Galveston area (Reid, 1973). Fault-associated deformation has been observed on both the hanging and footwall blocks; deformation measured by geodetic surveys extends more than 200 m from the scarp (Holzer, 1984b).

The localization of groundwater-related deformation along preexisting faults may complicate assessment of the activity and seismogenic potential of these faults. Unlike tectonic faults, however, neither abrupt movement nor seismicity are associated with the scarps produced by groundwater withdrawal. In addition, the faults commonly are temporally associated with periods of groundwater extraction, and spatially distributed within the zone of groundwater extraction. For example, in Houston, Texas, Reid (1973, cited in Holzer and Gabrysch, 1987) observed seasonal variations in rates of fault creep at two sites and concluded that natural processes and groundwater extraction were acting in unison and generating fault movement. Pampeyan et al. (1988) demonstrated that failures along pre-existing Holocene fault scarps in Fremont Valley, California, began to form in the 1960s, coincident with groundwater pumping.

Hydrocarbon Extraction

Ground subsidence also occurs as a result of the extraction of hydrocarbon reserves. Reviews of oil and gas field subsidence in the United States are provided by Poland and Davis (1969) and Martin and Serdengecti (1984). The latter paper reviews the fundamentals of subsidence related to oil and gas extraction, the mechanical behavior of reservoir rock, methods of estimating maximum subsidence and subsidence prediction and control.

Most oil and gas fields experience only small amounts of surface subsidence, typically less than 1 m (Martin and Serdengecti, 1984). The greatest subsidence related to oil extraction occurred at the Wilmington oil field in Long Beach, California between 1936 and 1966. This oil field recorded a maximum subsidence of 9 meters and a horizontal displacement of up to 3.7 m, yielding an average rate of subsidence of 300 mm/yr between 1936 and 1966 (Segall, 1989; Griggs and Gilchrist, 1983).

Important factors in understanding oil field subsidence are the reservoir fluid pressure (pore pressure), thickness (overburden), geometry (effective shear stress), and the mechanical rock properties (frictional sliding, strength) of the surrounding and overlying formations (Martin and Serdengecti, 1984). The ground subsidence is associated with in-situ rock failure by grain fracturing, crushing and rearranging (cataclasis), which facilitate the compaction of the reservoir rock (Martin and Serdengecti, 1984). As the oil is extracted, the effective lithostatic stress conditions (vertical and horizontal) of the porous rock is increased by the reduction in pore pressure. The reduction in pore pressure (reservoir fluid pressure), therefore, imposes additional stresses (horizontal and vertical) on the surrounding rock. Depending on the strength of the surrounding rock, and the degree to which the additional stresses are uniform (i.e., lithostatic or differential [capable of inducing shear]), the overlying and surrounding rocks may be deformed (Martin and Serdengecti, 1984). Most hydrocarbon reservoirs that have undergone compaction are characterized by bedded sand and shale (Martin and Serdengecti, 1984). The finer-grained deposits initially resist deformation, because they usually have undergone significant reorientation and compaction during and after burial. Therefore, much of the compaction is accommodated by fracturing, crushing and rearrangement of the coarse-grained components of a deposit.

Subsidence features resulting from oil extraction are similar to those produced by groundwater extraction. For instance, ground fissures and faults may be newly created and/or occupy pre-existing discontinuities. The most common type of faulting induced from excessive oil extraction are reverse and normal faults, with reverse faults concentrated near the center of the depression, and normal faults along the margin of the subsidence basin as ring faults (Figure 2.13).

Faults produced by the removal of hydrocarbons also can be associated with shallow earthquakes of

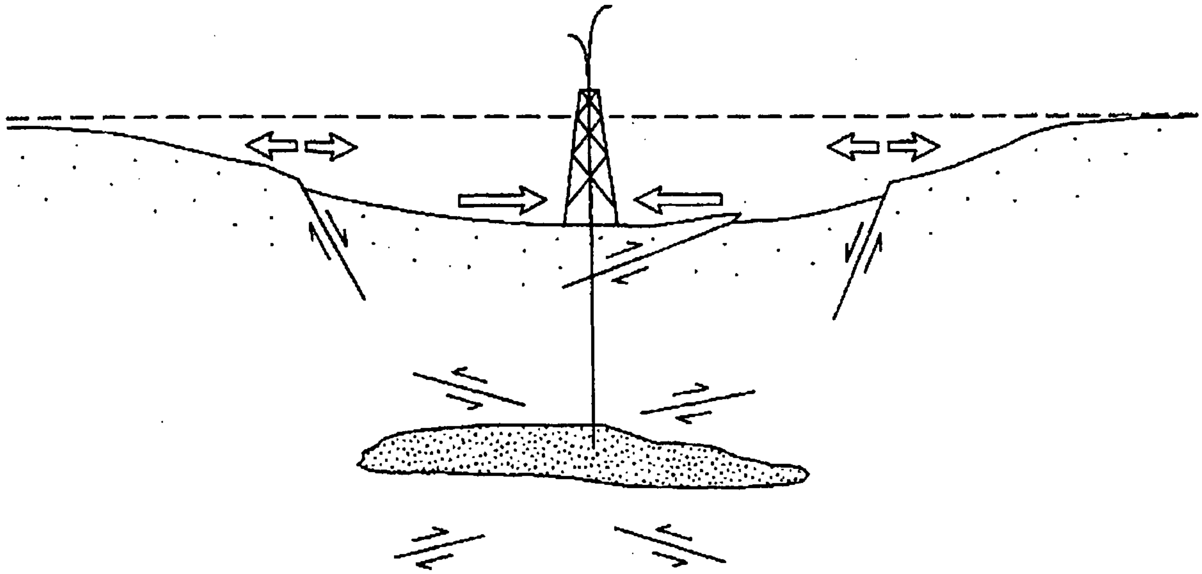


Figure 2.13 Schematic cross section summarizing surface deformation and faulting associated with fluid withdrawal. Open arrows indicate horizontal strain at Earth's surface. Normal faults develop on flanks of field, at a Goose Creek field. Reverse faulting occurred above reservoirs at Wilmington, Buena Vista Hills, Pau, and below reservoir at Strachan field (from Segall, 1989). Reprinted with permission of the publisher, the Geological Society of America, Boulder, Colorado, USA (Copyright © 1989 Geological Society of America).

< M5 (Segall, 1989). The seismicity pattern is concentrated within the boundaries of the production field. For example, Grasso and Feignier (1989, cited in Segall, 1989) demonstrate that 95% of the earthquakes that were attributed to development of a gas reservoir in the western Pyrenees had their epicenters fall within the boundaries of the production field.

Regional, local, and feature-specific characteristics of fluid-extraction features are summarized in Table 2.9.

Table 2.9 Summary of characteristics: fluid extraction-related subsidence structures

Regional context	Fluid extraction-related subsidence structures
Tectonic	May occur in intraplate or interplate regions.
Relation to tectonic stress/strain fields	No consistent relationship; may trigger seismicity with focal mechanisms influenced by regional stress field.
Geologic	Unconsolidated sedimentary aquifer or hydrocarbon reservoir systems.
Geophysical	May be associated with regional geophysical anomalies related to fluid reservoirs.
Seismologic	May occur in seismically active or quiescent regional seismic environments.

Table 2.9 continued

Local context	Fluid extraction-related subsidence structures
Topographic	Features are most common and prominent at or near the points of maximum convex-upward curvature in subsidence profile (e.g., margins of subsiding basin).
Geomorphic	Deformation is highly localized, generally extensional; features include normal fault scarps, closed depressions; fissures; linear troughs.
Stratigraphic	Presence of unconsolidated and compactible sediments at depth; features are often concentrated where geologic conditions have juxtaposed different thicknesses of compressible layers (e.g., near margins of basin; across buried faults, bedrock highs, and buried stream channels; at alluvial facies changes).
Structural	Linear and arcuate fissures and/or tension fractures; arcuate, high-angle, normal faults with down-to-the-basin net displacement. Not all historical faults coincide with preexisting faults. However, the coincidence of numerous historical faults with preexisting faults and the natural grouping of historical faults into structural provinces that coincide with the provinces defined by pre-existing faults indicates that most surface faults connect to pre-existing faults. Faults associated with oil extraction include reverse faults concentrated near the center of the depression as well as normal faults along the margin of the subsidence basin as ring faults.
Geophysical	Presence and downdip extent of subsidence-related structures are not easily imaged on geophysical data due to low total strains.
Seismologic	Fault scarps produced by groundwater withdrawal and associated subsidence are aseismic; seismicity (<M5) has been associated with subsidence and surface faulting in some hydrocarbon fields, possible association with events larger than M5.
Hydrologic	Significant groundwater withdrawal from major aquifers or hydrocarbon and/or groundwater extraction from hydrocarbon reservoir regions. Spatial and temporal associations with historic fluid withdrawal; may be evidence locally for artesian conditions; near boundaries of different piezometric head or aquifer confinement conditions within the basin.

Feature characteristics	Fluid extraction-related subsidence structures
Spatial characteristics	
Morphology	Linear zones of fissures, sinkholes, and/or graben; topographic scarps (heights as much as 1 m and lengths as much as 16 km); extension cracks and fissures; fissures, which range in length from tens of meters to kilometers, typically open only a few centimeters by displacement., erosional enlargement creates gullies 1 to 2 m deep and wide, maximum depths and widths of 5 to 6 m. Fissures may form complex polygonal patterns.

Table 2.9 continued

Geometry	*Plan view -*	Laterally discontinuous; individual lengths as much as 2 km, total length of zones as much as 16 km; arcuate to linear.
	Cross section -	Upward divergence (widening due to erosion) of the fissure within the upper 2.5 to 3 m of the ground surface; vertical, relatively constant width (13 to 50 mm) infilled fractures to observed depths of 17 m.
	Aspect ratio (H:V) -	Generally low to very low.
Scale	Variable - related to extent of reservoir experiencing drawdown; may also be controlled by location of "weaker" regions within overlying strata.	
Sense of deformation	Historical subsidence; extensional displacement toward basin center; contractional deformation near basin center overlying hydrocarbon reserves.	
Depth	Limited by depth to reservoir or depth of zone stressed by water-level decline; depths as large as 25 m have been documented, but probably extend farther.	
Spatial associations	Concentrated at the zone of maximum curvature of subsidence in a geologically homogeneous basin or at buried structural or stratigraphic discontinuities (e.g., faults and shallow bedrock highs) within the basin.	
Hydrologic	There may be evidence for groundwater seepage forces, or natural groundwater drainage from a perched aquifer.	
Temporal characteristics		
Rate of deformation	Slow strain rate; scarps generally grow by aseismic creep at rates of approximately 4 to 60 mm/yr; fissure systems may remain intermittently active for more than 30 years.	
Episodicity	May reflect climatic change (e.g., groundwater levels) or may coincide with pumping cycles (e.g., mining activities, increased groundwater withdrawal during drier periods).	
Duration of deformation	Post-fluid withdrawal activities; related to nature and timing of subsidence.	
Recurrent deformation	Commonly there is evidence of recurrent movement, but may be a result of a single event.	
Temporal associations	Development correlates with timing and magnitude of fluid-level changes (induced or natural); may be triggered by strong ground motions or extreme climatic events.	
Investigative techniques	Analysis of leveling data to define amount and locus of subsidence; review of temporal and spatial evidence for relation between ground failure and declining fluid levels; geomorphic/geologic mapping; exploratory trenching; drilling and analysis of available subsurface data to identify structural and/or stratigraphic inhomogeneities within basin.	

Table 2.9 continued

References	Borchers (1998); Griggs and Gilchrist (1983); Holzer (1984a, b; 1989); Martin and Serdengecti (1984); Poland and Davis (1969); Reid (1973); Segall (1989); Yerkes and Castle (1969)

Hydrocompaction

Hydrocompaction is a process by which dry, unconsolidated, porous sediments lose their strength and collapse spontaneously after wetting. Hydrocompaction is restricted to dry, semiarid and arid areas, and usually occurs in wind-blown loess or alluvial deposits (*See* Lofgren, 1969 and Propkopovich, 1984 and references cited therein). Although the wetting may result from natural processes, it usually occurs as a direct result of human activity involving some form of water application, such as irrigation, construction of dams and canals, urbanization, or disposal of industrial wastewater (Prokopovich, 1984). Previously wetted sediments or sediments below the groundwater table are not susceptible to hydrocompaction.

The subsidence and collapse of both surface and subsurface soils as a result of hydrocompaction may result in the development of sinkholes, closed depressions, ground cracks, and slumps. Hydrocompaction features that developed in clayey piedmont alluvium in the California San Joaquin Valley included numerous, commonly round, 100-200 m diameter, 1-3 m deep sinkholes (Prokopovich, 1984). The thickness of deposits susceptible to hydrocompaction ranges from 1 to 2 meters to 50 meters or more, depending on maximum level of groundwater achieved in a basin, and the original thickness of the deposit (Prokopovich, 1984; Shaw and Johnpeer, 1985). The variability of soil conditions even within small areas commonly results in the spotty or uneven distribution of surface deformation due to hydrocompaction (Johnpeer et al., 1985 a, b). Other key factors that controlled the size, shape and severity of hydrocompaction sinkholes in the San Joaquin Valley are the amounts of irrigation water applied, rate and mode of application, and shape of wetted bodies (Prokopovich, 1984).

It is expected that surface deformation resulting from hydrocompaction would not be confused with tectonic deformation for a number of reasons, including: (1) the close spatial and temporal associations to wetting events, (2) the limited distribution of the features in the vicinity of the sources of water, and (3) the general lack of evidence for repeated events over a significant period of time.

2.6.6.2 Collapse Structures

Collapse structures are caused by both human-induced processes such as underground mining and naturally occurring *in-situ* processes such as dissolution of evaporites. Collapse structures related to underground mining typically occur in regions formerly mined for lignite (Gray and Bruhn, 1984) although other mining activities also may cause ground collapse. Dissolution of highly soluble material such as evaporites (salt and gypsum), carbonates (limestone and marble), and dolomite produce a characteristic geomorphology called karst. Karst geomorphology is characterized by closed depressions, caves, collapse features and the diversion of drainages underground (Easterbrook, 1993). The mechanics of the collapse processes are well described in proceedings from several multidisciplinary conferences that have been held on sinkholes and karst hydrogeology (e.g., Beck, 1984; 1989; 1993; Beck and Wilson, 1987). The collapse processes and their related structures are discussed below.

Underground Mining

For the United States, the most severe cases of mining-related collapse are recorded in the Appalachian region and the states of Pennsylvania, West Virginia, Kentucky, Ohio, and Illinois (Gray and Bruhn, 1984), where significant quantities of coal have been removed. Subsurface coal mining techniques in the United States use the "room and pillar" method, which purposely leaves behind thick pillars for support of the overburden (lithostatic load) formerly maintained by the excavated ore (Gray and Bruhn, 1984). When the strength of the remaining rock is exceeded, the overlying rock warps inward, with the maximum warping occurring near the cavity's center. The surrounding rock may undergo a gradual loss of support that is slowly transferred to the ground surface through subsidence processes that eventually may progress towards collapse (Griggs and Gilchrist, 1983).

Underground mining-related collapse features are characterized by circular depressions bordered by normal faults and tension cracks, as well as evidence of historical mining. These faults are restricted to the formations above the mine material. Linear depressions also may form above collapsed tunnels and adits. Seismicity associated with mining collapse/subsidence occurs as rock bursts that generally are less than M5.

Regional, local, and feature-specific characteristics of mining-collapse features are summarized in Table 2.10.

Table 2.10 Summary of characteristics: mining collapse structures

Regional context	**Mining collapse structures**
Tectonic	Interplate and intraplate tectonic settings.
Relation to tectonic stress/strain fields	No particular relation with contemporary tectonic stress/strain fields. Seismicity triggered by changes in elastic stress by removal of large rock masses.
Geologic	Restricted to regions of active or former underground mining of ore deposits (commonly coal).
Geophysical	No association with a particular geophysical setting.
Seismologic	No association with a particular seismologic setting.
Local context	**Mining collapse structures**
Topographic	No association with a particular topographic setting; may occur in areas of low to moderate topographic relief.
Geomorphic	Circular depressions, closed basins bordered by extensional scarps and fissures; linear surface depressions also may form above collapsed tunnels or adits.
Stratigraphic	Commonly associated with subsurface coal-bearing strata.
Structural	Depressions bordered by shallow normal faults and tensional fissures that are restricted to formations above mined material.

Table 2.10 continued

Geophysical	Seismic reflection may delineate extent of underground mining (rooms and tunnels); geodetic surveying across mining regions may identify regions undergoing gradual subsidence.
Seismologic	May generate microseismicity of <M5 related to shallow "rock bursts" triggered by removal of large rock masses (lithostatic load) and release of residual stress.
Hydrologic	Underground mining in areas of low relief may cause flooding or dewatering of an aquifer.

Feature characteristics	**Mining collapse structures**
Spatial characteristics	
Morphology	Circular and linear depressions, tension cracks, and compression bulges.
Geometry	*Plan view -* circular depressions and linear depressions, potentially interconnected.
Scale	From <0.1 km to 10 km (?).
Sense of deformation	Extensional (subsidence and collapse).
Depth	Restricted to depth of mining.
Spatial associations	Structures are associated with historical and active zones of mining.
Hydrologic	*See* above.
Temporal characteristics	
Rate of deformation	Continuous (subsidence) to episodic (collapse).
Episodicity	Deformation can occur episodically; usually longer-term.
Duration of deformation	Coincident with and following mining activities. Can occur long (100 yr) after cessation of underground mining (deformation continues as long as instabilities exist between overburden and underground voids).
Recurrent deformation	Subsidence and collapse may occur repeatedly.
Temporal associations	May occur contemporaneously with active mining.
Investigative techniques	Regional geologic mapping and historical land-use review; subsurface exploration via drilling and seismic reflection.
References	Borchers (1998); Cook, 1976; Dunrud, 1984; Gray and Bruhn, 1984; Simpson, 1986.

Salt Dissolution Collapse Structures

Dissolution collapse structures associated with salt bodies include **breccia pipes**, **collapse anticlines**, **subsidence basins**, and **sinkholes** (closed depressions). There are three major salt basins in the United States where these types of collapse structures are prevalent: Michigan-Appalachian Basin, Gulf Coast Basin and Permian Basin (in the southwestern to central United States; Ege, 1984). Salt dissolution collapse structures, which may result from natural or human-induced processes, are characterized by karst-like geomorphology and have been described in detail by Huntoon (*See* Appendix A) and Ege (1984).

The primary cause for the dissolution of underground salt is by circulating groundwater (Huntoon, Appendix A). Dissolution of the salt creates a progressively expanding cavity that reduces support for overlying rock. When the overburden exceeds the strength of the surrounding rock, the overlying rock begins to warp and/or spall off as blocks into an enlarging cavity (Ege, 1984). If sufficient underground space is available for the loosely packed rock debris to collect, the void can migrate to the surface and produce surface subsidence. This process may lead to progressive subsidence or catastrophic collapse of the ground surface, leaving behind a depression or sinkhole (Figure 2.14). Another mechanism that results in surface subsidence is subsurface erosion of susceptible layers (sandstone, silt, loess) overlying salt cavities (Ege, 1984). The sediments are eroded by groundwater and transported down subsidence-induced and natural cracks, or drill holes into the salt cavity. The voids formed in the higher eroded beds can then cause surface subsidence.

Salt dissolution collapse structures include: (1) small-scale sinkholes measuring a few meters across; (2) vertically extensive breccia pipes; (3) dissolution basins measuring a few kilometers across; and (4) large-scale valley collapses measuring tens of kilometers in length and a few kilometers in width (Sugiura and Kitcho, 1981, cited in Huntoon, Appendix A). The dissolution basins often are filled with recent sediments.

(1) **Sinkholes**
Sinkholes are conical surface depressions that range in size from a few meters to 1 km wide. They are nearly identical in morphology and in development to karst topography-related sinkholes (see following section on limestone dissolution structures).

(2) **Breccia Pipes**
In breccia pipes, the pipe cores are comprised of brecciated roof rocks that have fallen or subsided into the upward stoping structure (Huntoon, Appendix A). They occur within the overburden rocks of the larger salt collapse structures, and are identified by: 1) wall rocks folded downward toward the pipe and 2) ring fractures surrounding the breccia pipe (Huntoon, Appendix A). The pipes stope upward and maintain constant diameters. The downward displaced rocks are usually highly altered owing to circulating brines.

(3) **Collapsed salt anticlines**
In the case of collapsed salt anticlines, parallel rows of graben and normal faults are adjacent and parallel to the axis of the structure, with grabens commonly within the axis of the structure (*See* Figure PH-8, Huntoon, Appendix A). They generally occur in thin-bedded sedimentary deposits. The normal faults dip at high-angles toward the basin, extend only to the depth of dissolution, and structurally are indistinguishable from tectonic faults at the outcrop scale (Huntoon, Appendix A). The normal faults are aseismic, and commonly grow by aseismic creep. Chevron folding is a common structural style associated with and trending parallel to the collapsed anticline. In addition, conjugate shears often pervade the collapsed rock due to the vertically-oriented

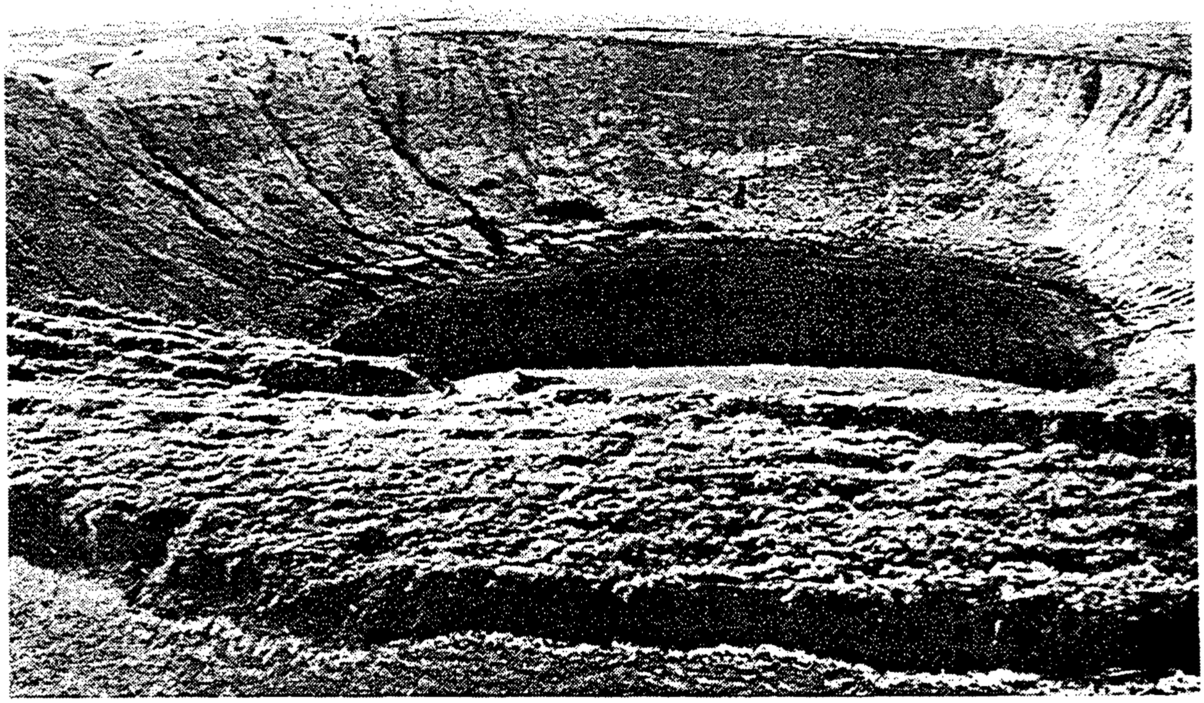


Figure 2.14 View of the Meade Salt Well, a sinkhole formed from dissolution of underlying salt beds, Meade County, Kansas, (from Ege, 1984). Reprinted with permission of the publisher, the Geological Society of America, Boulder, Colorado, USA. (Copyright © 1984 Geological Society of America).

causative maximum principal stresses (Huntoon, Appendix A). Active rising anticlines will exhibit youthful fluvial geomorphic features including high-angle steep-walled drainages and uplifted fluvial terraces and may be out of character with the drainages in the surrounding region.

(4) **Dissolution**
Subsidence due to dissolution of underlying salt may result in basin formation and collapse at a variety of scales. Kirkham and Streufert (1996) report that regional collapse due to evaporitic flowage and dissolution may be occurring over an area of about 500 km^2 in the Glenwood Springs-Carbondale-Gypsum area of Colorado. They note that vertical deformation across this collapse zone may exceed 1000 m and individual structures have up to 300 m or more of relief. A subsidence trough about 4 km long in Pleistocene outwash gravel filled with locally derived clastic sediments also is present in this area.

The key stratigraphic criterion for identifying salt-related deformational features is the presence of subsurface salt strata (Huntoon, Appendix A). The use of aerial photography and satellite imagery repeated over time often can highlight subtle vegetative and fracture features indicative of collapse structures, and provides a regional perspective of the geology and tectonics. In addition, geophysical investigative techniques such as seismic reflection, seismic refraction, electrical resistivity, induced polarization and low frequency electromagnetic waves (ground penetrating radar) are useful fordetecting collapse structures (Ege, 1984). These methods are most useful when supplemented with a subsurface sensing program through drilling, displacement transducers, thermal peizometers, flowmeters, dyes and "sonar calipers." Myers (1963, cited in Ege, 1984) describes the sonar caliper subsurface-logging technique, as an acoustic sounding device that is lowered through a borehole and extended into a solution cavity to determine the dimensions of the cavity.

Regional, local, and feature-specific characteristics of salt-dissolution structures are summarized in Table 2-11.

Table 2.11 Summary of characteristics: salt dissolution structures

Regional context	Salt dissolution structures
Tectonic	Generally not restricted by tectonic setting; common in former regional basin settings.
Relation to tectonic stress/strain fields	No consistent relationship to regional tectonic stress or strain fields; orientation of features may reflect stress field in which underlying salt flowage features developed (e.g., extensional basins, anticlines), or in which regional joint sets were formed that act as conduits for subsurface flow of dissolving fluids. Maximum principal stresses usually are vertical in regions experiencing salt dissolution.
Geologic	Structures present in former large evaporitic basins and in areas underlain by major bedded salt deposits.
Geophysical	Presence of underlying salt body may be indicated by low gravity anomaly; regional seismic reflection studies also useful for characterizing salt bodies.
Seismologic	No correlation with regional seismicity.
Hydrologic	Circulating groundwater is a primary cause of salt dissolution.
Local context	**Salt dissolution structures**
Topographic	Variable; observed in low- to high-relief terrain. Local topography characterized by pseudo-karst including closed depressions and lost drainages. Regional collapse basins (tens of km) may occur due to evaporitic flowage and dissolution.
Geomorphic	Closed basins up to several km wide; semi-circular subsidence and collapse basins from several tens to hundreds of meters in diameter; fissures, arcuate fault scarps, slump structures and chaotic tilting of adjacent bedrock blocks.
Stratigraphic	Dissolution is common in areas where there is an underlying regional salt deposit and an overlying, thin-bedded sedimentary cap; large blocks of chaotic collapse debris may occur over large areas.

Table 2.11 continued

Structural	Preponderance of high-angle, concentric normal faults dipping basinward; tension fissures and graben on crests of salt anticlines and domes; fault swarms varying from uniform orthogonal sets of faults to very complex fault patterns.
Geophysical	Local low-gravity anomaly indicates presence of salt body.
Seismologic	No correlation with seismicity; salt dissolution structures typically are aseismic; may have very low contemporary seismicity along normal faults, where earthquake foci are restricted to depths of salt and suprasalt strata.

Feature characteristics	**Salt dissolution structures**
Spatial characteristics	
Morphology	Salt dissolution commonly is revealed by features similar to those in karst areas, including closed topographic depressions that range in size up to a kilometer, and elongate subsidence basins several kilometers long.
Geometry	*Plan view -* Various; breccia pipes, small-scale sink holes (< 3 m), subsidence basins (> 10 km); parallel rows of graben and half-graben of varying sizes adjacent to collapse anticlines. *Cross section -* Dominated by high-angle normal faults dipping basinward; faults and fissures increase in density with depth and are restricted to depth of salt. Fissures and arcuate high-angle normal faults are dominant structures. Closed depressions and linear valleys may be aligned along pre-existing faults or fissures that act as conduits for fluid slow.
Scale	Variable: sinkholes can form in large numbers (40-50 sinkholes/km^2); up to 1 km wide; locally faulted synclinal sags range from a few hundred to a few thousand meters in length; individual collapse structures may have less than ten to hundreds of m of structural relief.
Sense of deformation	Extensional: normal faults and related fissures.
Depth	Faults are restricted to depth of salt body.
Spatial associations	May be coincident with human-induced subsidence produced from mining operations, oil and gas activities, or construction of dams, reservoirs, and highways over saline rock.
Hydrologic	Circulating groundwater is necessary for formation of salt dissolution structures.
Temporal characteristics	
Rate of deformation	Faults generally are aseismic with high strain rates (mm/yr).
Episodicity	Collapse or subsidence of overburden can be continuous or episodic. Most structures form continuously.

Table 2.11 continued

Duration of deformation	Depending on depth of salt, some salt structures can have an extensive, complex history of deformation (i.e., as long as 100 million years). Deformation processes dependent upon abundance of both groundwater or surface water and source rock. Duration of deformation may be limited by processes that remove debris from base of the collapse structure and allow additional collapse.
Recurrent deformation	Structures show evidence of recurrent deformation.
Temporal associations	Development likely correlates with timing of naturally occurring (e.g., climatic, base-level changes) or human-induced fluid level changes.
Investigative techniques	Regional and local geomorphic mapping; remote sensing; subsurface geophysical techniques, particularly seismic reflection and gravity; drilling and downhole instrumentation.
References	Huntoon (Appendix A); Huntoon and Richter (1979); Ege (1984), Benson (1977, 1979); Kirkham and Streufert (1996).

Limestone Dissolution Collapse Structures

Classic karst topography requires the presence of limestone, abundant water (surface and groundwater), and dissolved carbon dioxide (Ritter, 1978). The rapid intake and flow of groundwater through soluble rock differentiates karst topography from other types of terrain (Kastning and Kastning, 1994). For instance, aquifers in karst terrain are highly transmissive because of the underground framework of connected "pipelines" and caverns (Kastning and Kastning, 1994). The reaction between the groundwater and dissolved carbon dioxide produces carbonic acid, which provides the driving force for the dissolution processes. In addition, Corbel (1957, cited in Easterbrook, 1993) indicates that for well-developed karst topography, the mineralogy of the limestone requires about 90% $CaCO_3$, and for initiation, at least 60% $CaCO_3$. Karst features commonly form in humid regions, which provide an optimal combination of temperature and precipitation to facilitate dissolution processes (Easterbrook, 1993). Finally, secondary permeability by fractures and bedding plane partings is necessary to facilitate the circulation of groundwater and dissolution of the limestone. The existence of entrenched valleys below uplands underlain by soluble and well-jointed rocks favors the development of karst.

Karst topography is dominated by the presence of relatively small, shallow, conical, closed depressions termed sinkholes. Sinkholes form by one of two processes: (1) the downward solution of limestone from the surface (solution sinkhole) or (2) the collapse of the roof into a solution cavity from dissolution taking place beneath the ground surface (collapse sinkhole) . Sinkholes may be randomly spaced and usually form in large numbers, commonly 40 to 50 in a km^2 (Easterbrook, 1993). Active sinkholes are in hydrologic communication with underground streams. They vary in size from 10 to 100 m in diameter and 2 to 100 m in depth, although they can grow as large as 1 km in diameter and hundreds of meters deep. Sinkholes can be classified according to their dimensions (Easterbrook, 1993):

(1) **Bowl-shaped** sinkholes are categorized as solution sinkholes. They are very shallow relative to

the diameter of the sinkhole and have slopes of 10° to 12°.

(2) **Funnel-shaped** sinkholes also are categorized as solution sinkholes. They have diameters that are two to three times their depth and side slopes of 30° to 45°. They are less common than the bowl-shaped sinkholes.

(3) **Cylinder-shaped** sinkholes have a depth greater than their diameter and steep to vertical slopes. They are very rare relative to the other two types of sinkholes.

The second major topographic feature in karst topography is karst valleys. In general, these can be divided into four types with characteristic properties (Ritter, 1978):

(1) **Allogenic valleys** are characterized by steep narrow gorges, with canyon-like walls (Ritter, 1978). The heads of the valleys are developed in impermeable rocks adjacent to the karst region. They are often associated with subsurface drainages, meandering caves, and natural bridges. The longitudinal extent of the valley depends on the discharge of the drainage basin above the karst terrain (Ritter, 1978).

(2) **Blind valleys** are characterized by rivers that traverse a karst surface and eventually sink into an underground drainage system (Ritter, 1978). The point of infiltration is often identified by a limestone scarp. The limestone scarp forms as a result of differing erosion rates between the upstream and downstream drainages at the locus of infiltration into the subsurface hydrologic system. The scarp can range from a few meters to tens of meters (Ritter, 1978).

(4) **Pocket valleys** are the opposite of blind valleys in that they develop in the regions where groundwater emerges rather than sinks (Ritter, 1978). They are commonly associated with large springs that discharge from impermeable bedrock. They are usually U-shaped in cross-profile with steep sidewalls, and a steep headwall near the spring.

(5) **Dry valleys** represent the most common type of karst valley and are similar to normal fluvial drainage valleys except that they have no well-defined surface source (Ritter, 1978). They are steep-sided and have a well-integrated drainage pattern with numerous dendritic branches.

Regional, local, and feature-specific characteristics of karst-dissolution features are summarized in Table 2-12. For a detailed bibliography of principal references related to karst geomorphology and hydrogeology see Kastning (1994).

Table 2.12 Summary of characteristics: limestone dissolution (karst) structures

Regional context	Limestone dissolution (karst) structures
Tectonic	Not restricted; interplate and intraplate tectonic settings.
Relation to tectonic stress/strain fields	No consistent relationship to contemporary tectonic stress or strain fields.

Table 2.12 continued

Geologic	Occurs in regions with limestone and other soluble carbonate rock (i.e., marble, dolomite and gypsum). Karst areas developed in dolomite bedrock usually have less well-developed sinkholes and more bedrock rubble.
Geophysical	No association with a particular geophysical setting.
Seismologic	No association with regional seismicity.

Local context	**Limestone dissolution (karst) structures**
Topographic	Generally confined to areas of low to moderate relief; entrenched valleys below uplands of soluble and well-jointed rock favors development of karst.
Geomorphic	Characterized by karst morphology (i.e., conical and/or closed depressions and karst valleys with underground streams).
Stratigraphic	Dissolution of limestone causes collapse of surface deposits, which dip basinward. Fine-grained sediment may be deposited in center of sinkhole if there is no surface water drainage.
Structural	Tensional fractures and arcuate normal faults surrounding sinkhole.
Geophysical	No geophysical anomalies; voids may be imaged by geophysical profiling techniques.
Seismologic	Limestone dissolution structures are aseismic.
Hydrologic	Regional groundwater and surface water (chemistry and distribution) control development of karst topography. Karst topography is controlled by aquifers of high transmissivity, mildly acidic water, and/or regions with abundant surface water.

Feature characteristics	**Limestone dissolution (karst) structures**	
Morphology	Classic karst topography (circular depressions and karst valleys).	
Geometry	*Plan View -*	Concentric depressions and elongate valleys with/without subsurface drainages.
	Cross section -	Subsurface drainages are via solution fractures, partings, and caverns.
Structure	Faults and fractures encompass depression and exhibit normal displacement that does not extend below soluble rock; although karst-related faults reactivate pre-existing structures.	
Scale	Common 10 to 100 m in diameter, 1 to 100 m in depth, and as large as 1 km wide. Sinkholes can form in large numbers (40-50 sinkholes/km). Vertical shafts may be as much as 10 m in diameter and hundreds of meters deep.	
Sense of deformation	Extensional: collapse of overburden into depression by normal faulting processes, or toppling; depressions may form by subsidence of bedrock blocks.	
Displacement/event	Variable; as much as tens of meters, although individual collapses may involve only parts of entire sinkhole.	

Table 2.12 continued

Depth	Structures extend from a few to hundreds of meters in depth; sinkholes usually collapse to depths at or just below the water table.
Spatial associations	Karst topography forms where there is abundant surface or groundwater in contact with limestone.
Hydrologic	Acidic groundwater in contact with limestone is responsible for dissolution. Rapid subsidence or collapse may occur during floods, when subsurface streams have high erosion capability, or during drought, when collapse is triggered by decreases in pore pressure.
Temporal characteristics	
Rate of deformation	Rates of deformation range from uniform to episodic for collapse.
Episodicity	Collapse or subsidence of overburden can be continuous or episodic.
Duration of deformation	Deformation processes are dependent upon abundance of groundwater or surface water and source rock abundance. In addition, duration of deformation may be limited by processes that remove debris from the base of the collapse structure and that allow additional collapse.
Recurrent deformation	Deformation long-term.
Temporal associations	Initiation of dissolution processes coincides with contact between water and limestone; collapse ceases when volume of material spalled into depression stabilizes overburden forces, or there is a reduction in groundwater.
Investigative techniques	Regional and local geomorphic mapping combined with drilling, aquifer characterization, flowmeters, dyes and "sonar calipers." Remote sensing to identify geometric or vegetative patterns; geophysical surveys (seismic reflection and refraction, ground penetrating radar, and electrical resistivity) may provide additional data to characterize localized karst zones.
References	Ritter, 1978; Easterbrook, 1993; Kastning, 1994; Kastning and Kastning, 1994; Palmer, 1991.

2.6.7 Volcanic-Related Faults

Magma intrusion is an important component of worldwide crustal extension (Parsons and Thompson, 1991; Gans, 1987; Coney, 1987; Forselund and Gudmundsson, 1991; Lepine and Hirn, 1992; Rubin and Pollard, 1988). Seismicity, surface faulting, magma intrusion, and volcanism are expressed within many tectonic settings, and extension of the brittle crust is accommodated by a combination of normal faulting and magmatic (dike intrusion) processes (Bursik and Sieh, 1989; Parsons and Thompson, 1991; 1993; Forslund and Gudmundsson, 1991). However, areas of ambiguous or complex interplay of tectonic and magmatic processes exist. An important issue pertaining to seismic hazard evaluations in such areas is differentiating surface deformation and seismicity associated with tectonic processes from that associated with magmatic processes. In addition, the geologic record may contain evidence of faulting related to volcanic processes, but which could be interpreted as a result of tectonic processes during large earthquakes. This section addresses two such volcanic processes: (1) caldera collapse, and (2) magma intrusion.

2.6.7.1 Caldera Collapse Structures

Calderas are broad collapse depressions that form from the expulsion or injection of shallow (< 10 km) volcanic material (basaltic and silicic magmas), and are found within any tectonic or volcanic setting (Hackett et al., 1996). Structures formed by calderas accommodate principally vertical strain (lumesence, collapse, and resurgence) over the subcaldera magma body and radial and concentric strain peripheral to the magma chamber. Calderas are characterized by incipient ring fractures, which after eruption and collapse of the caldera, show normal displacement. Ring fractures extend to at least 5 km in depth, and at the ground surface are often associated with a ring of volcanic domes (Nowell, 1996; Christiansen, 1979). Calderas range from 2 to 60 km in diameter, and are proportional in size to the volume of ash-flow material that was originally displaced from the upper magma chamber (Christiansen, 1979).

Calderas can range in complexity from simple, single-eruption calderas to nested calderas that have undergone successive eruptions (Christiansen, 1979). Shallow (<10 km) microseismicity may be associated with calderas during episodes of resurgent uplift or subsidence, due to periodic magmatic injections (e.g., Long Valley caldera, California), as well as during the formation of the caldera. Defining the activity of the ring faults can be complicated by regional tectonic faults that intersect or are in close proximity of a caldera (e.g., Valles caldera, New Mexico).

2.6.7.2 Dike-Related Faults

Faults produced by volcanic processes that may be misinterpreted as tectonic surface rupture generally are restricted to volcanic rift zones and are related to dike intrusion (Smith et al., 1996, Appendix A; Hackett et al., 1996). Volcanic rift zones are the surface expression of dike swarms.

Dike intrusion may cause extensional deformation and seismicity within the volcanic rift zones. Tensile fissures with little or no vertical displacement, fissure swarms, flexural monoclines, and normal faults commonly are symmetrically distributed about a central eruptive fissure, sometimes forming a graben above shallow dikes. These structures typically occur within broad zones reflecting their origin by repeated dike injection.

Extensional faults in volcanic terranes are the surface expression of shallow intrusion and can be misinterpreted as structures associated with major tectonic faults. Dike-induced structures have the following general characteristics:

(1) An inferred or demonstrated association with cogenetic volcanic rocks (*See* Figure RS-1b, Smith et al., Appendix A). In near-vent areas, the structures commonly are buried completely by cogenetic volcanic materials.

(2) Occurrence of dike-induced surface ruptures in diffuse belts up to several kilometers wide (*See* Figure RS-2, Smith et al., Appendix A).

(3) A graben or two zones of non-eruptive fissures, commonly are symmetrical about an eruptive fissure (*See* Figure RS-1b, Smith et al., Appendix A).

(4) Downdip extents (fault widths) of dike-induced faults and fissures are only slightly greater than the depth to the top of the dike (Rubin, 1992). Because dikes intrude to shallow depths within the crust, and dike tops are usually <5 km from the surface (Ryan, 1987; Gudmundsson, 1984), the rupture areas of dike-induced faults are small.

(5) Tensional fissures are the most abundant structural feature (*See* Figure RS-2, Smith et al., Appendix A) where there is little net vertical displacement across graben, and slickensides are seldom observed on fault scarps, suggesting that most movement is purely dilational.

(6) Vertical offsets on faults range from less than a meter to several tens of meters, depending on the composition and number of associated dikes. Colluvial wedges containing records of single large earthquakes generally do not form in dike-induced structures because the surficial materials commonly are volcanic bedrock with little or no colluvium. Instead, fault scarps with several-meter vertical displacements in volcanic bedrock may reflect the cumulative effects of many decimeter-scale displacement events from several dike-injection episodes.

(7) Monoclinal flexures are common because the structures typically occur within volcanic rock sequences (particularly lava flows) which tend to drape over shallow faults rather than break cleanly to the surface.

(8) Vertical displacements vary abruptly along strike, and individual faults are short (hundreds of meters to about 10 km), commonly grading into monoclines or purely tensional fissures (*See* Figure RS-2a, Smith et al., Appendix A).

(9) Structures may be associated with symmetrical geophysical anomalies (Flanigan and Long, 1987; Schoenharting and Palmason, 1982). This contrasts with the asymmetrical anomalies typical of large-displacement tectonic faults.

(10) On a regional scale, even after millions of years of activity, belts of dike-induced deformation maintain a subdued topography (i.e., the eastern Snake River Plain of Idaho and the island of Iceland) (Parsons and Thompson, 1991). This contrasts with conspicuous structural and topographic relief developed over similar time periods in many tectonic environments.

(11) Rupture and displacement on dike-related faults and fissures migrate incrementally at about the velocity of propagating dikes (0.5 m/s) (Rubin, 1992).

(12) The small offsets and rupture areas of dike-induced faults, and their incremental growth in tandem with dike propagation, suggest that the magnitudes of associated earthquakes will be small to moderate. Observations of seismicity from volcanic rift zones worldwide corroborates this inference, showing the maximum magnitudes of dike-induced earthquakes are M 5.2 ± 0.9 (Jackson, 1994; Einarsson, 1991; Brandsdottir and Einarsson, 1979; Hauksson, 1983).

Regional, local, and feature-specific characteristics of dike-induced structures are summarized in Table 2.13.

Table 2.13 Summary of characteristics: volcanic (dike-related) faults

Regional context	Volcanic (dike-related) faults
Tectonic	Divergent interplate boundaries; extensional hotspot intraplate settings.
Relation to tectonic stress/strain fields	Possible - dike emplacement in upper crust may be affected by shallow crustal stress (they would be predicted to form parallel to the orientation of maximum horizontal compression).
Geologic	Volcanic settings are characterized by broad basalt fields and shallow dike intrusions.
Geophysical	Areas of high heat flow.
Seismologic	Seismically active areas.
Hydrologic	No association with regional hydrologic characteristics.

Local context	Volcanic (dike-related) faults
Topographic	Subdued topography and low-relief terrain.
Geomorphic	Diffuse belts (up to several kilometers wide) of graben, normal fault scarps, subdued scarps (monoclinal flexures), and fissures along rift zones.
Stratigraphic	Basaltic lava flows (flood basalts); absence of colluvial wedges adjacent to bedrock fault scarps because of erosional resistance of basalt and small fault displacements.
Structural	Extensional features - normal faults; graben or zones of non-eruptive fissures are symmetrical about an eruptive fissure; tensional fissures (slickenslides rare); monoclinal flexures.
Geophysical	Structures may be associated with symmetrical geophysical anomalies (e.g., aeromagnetic or gravity highs) related to dike swarms at depth.
Seismologic	Structures can be seismogenic; earthquakes associated with dike intrusion generally range from M3 to M5.3; migrating seismicity swarms accompany dike intrusion.
Hydrologic	May be associated with hot spring activity - springs may exhibit mantle-derived chemical composition.

Feature characteristics	Volcanic (dike-related) faults	
Spatial characteristics		
Morphology	Linear graben; tensile fissures; monoclines; vertical normal faults.	
Geometry	*Plan view -*	Fissures and faults are symmetric about central eruptive fissure; vertical displacement of faults vary abruptly along strike and commonly grade into monoclines or purely tensional fissures.

Table 2.13 continued

	Cross section -	Vertical to steeply-dipping graben-bounding normal faults; drape folding (monoclines); vertical offset on faults range from less than 1 m (associated with single dike) to several tens of meters (associated with thick, silicic, dikes or repeated injection of closely spaced basaltic dikes); normal faults terminate above or within dike tops.
	Aspect ratio (H:V) -	Low to very low (2:1 or less) for individual scarps.
Scale	Graben are up to 10 km long and 3 km wide; downdip extent of faults is generally less than 5 km.	
Sense of deformation	Extensional deformation (fissures, graben) above dikes at depth.	
Depth	Dike-induced graben faults may extend to about 5 km.	
Spatial associations	Brittle features typically are associated with cogenetic volcanic materials at surface; structures are associated with dikes at depth.	
Hydrologic	May be associated with hot-spring activity.	
Temporal characteristics		
Rate of deformation	Variable - dike intrusion events range from hours to centuries; velocity of a propagating dike is about 0.5 m/s; volcanic recurrence generally on order of tens to hundreds of thousands of years.	
Episodicity	Dike intrusion events can be episodic.	
Duration of deformation	Hours to centuries.	
Recurrent deformation	Graben faults may exhibit evidence of recurrent movement associated with multiple dike injection.	
Temporal associations	Migrating swarms of small to moderate earthquakes associated with dike intrusion.	
Investigative techniques	Detailed geologic and geomorphic mapping; paleoseismic analysis. Numerical modeling and geodetic monitoring provides data to evaluate the relationship of dike geometry to stress and strain distributions and to surface deformation.	
References	See citations listed in Smith et al. (Appendix A).	

2.7 Non-deformation Features

A variety of non-deformation features may be confused with tectonic faulting. These features may be produced by geomorphic, stratigraphic, pedogenic, or hydrologic processes. Non-deformation features most commonly confused with tectonic faulting include the following:

- Geomorphic
  - lineaments and scarps of nontectonic origin, including marine and fluvial terrace risers, and differential erosion controlled by pre-existing faults (i.e., fault-line scarps), bedding or joints
- Stratigraphic
  - buttress unconformities and other anomalous relationships between

sedimentary facies not associated with tectonic deformation

- Pedogenic
  - various soil phenomena formed by differential weathering and development of soil profiles, including the development of soil "cutans" (Abbott et al., Appendix A)

- Hydrologic
  - abrupt groundwater level changes or gradient anomalies resulting from facies changes or other stratigraphic variations

In particular, pedogenic soil cutans or "clastic dikes" are common in the eastern and southeastern United States, and have been attributed to surface faulting or strong vibratory ground motion (Abbott et al., Appendix A). Many of these features result from preferential soil weathering along pre-existing joints, fractures, or faults, and thus have preferred orientation and lateral continuity. The origin of these features is discussed in greater detail in Section 2.7.1, Soil Phenomena.

By definition, "non-deformation" features result from processes other than deformation, whether tectonic or nontectonic. The simplest criterion to differentiate these features from deformation-related features, therefore, is to demonstrate the absence of deformation associated with the feature. If it can be shown that there is no fault or fold spatially associated with one of these features, then the feature likely has a non-deformation origin.

In addition to subsurface information to demonstrate the absence of faulting associated with these features, a variety of geologic and geomorphic criteria can be used to identify and differentiate non-deformation features. For example, both marine and fluvial terrace risers produce nontectonic scarps that may be confused with fault scarps. These features may be differentiated from tectonic scarps using the following criteria:

- Marine and fluvial terrace risers generally are subparallel to the coastline or river, respectively, and typically have cuspate or arcuate forms in plan view.

- Marine and fluvial terrace risers commonly occur in an inset sequence of terraces. Younger, lower terraces commonly are cut into and erode older, higher terraces, but not *vice versa.*

- The marine terrace backedge maintains a relatively consistent elevation, except where deformed.

- The fluvial terrace backedge forms a uniform downstream gradient similar to the current stream channel gradient, except where deformed.

- Marine and fluvial terrace risers commonly are sinuous depending on bedrock lithology and structure, incidence angle of marine waves, and competence and capacity of fluvial drainages.

Fault-line scarps form by preferential erosion and exhumation of a pre-existing bedrock fault. Fault-line erosion occurs most commonly where the pre-existing fault has juxtaposed bedrock lithologies of varying erosional resistance, such as soft clay or shale against resistant sandstone or crystalline rock. Because of the different rock types, fault-line scarps may face the opposite direction than expected if related to surface rupture. Fault-line scarps may be differentiated from tectonic scarps using the following criteria:

- independent evidence of fluvial erosion and/or mass wasting (e.g., landsliding) along the fault trace

- non-uniform scarp height, particularly where larger scarps are developed on younger surfaces

- continuity and scale of the scarp is limited to (and coincident with) areas of contrasting bedrock lithologies

- presence of unfaulted Quaternary deposits across the fault (i.e., direct evidence of no faulting)
- absence of geomorphic and stratigraphic evidence of Quaternary faulting (i.e., indirect evidence of no faulting)
- scarp aspect inconsistent with sense of fault slip

Buttress unconformities include features such as cut-and-fill channel margins, buried fault scarps, glacial ice contacts, and landslide buttresses. These unconformities commonly produce abrupt, near-vertical stratigraphic contacts. Minor shearing and/or clast imbrication may occur along the contact during primary deposition or post-depositional differential settlement. In addition, these unconformities commonly produce changes in groundwater flow parameters (e.g., permeability) and clay may accumulate along the contact. Buttress unconformities may be differentiated from tectonic faults using the following criteria:

- absence of offset strata
- absence of tectonic shear fabric along the fault
- recognition of depositional environment (e.g., fluvial, marine, landslide)
- recognition of channel geometry related to buttress unconformity
- absence of lateral continuity over local and regional scale.
- sinuous nature of buttress unconformity in plan view
- presence of "rip-up" clasts within inset fluvial deposit that are composed of adjacent or nearby rock or soil types

The recognition of non-deformation features requires a thorough understanding of the geologic context within which the features exist. All observations and data must be consistent with the local geologic and tectonic setting. If a feature cannot be explained by a non-deformation process, then a tectonic or nontectonic deformation process should be considered.

2.7.1 Soil Phenomena

The process of weathering and soil formation may produce features suggestive of active faulting and/or earthquake-induced ground deformation. For example, expansive, clay-rich soils may contain a distinctive shear fabric due to shrink-swell processes and differential soil movement. The shear fabric may be pervasive or it may be localized in areas of abrupt nontectonic changes in soil thickness or parent material texture. Shear surfaces within the soil may be striated or slickensided. Pedogenic shear fabric may be distinguished from tectonic faulting or earthquake-induced ground deformation by the following:

- The shear fabric will be confined to expansive soils.
- The shear fabric will not extend to depth into unweathered sediment or rock.
- The shear fabric typically will be pervasive and not confined to narrow zones.
- The shear fabric will not be concentrated over faults in underlying bedrock.
- The shear fabric will contain shears with multiple orientations.
- The shear fabric will not persist laterally with a consistent orientation.

Another common phenomenon of surface weathering and soil formation is the development of soil cutans. Soil cutans are tongues of deep weathering that penetrate deeply into the substrate (*See* Abbott et al., Appendix A). Soil cutans are common in the Tertiary Coastal Plain of the southeastern United States, where they have been referred to as liquefaction-induced "clastic dikes" (e.g., Siple, 1967; Zupan and Abbott, 1975; Seeber and Armbruster, 1981). They also have been interpreted to result from solution of underlying carbonate horizons and sediment collapse (Siple, 1967). Reconnaissance studies (Johnson and Heron, 1965; Heron et al., 1971) and detailed studies (Abbott et al., Appendix A) show, however, that the "clastic dikes" are not the result of forced injection of material but, rather, are a result of differential soil weathering.

Abbott et al. (Appendix A) performed a detailed study of the "clastic dikes" described by earlier authors at the Savannah River site in South Carolina. They reinterpret the "clastic dikes" as soil cutans, which are defined as "a modification of the texture, structure, or fabric of the host material by pedogenic processes, either by a concentration of particular soil constituents or in-situ modification of the matrix". Abbot et al. show that the similarity of material within the features and that of the host material, as well as undisrupted bedding within and across the features, demonstrates that the features are not seismically induced or aseismic liquefaction features. Furthermore, the undisrupted horizons show that the features are not tree throws, root casts, animal burrows, or ice wedges. The absence of offset across virtually all of the features demonstrates that they did not develop as faults. The features were interpreted by Abbott et al. (Appendix A) to have developed through pedogenic processes based on:

- their similarity in appearance and apparent concomitant development of the features to a sub-horizontal zone of geochemical alteration present at the top of each exposure in the soil profile

- an overall thinning and pinch-out of the features with depth, and their common occurrence in the upper parts of exposures

Two types of soil cutans were identified by Abbott et al. (Appendix A): irregular cutans and structurally-controlled cutans. Irregular cutans are irregular or polygonal in shape, and do not appear to be controlled by any pre-existing structure. Structurally-controlled cutans are planar to curviplanar and generally exhibit locally strong preferred orientations. Within most exposures, structurally controlled cutans share a common orientation with adjacent joints. At one locality, the soil cutans were coincident with fault planes. These relationships suggest that the location and orientation of some soil cutans may be controlled by pre-existing structures.

In summary, soil weathering processes are responsible for near-surface soil features that were previously interpreted as tectonic features. Differentiation between these soil phenomena and tectonic features requires on understanding of soil-forming processes.

3 DIFFERENTIATING TECTONIC AND NONTECTONIC FAULTS

Assessing the tectonic or nontectonic origin and seismogenic capability of faults exposed at the surface involves: (1) identifying diagnostic criteria for faults of varying origins, (2) evaluating the relative value of the individual criterion or suites of criteria, (3) comparing the characteristics of the fault in question to these criteria and interpreting the fault's origin, and (4) developing an internally consistent model that explicitly outlines uncertainties in the available data. The first three steps in this process are discussed in the following sections, and the latter step is addressed in Section 4.

Table 3.1 Data sets needed to evaluate fault origin

Regional context	Local context	Fault-specific characteristics	
		Spatial	***Temporal***
Physiography	Topography	Morphology	Rate of Deformation
Tectonic Setting	Geomorphic Setting	Geometry	Episodicity
Relation to Regional Stress or Strain Fields	Stratigraphic Setting	Scale	Duration
Geologic Setting			
Geophysical Setting	Structural Setting	Sense of Deformation	Recurrence
Seismologic Setting	Geophysical Setting	Depth	Associations
Hydrologic Setting	Seismologic Setting	Displacement/Event	
	Hydrologic Setting	Associations	

3.1 Criteria for Assessing Fault Origin

Assessing the origin of near-surface faults generally requires: (1) an understanding of the types of tectonic and nontectonic processes that can produce faults; (2) an understanding of the regional and local geologic setting within which the fault occurs; and (3) knowledge of the specific characteristics of the feature in question. Various types of data and information are required to differentiate between a tectonic and nontectonic origin. These types of data are listed below, and briefly described in the following section.

3.1.1 Regional Context

Many faults are or are not consistent with certain geologic settings, thus making the assessment of regional geologic and tectonic framework an important part of determining fault origin. **Physiography** describes the shape and dimensions of the landscape. Evolution of the landscape is strongly influenced by underlying bedrock lithology, geologic structure, and the geomorphic processes that have modified it. Although many nontectonic faulting mechanisms are not restricted to any particular physiographic setting, some associations can be made and should be considered. For example, some structures such as sackungen and deep-seated landslides require significant topographic relief to develop, and generally are restricted to mountainous regions;

growth faults are ubiquitous along passive margin settings where large-scale gravity gliding/spreading is common; subsidence-related faults and fractures that are caused by groundwater withdrawal are generally observed in basins in semi-arid to arid regions; and some types of soil phenomena that may be confused with tectonic faults, fractures, or paleoliquefaction features are commonly observed in the Coastal Plain of the southeastern U.S.

A basic understanding of the contemporary **tectonic setting** of a region is necessary to evaluate the kinematic and structural relationship of recent surface deformation to tectonic features of varying ages. The general tectonic setting (e.g., transcurrent, extensional, convergent) should be considered in any evaluation of fault origin. Some types of faulting are typical of or restricted to certain types of tectonic provinces. For example, salt flowage structures, including salt anticlines, salt domes, salt diapirs and growth faults, generally are restricted to regions that experienced extension. In addition, there may be significant differences between faults in interplate regions and those in stable continental regions (SCR). Interplate tectonic settings generally are characterized by relatively high rates of seismic activity and active faults with moderate to high slip rates. Active faults typically exhibit geologic, geomorphic, or seismologic evidence of surface deformation. Less active or secondary structures, even in more tectonically active regions, may be more difficult to identify than active primary structures.

On the other hand, the overall rate of seismicity in SCR's is low in comparison to plate boundaries (e.g., Johnston, 1989). The low strain rate combined with high rates of deposition and erosion that may occur in SCR lead to difficulties in identifying and characterizing active seismogenic faults. Paleoseismic studies of surface rupturing SCR earthquakes show that seismogenic faults have either very low rates of deformation (e.g., Crone et al., 1992; Machette et al., 1993), or little or no brittle tectonic deformation prior to a historic rupture (e.g., Adams et al., 1991). Recent data on intraplate earthquakes and associated faults suggest that potentially seismogenic faults are widely scattered and commonly are reactivated pre-existing faults that have very small cumulative displacements in the current tectonic setting (Adams, 1996; Rajendran et al., 1996; Seeber et al., 1996). Seeber et al. (1996) conclude that future earthquakes are likely to occur on faults with little or no evidence of past earthquakes and that moderate-size earthquakes may occur at shallow depths and rupture to the surface.

If the sense of slip on a fault is not consistent with the **contemporary stress or strain field** as determined from instrumental measurements of *in situ* stress (e.g., borehole blowouts, overcoring experiments, etc.) and indicators of regional strain (e.g., earthquake focal mechanisms, kinematic analysis of geologic structures), then a nontectonic or secondary tectonic origin should be considered likely. For faults having a long history of deformation that may involve reactivation, it is important to differentiate the most recent period of movement from earlier displacements that may reflect a different style of faulting in a previous tectonic regime. Inversions of earthquake focal mechanisms for brittle strain (Unruh and Lettis, 1998; Unruh et al., 1996) may provide evidence of distinct seismotectonic domains, which may be characterized by different styles and orientations of fault movement.

An understanding of the general **geologic setting**, including the lithologies, distribution, thickness, and character (e.g., soluble, compressible, mobile, or weak units) of geologic units, both in the surface and subsurface, is needed to evaluate the potential for several nontectonic faulting mechanisms. The history of natural or human-induced loading should be noted. Evidence for nearby volcanic, hydrothermal or geothermal activity, glaciation, and strong ground motion also should be identified. **Geophysical** anomalies (e.g., magnetic and aeromagnetic lineaments, heat flow data) in the site region that are indicative of specific geologic environments or structural features should be noted.

Seismologic data, including the spatial and temporal distribution of historical and instrumental data, and possible associations with tectonic faults, depth of seismicity, and focal mechanisms, provide data to evaluate the location, activity, style, geometry, and depth of faulting.

It is also important to consider the regional **hydrologic setting** in assessing a fault's origin. Both long-term climatic cycles as well as short-term (e.g., meteorologic or human-induced) influences may be significant. In particular, evidence for significant natural or human-induced fluid withdrawal or groundwater level changes should be considered.

3.1.2 Local Context

Local conditions may strongly influence the style and scale of surface faults and other features that could be interpreted as related to tectonic processes. Thus, interpretation of a fault's origin should address the local topographic, geomorphic, stratigraphic, structural, geophysical, seismologic, and hydrologic conditions of the site vicinity. The location and orientation of faults with respect to local **topography** is critical and should be noted. For example, assessing whether a fault trace trends across topography or is restricted to the flanks or crests of ridges is a potentially useful diagnostic criteria of a fault's origin. The latter may suggest a gravitational origin rather than a tectonic origin. Topographic position and aspect also influence the climatic condition at a site, which, in turn, influence rates of surficial processes (e.g., erosion, weathering) that produce or modify evidence of surface faulting. In addition, local paleo-topographic conditions also should be considered, if possible, when evaluating the environment in which a particular fault may have formed. **Geomorphic features** and assemblages of landforms that may provide an indication of faulting processes within the site area should be described and noted. In particular, features that demonstrate the lateral and vertical continuity of a fault, or provide evidence of the timing and duration of recent surface deformation, should be noted.

Local **stratigraphic** characteristics also are important. The presence of unstable geologic strata, weak (detachment) horizons, soluble strata, or unconsolidated, compactible sediments may provide evidence that a fault is or is not related to tectonic processes. Stratigraphic relationships that provide evidence for continuous or episodic deposition or displacement should be described, and evidence for cumulative displacement on a fault should be noted. An understanding of the general orientations and styles of **structural features** (e.g., faults, joints, folds), and their relative ages is critical to constructing the local structural framework of fault-related features and determining their origin. The degree of deformation and orientation of underlying bedrock structures (both tectonic and nontectonic) may play a significant role in identifying nonseismogenic faulting. For example, regions of localized active uplift may be more susceptible to nontectonic gravitational failures due to over-steepening of hill slopes.

Available **geophysical data** (e.g., seismic reflection and refraction data, gravity and magnetic surveys, heat flow data) should be reviewed to identify and characterize the spatial distribution of geologic anomalies that may be significant. For example, seismic reflection data have played a critical role in characterizing salt-related features and the three-dimensional geometry and structural associations of related faults.

Comparison of historical and instrumental **seismicity** patterns to specific features should be made to evaluate possible associations of moderate- to large- magnitude earthquakes and microseismicity. The style of seismicity, perhaps interpreted from focal mechanisms, should also be noted.

Hydrologic features and conditions may indicate the location and activity of surface or near-surface faults. Major earthquakes cause significant hydrogeological changes, both temporary and very long-lived. Temporary or short-term effects include regional increases in ground-water discharge that typically last a few months and short-lived fountains (e.g., resulting from spontaneous hydrofracture in the hanging wall of a normal fault) (Muir-Wood, 1992). Seismic pumping caused by coseismic volumetric strain affecting crack apertures and crustal porosities may result in water table rises on the order of a few tens of meters. Permanent changes generally relate to hot springs, and can involve the staunching of pre-existing springs, changes in discharge rates, and the creation of new springs.

As summarized by Muir-Wood (1992) these phenomena can be recorded geologically in a variety of ways. Certain kinds of banded mineralization may reflect transient changes in groundwater chemistry or temperature, in effect producing earthquake varves. Earthflows and craters resulting from landslides induced by the coseismic release of groundwater beneath poorly drained unconsolidated materials may provide geomorphic evidence of recent earthquakes. And lastly, the presence of hot springs themselves may be indicative of near-surface faults. Pazzaglia (Appendix A) notes that in the eastern United States the alignments of hydrologic features, such as springs or hydrothermal deposits may be indicative of tectonic faults.

Hydrologic conditions favorable for surface or near-surface faulting should be noted. Alignments of hydrologic features, such as springs or hydrothermal deposits may be indicative of tectonic faults (Pazzaglia, Appendix A). Tectonic faults (active or inactive) also may act as hydraulic conduits and/or aquitards that affect local groundwater conditions and influence slope stability and near-surface ground deformation. Historical records of groundwater fluctuations should be reviewed in areas where subsidence and related groundsurface deformation are suspected.

3.1.3 Fault-Specific Characteristics

Commonly, data specific to a fault are required to assess whether the fault is related to tectonic processes. Both spatial (morphology, geometry, scale, depth, sense of deformation, displacement, etc.) and temporal (rate of deformation, episodicity, duration, recurrence, etc.) characteristics of a fault should be described and evaluated.

Detailed data on the **morphology** (length, height, continuity, lateral variability, distribution, etc.) of surface scarps, fissures, and related surface deformation are important for assessing the style of deformation, and the timing and displacement of faulting events. The **geometry** of a fault in three dimensions also is an important characteristic that provides information on the processes of formation. The pattern of deformation commonly provides data to interpret the kinematics associated with the formation of a fault, and thus its origin. In plan view, both the plan shape (linear, arcuate, etc.) and the distribution of faults (e.g., single, multiple, en echelon, clustered) should be described. In cross section, the geometry of a fault at depth (e.g., listric, high-angle, continuous) should be evaluated. Horizontal-to-vertical (H:V) aspect ratio is important for evaluating fault origin. Aspect ratios of less than 0.5 (H:V) likely are not tectonic and, if tectonic, probably are not seismogenic. For tectonic faults, high (> 5 m) scarps formed during a single event commonly occur on faults with long (several km to tens of km) traces (Wells and Coppersmith, 1994). Faults with relatively high fault scarps and limited lateral extent (i.e., anomalously low H:V ratios) may be indicative of a nontectonic mechanism.

The areal extent of a fault is an excellent indicator of the **scale** of the processes by which it formed. Tectonic faults that are capable of generating moderate- to large-magnitude earthquakes require significant crustal extent in both lateral and downdip directions. The extent and continuity of a fault yield information on process, and thus may enable differentiation among various origins. The **depth** of faulting, therefore, is an important factor in determining the seismogenic potential of a fault. Only faults that extend to seismogenic depths (> 3 km depth) are likely to generate moderate- to large-magnitude earthquakes with associated surface deformation. Subsurface data (e.g., drilling records, geophysical surveys) and seismologic evidence for the geometry and depth of faults should be described and evaluated.

Structural, stratigraphic, and seismologic data that provide evidence for the **sense of deformation** in a local region and/or for a specific fault should be described and evaluated with respect to the local setting. Offset linear markers are one of the most reliable indicators of the sense and amount of displacement, but often are not available for an individual fault with limited exposure. An alternative method to determine the sense, but not amount, of slip on a fault is the analysis of kinematic indicators, or shear criteria, that are formed within or adjacent to the fault zone itself. These include, but are not limited to, rotation of geologic markers such as bedding, asymmetric folds, lineated slickensides (slickenlines), riedel shears and their intersection with high strain, fault-parallel y-shears, dismembered and rotated clasts, and shear fabrics formed in foliated gouge. Inconsistency among shear criteria within a single fault zone is common, due to misinterpretation of structures and evidence of non-parallel slip events on a single fault. Therefore, it is best to gather as much data of this kind as possible prior to making a conclusion regarding sense of movement.

Hydrologic features and local groundwater conditions may be important for evaluating fault origin. Evidence for groundwater anomalies (e.g., springs, ponded areas, vegetation alignments) along or in the vicinity of a fault should be evaluated. Fluctuating or high groundwater conditions that may influence the stability of a site or result in the formation of anomalous soil features should be evaluated.

Geologic (i.e., stratigraphic or structural data) and seismologic data that provide evidence for the timing of surface deformation are important to the assessment of the origin of faulting. The **rate** or magnitude of strain associated with a fault may be indicative of the mechanism of formation. For example, if the rate of deformation is not consistent with tectonic rates of activity in the region (i.e., is anomalously high or of relatively short duration), a nontectonic origin should be suspected. The **episodicity** and **duration of faulting** also should be evaluated. Both tectonic and nontectonic faults may be characterized by episodic behavior. Surface faulting events along tectonic faults result from the sudden release of accumulated strain on a fault and are chararacterized by episodic displacement events. Although creep may occur along some tectonic faults (chiefly strike-slip faults), the creep generally occurs within the upper few kilometers of the fault, and the accumulated strain along the lower portions of the fault generally is released only during the larger magnitude events. Evidence for continuous surface faulting is more typical of some types of nontectonic deformation processes (e.g., subsidence-related surface faulting, some sackungen). Surface deformation that occurs during moderate- to large-magnitude surface faulting events occurs rapidly (within seconds to minutes). Faulting that occurs over an extended period (hours to years) is more likely the result of nonseismogenic, nontectonic processes.

Evidence for **recurrence** provides additional data to evaluate the long-term behavior of a fault. Single-event displacements in the absence of evidence for a longer term displacement history generally are not characteristic of tectonic faults. This criterion may be less applicable in stable continental regions, however, where there is some evidence from historical surface-faulting earthquakes to suggest that some faults are characterized by very small displacements accumulated in the current tectonic regime and/or long times since the previous rupture (Seeber et al., 1996).

Spatial and temporal **associations** with climatic (major climatic cycles), volcanic, glacial, isostatic-rebound, subsidence, collapse, fluid-extraction, or seismic events also provide supporting evidence for evaluating the tectonic or nontectonic origin of faulting. Evidence for the presence or absence of paleoseismic faulting, or features suggestive of strong ground motion in the vicinity of the feature in question should be described and evaluated. A fault that has a reasonable association with one or more large earthquakes or sustained earthquake activity that are usually accompanied by significant surface deformation is very likely a tectonic fault.

3.2 Diagnostic Criteria for Identifying Nonseismogenic–Tectonic Faulting

This section focuses on criteria that can be used to differentiate nonseismogenic faults from seismogenic faults. Based on our analysis of available data and information, there are few individual criteria that can be used to document the origin of a fault. Thus, distinguishing between nonseismogenic and seismogenic faults generally requires the use of several criteria, coupled with sound professional judgment. Direct evidence of a seismogenic fault origin is an exception, rather than a rule, and thus many determinations of origin rely on the preponderance of evidence and the relative validity of each criterion used in formulating a judgment. As noted in the previous section, the determination of fault origin will require an integrated analysis of (1) fault-specific attributes such as scale, geometry, and kinematics; (2) the geologic and tectonic setting (i.e., context) within which the fault occurs; and (3) a thorough understanding of the types of processes that can produce faults. Determining the geologic context requires an understanding of the geologic, seismologic, and tectonic setting of the area within which the fault is present. Most tectonic and nontectonic processes form faults with a characteristic geometry and sense of movement. Thus, the three-dimensional geometry, scale and kinematics of an individual fault are important for determining origin. In particular, the scale of a fault relative to topography and other geologic and tectonic features (e.g., bedding, folds, the Earth's crust) is critical for determining the origin of the phenomena. The following sections provide criteria that may be used to determine the origins of the nonseismogenic phenomena described in Sections 2.5 and 2.6. As above, we subdivide these phenomena into those directly related to tectonic processes (this section) and those related to nontectonic processes (Section 3.3).

3.2.1 Secondary Tectonic Faults

As described in Section 2.5.2, secondary tectonic faults are subordinate structures that accommodate deformation during slip events on primary seismogenic-tectonic faults. Secondary tectonic faults are considered to not contribute significantly to the seismic moment release during earthquake rupture (Yeats, 1986). Thus, secondary tectonic faults form as a passive mechanical response to non-uniform slip, and to the development of localized stresses during displacement on non-planar fault segments. However, primary tectonic faults also may experience secondary rupture during large earthquakes on adjacent primary tectonic faults. Because such primary tectonic faults also may produce surface rupture earthquakes independent of rupture on adjacent tectonic faults, this scenario is not considered as movement on a secondary tectonic fault.

Overall, secondary surface faulting may be very similar to that produced by primary fault rupture, and differentiating these faults produced by secondary rupture from those produced by primary fault rupture is difficult and commonly inconclusive. Secondary surface faults commonly have subtle or distributed surface expression, and may exist for only a short period of time before being buried, eroded, or disturbed. In addition, some features are similar in appearance to tectonic fault ruptures, such as ground cracks adjacent to the Monte Vista fault produced by the 1989 Loma Prieta earthquake (Langenheim et al., 1997).

There are no *highly diagnostic* criteria for differentiating features associated with secondary fault rupture from those produced by primary rupture on seismogenic tectonic faults. Therefore, investigations of features possibly produced by secondary fault rupture should consider the scale, timing, and patterns of the deformation, and should evaluate whether all the characteristics of a feature could have originated by primary or secondary deformation processes. Whereas the style and sense of primary faulting should be consistent with the regional stress field, secondary faulting may be variable and more dependent on local stress fields induced by basin geometry, sediment thickness, or other local features. In the absence of highly diagnostic criteria, using several *moderately diagnostic* criteria, as described below, may be sufficient to differentiate secondary tectonic features from primary tectonic rupture. In general, the use of a greater number of criteria translates into higher confidence in interpreting whether the feature is related to secondary or primary fault rupture.

Differentiating secondary faults from primary fault rupture requires local and regional geologic perspectives. Based on empirical data from past earthquakes, there are three *moderately diagnostic* criteria for differentiating between secondary and primary tectonic faults:

- Scale of faults
- Consistency with kinematic framework
- Displacement per event

Scale of Faults

Criterion: If the length or depth of a fault is such that it could not generate a surface-rupture earthquake, then it probably is a secondary tectonic fault or the result of a nontectonic process.

Generally, moderate to large earthquakes are produced by energy release over at least tens of square kilometers along a fault plane. If faults have lengths and/or depths of less than a few kilometers, then it is likely that they are not related to a seismogenic tectonic fault. This is based on the assumption that the surface extent of the fault represents the length along which rupture occurred. If subsurface data suggest that the fault extends to only shallow depths, then it is likely that they are related to secondary rupture. For example, secondary faults may be antithetic to primary tectonic faults, and thus intersect a seismogenic fault at depth, or they may be bending-moment faults produced by deformation in the hanging wall of a seismogenic reverse fault.

Consistent Kinematic Framework

Criterion: If the geometry, distribution, spatial relationship, and sense of movement of a possible secondary tectonic fault is consistent with the regional and/or local kinematic framework of a seismogenic deformation, then the fault may be a result of secondary tectonic faulting. Conversely, if the fault is inconsistent with the regional or local kinematic framework, then it likely is unrelated to secondary (or primary) tectonic deformation, and, more likely, is related to a nontectonic process.

As noted in Section 2.5.2, secondary tectonic faults have been recognized in virtually all tectonic environments (compressional, extensional, and translational [strike-slip]). In compressional environments, secondary deformation may include flexural-slip faults, bending-moment faults, chordal faults, relay (tear) faults, conjugate strike-slip faults, and hanging-wall collapse faults. Interpretation of a fault as one of these types of faults, therefore, requires a compressional tectonic framework. Similarly, extensional environments may include antithetic or sympathetic normal faults, antithetic reverse faults, and cross faults. Strike-slip environments may include secondary extensional or contractional faults. Interpretation of these faults also requires consistency with the regional tectonic framework.

Displacement Per Event

Criterion: If a fault is associated with large amounts of displacement during individual ruptures, then it likely is related to primary tectonic faulting or a nontectonic process.

Secondary tectonic faulting commonly includes distributed deformation, or deformation that typically is less than that produced by primary surface faulting. Thus, displacement of less than a few tens of centimeters may be attributed to secondary faulting. However, other non-seismogenic features may be associated with displacements that are much greater than a few tens of centimeters, and interpretations should consider other possible fault origins. In addition, primary tectonic faulting at the surface may be only a few tens of centimeters, if produced by a smaller-magnitude earthquake or distributed deformation at the ground surface.

3.2.2 Features Associated with Strong Ground Motions

Strong ground motion may produce secondary surface faulting similar to that produced by primary fault rupture and differentiating faults produced by strong ground motions from those produced by primary or secondary fault rupture is difficult and commonly inconclusive. Features produced by strong ground motion include earthquake-induced landslides, settlement, ground cracks, and liquefaction-induced lateral spreads and clastic dikes (Section 2.5.3). Faults associated with the processes may be poorly expressed at the ground surface and may exist for only a short period of time before being buried, eroded, or disturbed.

The scale, geometry, and kinematics of **seismically induced landsliding** are comparable to those of non-seismically induced landsliding (Fleming and Johnson, 1989; Mills et al., 1994; Gomberg et al., 1995). The criteria used to differentiate faults formed by landslide processes (*See* Section 3.3.1 below) from tectonic faults, therefore, also apply in the case of seismically induced landslides.

Differentiating between **ground cracks produced by strong ground motions** and small amount of tectonic surface rupture is very difficult. Extensive field investigations following the 1994 Northridge earthquake (Rymer et al., 1995; Holzer et al., 1996; Johnson et al., 1996) show that differentiating the effects of strong ground motion and primary surface rupture may be difficult and perhaps irresolvable. For example, fault scarps produced by liquefaction-induced lateral spreading in Potrero Canyon during the 1994 Northridge earthquake (Rymer et al., 1995) are similar in appearance to tectonic fault scarps. Small scarps and fissures on relatively level ground may form in response to settlement caused by compaction of unconsolidated sediment or densification related to liquefaction of saturated sediments. In paleoseismic trench exposures, **liquefaction-related phenomena** such as sand blows, dikes, and sills also may be confused with surface faulting, especially if the trench is limited in depth or length and fails to expose the liquefaction feature in three dimensions. Linear scarps and troughs formed by large lateral spreads locally may be confused with tectonic fault scarps. Complicating the determination of fault origin are the historical observations that locations of ground fractures produced by strong ground motion may have little or no relationship to the distance from the fault responsible for the earthquake, and instead may be highly dependent on site-specific geologic and hydrologic conditions.

There are no *highly diagnostic* criteria for differentiating faults caused by strong ground motion from those produced by primary rupture on seismogenic tectonic faults. Therefore, investigations of faults possibly produced by strong ground motions should consider the scale, timing, and patterns of deformation, and should evaluate whether all the characteristics of a fault could have originated by primary or secondary deformation processes. Whereas the style and sense of primary fault rupture should be consistent with the regional stress field, deformation resulting from strong ground motion may be variable and more dependent on local stress fields induced by basin geometry, sediment thickness, or other local features. In the absence of highly diagnostic criteria, using several moderately diagnostic criteria may be sufficient to differentiate strong ground motion-induced features from primary tectonic rupture. In general, the use of a greater number of criteria translates into higher confidence in interpreting whether a fault is or is not related to primary fault rupture.

As noted above criteria for differentiating faults associated with seismically induced landslides are described in a following section. In the following paragraphs we describe criteria for differentiating faults related to strong ground motion, particularly seismically induced liquefaction faults, from primary tectonic fault deformation. Based on empirical data from past earthquakes, the following *moderately diagnostic* criteria should be considered in evaluating the origin of suspected liquefaction or settlement-induced ground deformation:

- Regional geologic and tectonic setting
- Stratigraphic relations
- Hydrologic features
- Scale of features

Regional Geologic and Tectonic Setting

Criterion: If the regional distribution and abundance of possible liquefaction-related faults are consistent with the distribution of strong ground motions associated with a specific fault zone or scenario earthquake, rather than nontectonic phenomena such as flooding or frost heaving, then the features may be related to strong ground motion.

As with all secondary faulting, strong ground motion-induced deformation must be understood within the context of the regional tectonic and geologic setting. Faults associated with liquefaction-related features generally are concentrated locally at multiple sites distributed over regions with similar geologic and hydrologic conditions. For example, liquefaction-related features will occur only in those areas where the deposits are susceptible to liquefaction (i.e., saturated, unconsolidated, granular deposits) and the opportunity is present (i.e., ground motion occurs above some specified threshold level). The regional distribution and abundance of possible liquefaction-related features, therefore, should be consistent with the distribution of susceptible deposits and with strong ground motions associated with an earthquake, and not a nonseismogenic event (i.e., flooding-induced liquefaction). The regional pattern of liquefaction-induced deformation will be distributed broadly about the epicentral region and extend over greater distances for larger magnitude earthquakes. If sand blows or other features are distributed over a broad region, it is unlikely that the feature(s) may have been caused by artesian conditions related to flooding. For example, the M8 New Madrid earthquakes of 1811-1812 resulted in seismically induced liquefaction over an area approximately 160 km long and 80 km wide (Obermeier, 1995), and the 1886 Charleston earthquake caused liquefaction as much as 160 km from the Charleston area (Gelinas et al., 1994). In contrast, flooding-induced sand boils are restricted to a belt within 0.5 to 1.0 km from man-made or naturally-made levees (Obermeier, 1996; Li et al., 1996). The regional extent and distribution of liquefaction are governed by factors controlling the amount of strong ground shaking (earthquake magnitude, distance from fault plane, directivity, and topographic effects), as well as local geologic conditions (e.g., depth to groundwater, slope, geologic materials, and structure).

Stratigraphic Relations

Criterion: If surface deformation is present in unconsolidated granular deposits, then it may be related to seismically induced liquefaction or compaction.
Criterion: If stratigraphic evidence indicates that the feature was produced by a short, upward-directed hydraulic force, then it may be related to seismically induced liquefaction.

Criterion: If there is evidence of mobile granular material (e.g., sand dikes or sills), then the feature may be related to liquefaction.

Surface deformation produced by liquefaction is associated with unconsolidated late Quaternary basin-fill, fluvial and eolian deposits. Compaction-related differential settlement will be localized at abrupt thickness changes in the consolidating sediment, such as along the margins of depositional basins or across buried bedrock "highs" in the basin. Liquefaction-related features generally are associated with unconsolidated, saturated silty sand and sand deposits (fluvial, lacustrine or eolian) that are overlain by nonliquefiable strata. Gravelly sand deposits (<40-50% gravel) also are susceptible to liquefaction (Obermeier, 1995; 1996); whereas sediments with 15% or more clay content generally do not liquefy. Furthermore, deposits older than late Pleistocene, which usually are moderately to well consolidated, have lower susceptibility to liquefaction and settlement than younger, poorly compacted saturated deposits (Tinsley et al., 1985; Gelinas et al., 1994). Tokimatsu and Seed (1987) provide data on the amount of settlement that may be expected during strong ground motion, with volumetric strains being generally less than 5% in unconsolidated sediments. Assessing the stratigraphy of sediments beneath a site may allow a comparison of displacements observed at the surface and the expected amount of settlement.

In identifying a liquefaction-related feature, the feature's characteristics must be consistent with an upward-directed hydraulic force of short duration. For instance, the presence of angular rip-up clasts of a nonliquefiable stratum within an injected dike of clean sand suggest a sudden upward-projected force. In a trench exposure, most sand blows and lateral spreads have well-defined internal stratigraphic features, such as a general upward-fining sequence of vented dike material, that are useful for differentiating liquefaction features from other fault-related sedimentologic processes (Obermeier, 1996).

Liquefaction commonly is expressed as tabular sand dikes and sills at sites with a thick clay-rich cap or "top stratum" (Tuttle and Schweig, 1996). Sand dikes generally widen with increasing depth, whereas sills typically spread laterally at three preferential locations: (1) along the base of the top stratum, (2) along bedding planes and other horizontal planes of weakness in the cap, and (3) beneath dense, strong root mats (Obermeier, 1996). Sills also tend to form irregularly within thin beds of silt or sand between clay-rich beds. Finally, because dikes represent upwardly mobile sediments, horizontal bedding in the sand is absent.

Hydrologic Relations

Criterion: If surface deformation is present in deposits that are (or were) saturated, then they may be related to seismically induced liquefaction.

Liquefaction occurs in saturated, unconsolidated granular deposits where strong cyclic ground motions produce upward hydraulic forces. Typical hydrologic groundwater settings include low-lying areas (valleys, lagoons, and tidal regions) where groundwater is less than about 15 m in depth, and where deposits consist of saturated sands overlain by interbeds of fine-grained clays and silts. Saturated conditions are essential for liquefaction. Of course, hydrologic conditions vary through time, and the absence of present-day shallow groundwater does not preclude the possibility that liquefaction may have occurred under different groundwater conditions.

Scale and Extent of Feature(s)

Criterion: If the scale (e.g., length, width) of a fault is less than several hundreds of meters, then it may be related to liquefaction rather than tectonic surface rupture.

Criterion: If a fault extends to depth below any potentially liquefiable layer, then it likely is unrelated to liquefaction and may be related to tectonic surface rupture.

Criterion: If a fault extends laterally beyond the zone of deposits susceptible to liquefaction and/or compaction, then it likely is unrelated to liquefaction and may be related to tectonic surface rupture.

In plan view, the pattern of liquefaction-induced deformation is controlled primarily by two phenomena: orientation of the basin margin and/or orientation of adjacent free face exposures (e.g., channel margins). Most lateral spreads and settlement-induced ground cracks form subparallel to the basin margin and/or nearest topographic free face. There is no direct relationship of the pattern of liquefaction to regional or local structural grain and/or tectonic stress/strain field, although there may be an indirect relation where these features control basin geometry.

At a local scale, liquefaction features may exist as: (1) linear sand ridges as much as 1 km long, (2) curvilinear-shaped sand-filled craters up to 1 m in diameter, (3) lateral spreads with displacements as much as hundreds of meters; and (4) sand blows with diameters as much as 10 m. In contrast, settlement features generally exist as semi-circular depressions tens of meters wide and a few meters deep. The limited length of these features indicates that they are less continuous than most seismogenic tectonic faults, and are characterized with shapes inconsistent with tectonic faulting (i.e. curvilinear vs. linear). At local and regional scales, liquefaction-induced features commonly occur in groups with more than one type of feature present (i.e., sills, dikes, vented sediment, lateral spreads, and occasionally soft-sediment deformation structures).

3.3 Criteria for Identifying Nonseismogenic–Nontectonic Phenomena

3.3.1 Landslide Faults

Landslide-induced ground deformation produces a wide-range of geologic and geomorphic features that individually are very similar to those produced by surface faulting. Landslides and their general features are described in detail in Section 2.6.1 and by Cotton (Appendix A). Although highly diagnostic characteristics specific to landsliding exist, differentiating landslide faults from tectonic faults is most effective when a suite of landslide characteristics are considered collectively. Several *moderately diagnostic* criteria can be used to differentiate landslide-induced faults from tectonic faults, including:

- presence of a shallow, listric basal slide plane subparallel to topography
- Map pattern and sense of slip of boundary faults
- Lateral continuity of fault features
- High strain rates and anomalous recurrence intervals

These criteria are described briefly below.

Presence of a shallow, listric basal slide plane

Criterion: If a fault merges with a shallow, listric basal slide plane subparallel to topography, then the fault most likely is related to landsliding rather than tectonic surface rupture.

The presence of a shallow (<200 m) basal slide plane is a highly diagnostic feature that readily differentiates a landslide from tectonic faults and other nontectonic faults (excluding sackung and gravity glide phenomena). As stated previously in Section 2.6.1, landslides commonly produce discrete shearing of earth materials, including development of rupture surfaces within and at the boundaries of the slide mass. In general, the basal slide plane forms a continuous surface of shearing that may express fault characteristics, such as clay gouge, slickensides, lineations, and mullions. The basal slide plane may extend to the ground surface and/or will merge with the boundary faults. The basal slide plane is usually listric in geometry and subparallel to the mountainside or slope on which it forms. The basal slide plane can be identified and characterized through geologic mapping, geophysics, trenching and detailed borehole logging.

Map pattern and sense of slip of boundary faults:

Criterion: If a fault is related to other faults of different senses of slip that are consistent with a landslide model, then the fault probably is related to landsliding rather than tectonic faulting.

Within an ideal landslide model, faults bound the entire landslide block and their location and sense of movement can be estimated (Cotton, Appendix A). For instance, the upper part of a landslide usually is bounded by a head scarp that shows normal displacement, whereas the toe is associated with thrust or reverse faults and sinuous, bulging ridges that overrun the former topographic surface. In profile, these two opposing faults are connected by a basal plane of detachment, with the normal fault dipping downslope and the reverse fault dipping upslope. For example, a reverse fault present at the base of a slope should be suspected as a landslide plane if it decreases in dip with increasing depth, reverses dip direction, and becomes parallel to the overlying slope. Conversely, if a reverse fault at the base of a slope extends into the subsurface at increasingly greater dip with depth it most likely is not the result of landslide failure. Downslope movement of the landslide also is accommodated by strike-slip faulting along the lateral margins (Figure WC.3, in Cotton, Appendix A).

Lateral Continuity

Criterion: If a fault is greater than 3 km long, it probably is not related to landsliding.
Criterion: If a fault is discontinuous or less than 3 km long, it may be related to landsliding.

In plan view, virtually all landslides are less than 3 km long, and most are less than 1 km long in their longest dimension. Therefore, it follows that if an observed fault exceeds these dimensions, it cannot be related to landsliding and more likely is related to tectonic fault rupture or some other nontectonic process. In contrast, tectonic faults, especially well-developed faults, typically are continuous over distances greater than three kilometers. Thus, if a fault is greater than 3 km long, it probably is not related to landsliding. Conversely, if a fault is discontinuous or less than 3 km long, a potential landslide origin should be considered.

High Strain Rates and Anomalous Recurrence Intervals

Criterion: If a fault has an anomalously high strain rate for the regional or local tectonic setting (e.g., m/sec vs. mm/yr), then it is unlikely to be related to tectonic faulting and more likely to be related to landsliding or some other nontectonic process.

Criterion: If a fault has an anomalously short recurrence interval between surface rupturing events for the regional or local tectonic setting (e.g., years or tens of years vs. hundreds to thousands of years), then it may be related to landsliding.

Differences in strain rates and recurrence intervals also are potentially diagnostic criteria for differentiating landslide-induced faulting from tectonic faulting. Active landslide faults commonly have high strain rates (mm/yr to m/sec) in comparison to lower strain rates of most active faults (mm/yr). The high strain rates of landslide faults usually are associated with currently active landslides, and may not be present, especially with older, dormant landslides. Strain rates can be assessed by measuring the recorded offset of cultural objects, stratigraphic units, or geomorphic features over specified time intervals across a fault. Note that the time-averaged strain rates for large dormant landslides may approach the strain rates for active tectonic faults, eliminating the usefulness of high strain rates as a compelling differentiation criterion. In addition, landslides may have recurrence intervals between surface rupturing events that are anomalous in the regional or local tectonic setting. If a paleoseismic investigation documents a strain rate or recurrence interval that is higher and more frequent, respectively, than that expected from the tectonic setting, a nontectonic origin should be considered for the fault.

Inconsistency with Regional Tectonic Framework

Criterion: If a fault or series of faults are inconsistent with the regional tectonic framework, and is more consistent with downslope movement, then they may be related to landsliding.

It is important to evaluate the sense of fault slip and fault orientation with respect to the regional tectonic framework and surrounding topography. For example, does the observed sense of slip and fault orientation agree with the predominant tectonic setting and/or stress regime of the region? If the style of faulting and/or orientation of a fault is not consistent with the contemporary tectonic and structural setting, it may suggest that the fault has a nontectonic origin. This evaluation is particularly useful for poorly developed landslides with limited exposure of only one or two of the diagnostic boundary faults.

3.3.2 Sackungen

Sackungen-like features may be expressed geomorphically as a variety of features: downhill-facings scarps, uphill-facing scarps, double-crested ridges, troughs, and notches on ridge axes (Section 2.6.2; McCalpin, Appendix A). Unfortunately, these features may also form by (1) displacement on tectonic faults (Section 2.4), (2) earthquake-induced slope failure (Section 2.5.3), or (3) nonseismic gravity failure (Section 2.6.1). These many possible mechanisms make the identification of sackungen difficult. In this section, we concentrate on the characteristics that allow differentiation between sackungen and tectonic faults.

There are numerous phenomena produced by primary and secondary surface fault rupture that are geomorphically similar to phenomena produced by nonseismic sackung processes. Furthermore, there are no highly diagnostic criteria to differentiate sackungen from tectonic faults that provide a high degree of confidence in determining the origin of a fault. The identification of sackungen, therefore, requires the integrated use of several moderately diagnostic criteria. The greater the number of criteria that support a sackungen origin, then the higher degree of confidence in the interpretation that a fault may be related to a sackungen process. Moderately diagnostic criteria include:

- Regional spatial pattern of sackungen
- Regional temporal pattern of deformation
- Regional/local length
- Local number and continuity of scarps
- Local topographic position and relief
- Feature-specific scarp aspect ratio
- Feature-specific displacement history
- Feature-specific sense of slip
- Feature-specific fault zone structure

These criteria are described below and discussed in greater detail by McCalpin (Appendix A).

Regional Spatial Pattern of Sackungen

Criterion: If a feature is spatially associated with other sackungen-like features that are not oriented similar to known tectonic faults, then the feature likely is related to sackung processes.

The presence or absence of similar sackungen within the region may provide data to assess the origin of the fault-like feature. For example, if sackungen-like features are present within only a limited zone coincident with a tectonic fault or fault zone, then it is more likely that they are related either to surface rupture along the fault or to strong ground motions generated by the fault. However, it is also possible that sackungen may form preferentially along pre-existing structural discontinuities or zones of weakness, and this spatial association is not indicative of the seismogenic nature of a fault in the present tectonic regime. If sackungen-like features of similar dimensions and characteristics are distributed throughout a large area with no apparent spatial relation to known tectonic faults, then an interpretation of a nontectonic origin for the sackungen-like features may be more reasonable.

Regional Temporal Pattern of Deformation

Criterion: If several sackungen-like features in an area developed simultaneously, then the features are more likely to be related to tectonic faulting.
Criterion: If several sackungen-like features in an area developed independently of one another, then the features may be related to nonseismogenic sackung processes.

As with other slope failures that may be generated by either seismic or nonseismic processes, the timing of formation of several sackungen-like features within a region provides important information for determining their origin. In other words, if the formation of several sackungen-like features within a region occurred simultaneously, then an interpretation of a coseismic surface faulting origin may be valid. Alternatively, if movements of several sackungen-like features within a region are not simultaneous, then it is less likely that all of the features were formed as a result of a single, regional influence such as surface faulting during a specific earthquake. In addition, the spatial distribution of sackungen-like features having comparable or noncomparable movement histories may yield information to assess feature origin. For example, if sackungen-like features having similar movement histories are separated widely within a region that also contains sackungen with different movement histories, then an interpretation of a nonseismic origin may be appropriate. In short, both the temporal and spatial patterns of sackungen-like features may provide data to assess the feature's origin. This approach is similar to that used for identifying seismically induced (1) paleoliquefaction features in the central and eastern United States (Tuttle and Schweig, 1996; Obermeier, 1996), (2) subduction-related subsidence in the Pacific Northwest (Carver and McCalpin, 1996), and (3) landsliding adjacent to Lake Washington (Jacoby et al., 1995). This criterion obviously entails identification of many features and development of detailed chronologies for each, which may be expensive and logistically difficult.

Regional/Local Length

Criterion: If sackungen-like features are continuous over lengths of more than about 3 km, they are more likely to be related to a primary tectonic fault.

Sackungen generally are less than a few kilometers (< 3 km) long, and tend to be discontinuous. Tectonic faults that exhibit evidence of repeated surface rupture, in contrast, typically are longer than 3 km. However, historical earthquakes show that tectonic surface ruptures also may be limited in length. Therefore, if sackungen-like features are continuous over lengths of more than about 3 km, they are more likely to be related to a primary tectonic fault. In contrast, the features may be related to either seismic or nonseismic processes if they are present over a short length.

Local Number and Continuity of Scarps

Criterion: If sackungen-like features are aligned and are constrained within a narrow zone along a tectonic fault, then it is more likely that they are related to surface rupture than to gravity failure.

Sackungen commonly have numerous short, discontinuous arcuate scarps, whereas tectonic faults generally have scarps in a narrow zone associated with the fault trace. If sackungen-like features are aligned and are constrained within a narrow zone along a tectonic fault, then it is more likely that they are related to surface rupture rather than to gravity failure. In contrast, many discontinuous, broadly distributed scarps on or near a ridge crest more likely represent a series of sackungen.

Local Topographic Position and Relief:

Criterion: If sackungen-like features are present *in areas of low relief or can be traced continuously across slopes having opposite aspects, they likely are not related to sackung processes.*

Sackungen require relief for formation, such that areas with low relief generally do not contain sackungen. Because surface-rupturing faults may be present across areas with little or no local relief, sackungen-like features in areas of low relief more likely represent tectonic surface rupture, or some other nontectonic phenomena. In addition, the position of sackungen-like features across topography provides a moderately diagnostic criterion. For example, if a feature crosses slopes with different aspects, then a gravitational sackungen origin is not likely.

Feature-specific Scarp Aspect Ratio

Criterion: If the ratio between the length (L) and height (H) of a scarp is small (L:H $<10^4$), then the feature may be related to sackungen processes.

The ratio between the lengths and heights of scarps associated with sackungen are generally smaller for sackungen ($<10^4$) than for tectonic faults ($>10^4$) (McCalpin, Appendix A). For example, tectonic surface rupture may produce a 2-m-high scarp that is tens of kilometers long (Wells and Coppersmith, 1994), although it is unlikely that a 2-m-high scarp produced by sackung processes will be as continuous. In other words, sackungen scarps commonly are shorter than the same height scarps produced by tectonic surface rupture.

Feature-specific Displacement History

Criterion: If a sackungen-like feature is produced by continuous movement, then it likely is a result of sackung processes.

Sackungen may form by continuous or episodic movement, but fault scarps are formed by episodic, sudden events (McCalpin, Appendix A). Thus, if information is available for determining the timing and recurrence of feature development, such as through a program of shallow trenching, features produced by continuous movement along near-surface faults probably are a result of sackung processes. If the features are formed by episodic movement on the near-surface faults, then another criterion likely is needed to assess the feature's origin.

Feature-specific Sense of slip

Criterion: If a sackung-like feature contains evidence of predominantly lateral slip, then it more likely is related to tectonic faulting or some other nontectonic process.

Criterion: If the sense of slip associated with a sackung-like feature is not consistent with the style of tectonic faulting in a region, then the feature may be related to sackung processes or some other nontectonic process.

Sackungen typically are associated with normal dip-slip displacement, which may be diagnostic if faults in a region are dominated by strike slip or reverse displacement. Minor lateral movement may occur along faults associated with sackungen, which may be consistent with gravitational processes when can be compared to local slope orientations. If the regional tectonic framework suggests the presence of lateral-slip faults, then other criteria may be needed to address whether a feature is related to sackung or tectonic processes. Because sackungen-like features are produced by gravitational processes, the presence of reverse faulting strongly indicates a non-sackungen origin. On the other hand, evidence of reverse faulting at the base of a slope may be indicative of slope failure (i.e., landsliding) or tectonic faulting. Again, the use of multiple criteria in determining fault origin is extremely helpful.

Feature-specific Fault Zone Structure

Criterion: If the near-surface fault zone morphology consists of a sharp upper contact, and a lower boundary that is transitional to underlying bedrock, then the feature may be related to sackung processes.

McCalpin (Appendix A) notes that sackung-related fault zones typically contain a discrete zone of pulverized, brecciated rock and fault gouge that transitions to the undeformed underlying bedrock. The fault zone morphology is asymmetrical, with a sharp upper contact and gradually less brecciation of bedrock with depth. In contrast, tectonic fault zones in bedrock commonly are symmetrical, with a central core of fault gouge, bordered by progressively less pulverized bedrock outward in both directions from the fault gouge. However, all tectonic fault zones in bedrock are not symmetrical, suggesting that this is not a universally diagnostic criteria.

3.3.3 Large-Scale Gravity Structures Above Salt or Shale

The process of gravity spreading/gliding consists of the large-scale translation of weak mobile rocks along a basal detachment plane under low tectonic or gravitational stresses. Gravity glide structures are similar to large, low-gradient landslides in which the underlying rock type (salt or shale) serves as the décollement or slip surface. Gravity spreading is the vertical thinning and horizontal widening of rock masses driven by a topographic gradient, and is regarded as a main trigger for growth faulting and diapirism along passive margins (e.g., the Gulf of Mexico region). Growth faults related to gravity spreading/gliding are discussed in Section 2.6.3 and by Huntoon (Appendix A) and Vendeville (Appendix A). The surface rupture expressions of growth faults and faults along and within gravity-glide blocks are structurally and geomorphically similar to tectonic faults. There are no highly diagnostic criteria available to conclusively differentiate gravity glide structures from seismogenic-tectonic faults. However, through consideration of multiple moderately diagnostic criteria, it may be concluded that a features is more likely related to gravity spreading/gliding than tectonic processes. The following criteria may be used for delineating large-scale gravity spreading/gliding structures:

- Presence of salt or shale strata
- Tectonic setting
- Hydrologic setting
- Geologic structure and downdip extent
- Spatial associations

Presence of Salt or Shale Strata

Criterion: If a fault is present in an area that does not contain salt or shale, then it cannot be related to gravity spreading/gliding.

Criterion: If a fault is present in an area underlain by salt or shale, then it may be related to gravity spreading/gliding.

The presence of salt or shale strata within a region is required for the development of large-scale gravity spreading/gliding phenomena. The ability of salt and overpressured shale to behave viscously leads to structural instabilities and continuous deformation. Gravity spreading/gliding also may remain active for as long as a few million years. If salt or overpressured shale is present near a fault, one should consider the possibility that the fault is related to gravity spreading/gliding. Unfortunately, the presence of shale is less diagnostic than the presence of salt, because shale is more common and may be present in many tectonic settings. Also, the presence of salt/shale is not effective in conclusively differentiating between nontectonic salt/shale, gravity-driven faults and tectonic faults. Typically, as with other nontectonic features described above (i.e., landslides and sackungen), the degree of confidence in the assessment of fault origin is higher when many criteria are used in the analysis.

Tectonic Setting

Criterion: If a fault is present in an area containing salt or shale and currently or formerly characterized by an extensional tectonic setting, then it may be related to gravity spreading/gliding.

Jackson and Vendeville (1994) suggest that salt basins with persistent deformation structures, such as growth faults and gravity glides, are associated with an active or inactive divergent continental margin or rift zone. The rift zones are characterized either by (1) thick-skinned extension that involving basement rock and sedimentary cover, (2) thin-skinned extension involving only the sedimentary cover, or (3) a combination of both of these processes. The underlying structures need to be assessed for their consistency for originating within an extensional setting. For instance, the presence of salt and normal faulting for a site located in the Michigan Basin might suggest that the fault is unrelated to gravity glide processes; whereas similar conditions identified in the Gulf Coast region might strongly suggest that the fault is associated with nontectonic salt/shale-induced deformation. In both instances, further assessment is necessary for conclusively determining the origin of the fault.

Hydrologic Setting

Criterion: If a fault is present in an area containing high fluid pressures within subsurface shale deposits, then it may be related to gravity spreading/gliding along a shale décollement.

Criterion: If a fault is present in an area that lacks high fluid pressures within subsurface shale deposits, then it probably is not related to gravity spreading/gliding along a shale décollement.

High fluid pressures are required to produce shale gravity glides. Movement along a shale décollement is induced by high fluid pressures within the shale deposits. Elevated fluid pressures in shale deposits occur from rapid sedimentation and compaction of overlying sediments. Overpressured shale deforms and translates along a shallow-dipping detachment under low tectonic or gravitational stress, and continues deforming while overpressured conditions exist.

Geologic Structure and Down-Dip Extent

Criterion: If a fault is listric into a basal glide plane at a depth within a shale horizon, then it may be related to gravity spreading/gliding.

Criterion: If a fault lies in an area containing a horst-graben complex oriented perpendicular to the regional slope, then it may be related to gravity spreading/gliding.

Criterion: If fault blocks are characterized by greater block rotation than those produced by tectonic normal faults, then they may be related to gravity spreading/gliding.

Criterion: If a fault is in an area containing rollover folds and footwall uplifts that are more localized than those associated with basement-involved normal faults, then it may be related to gravity spreading/gliding.

Gravity spreading/gliding processes translate blocks above a weak detachment plane or décollement layer, with relatively undeformed blocks separated by listric normal faulting (Vendeville, Appendix A; Huntoon, Appendix A). Listric normal faults typically form in the upper part of the spread/glide, and sole into a dipping basal slide plane that may be as much as 7 km deep. The basal slide plane deforms the ground surface updip as a series of listric normal growth faults, and downdip as folds and thrust faults. The detachment fault bounds the trailing edge of the plate, with normal faults deforming the block itself, and thus producing a horst-graben complex with faults that strike perpendicular to the direction of sliding. In vertical section, the listric normal faults dip downslope, with their dip decreasing with depth to the décollement. They generally extend less than 10 km laterally and in many cases terminate against diapirs. Normal faulting may allow for diapiric rise of salt, and thus gravity spreads/glides commonly are associated with regions containing salt diapirs.

Faults related solely to large-scale gravity deformation differ from basement-involved, tectonic faults in that they involve only the few uppermost part (<7 km) of the crust above the shale horizon, whereas tectonic faults typically extend to greater crustal depths. In addition, because fault spacing partly depends on the thickness of the faulted interval (the brittle overburden overlying the salt or shale layer), the spacing of normal faults associated with salt or shale gravity spreading/gliding may be less than that of tectonic faults.
Fault blocks produced by gravity spreading/gliding tend to experience greater block rotation than those produced by tectonic normal faults. Because gravity glide faults affect only the shallow sediments, block deformation such as rollover folding of the hangingwall or uplift of the footwall also are accommodated at depth by flow of the source layer, rather than by deformation of the basement. This results in rollover folds and footwall uplifts that are more localized than those associated with basement-involved normal faults.

Spatial Associations

Criterion: If a fault is spatially associated with salt or shale diapirs, then it may be related to gravity spreading/gliding.

Gravity spreads/glides commonly are spatially associated with salt and shale diapirs. The extension of the overlying suprasalt sediments along the basal detachment promotes the growth of salt and shale diapirs. Therefore, the presence of salt or shale in the shallow subsurface, and of other features that are characteristic of salt or shale deformation (Section 2.6.3; Vendeville, Appendix A; Huntoon. Appendix A), may also suggest the presence of faults related to gravity glide structures.

3.3.4 Salt or Shale Diapirs

Sedimentary diapirs usually consist of upwardly mobile salt or shale that causes doming, warping and normal faulting of overlying sedimentary deposits. Salt and overpressured shale are weak rocks that may deform and flow plastically under low tectonic or gravitational stress. Deformation structures produced by rising diapirs are remarkably similar in appearance to tectonic faults, making a determination of the origin of the structure difficult. Salt and shale diapirs and the mechanisms responsible for their origin are described in Section 2.6.3 and by Huntoon (Appendix A) and Vendeville (Appendix A). Because diapirs are triggered primarily by large-scale gravity spreading, they commonly are associated spatially with gravity glide features (as described in Section 3.3.3).

There are no *highly diagnostic* criteria that definitively identify a fault related to salt or shale diapirism, and there are only two criteria (described below) that demonstrates a fault cannot be related to salt or shale diapirism. In fact, the presence of salt or overpressured shale in a region raises the possibility that any observed fault may be related to salt or shale diapirism.

Stratigraphic relations

Criterion: If a fault is present in an area that does not contain salt or shale, then it cannot be related to diapirism.

Criterion: If a fault is present in an area underlain by salt or shale, then it may be related to diapirism.

If salt or overpressured shale is present near a fault, one should consider the possibility that the fault may be related to diapirism. For instance, the presence of subsurface salt or salt ridges exposed near the ground surface, and within a region of surface faulting or folding, suggests that the faulting may be related to diapirism. Salt behaves viscously under low gravitational stresses and commonly records structural instabilities and long-term deformation generally attributed to diapiric processes. Shale mobility requires conditions of elevated fluid pressures. The presence of shale is less diagnostic than the presence of salt, because shale is more common and may be present in many tectonic settings. Thus, to confirm that the origin of a fault is related to shale diapirism requires the identification of other criteria (see below). In general, however, in regions where salt and shale mobility has been documented or is suspected, normal faulting may be related to diapirism.

Tectonic Setting

Criterion: If a fault extends to depth below the horizon or salt or shale, then it cannot be related to salt or shale dipirism.

Criterion: If a fault has a ratio of lateral to vertical slip of greater than 1.0, then it more likely is related to tectonic faulting than to salt or shale diapirism.

Virtually all faults related to salt or shale diapirism have normal dip-slip displacement confined to the shallow crust above the horizon of salt and/or mobile shale.

Moderately Diagnostic Criteria

Diapir-related faulting also may be identified by the integrated use of several moderately diagnostic criteria, such as local stratigraphy, seismicity, structure, scale, and duration of deformation. Integrating these criteria with the diagnostic stratigraphic criteria described above provides a greater degree of confidence in determining the origin of a fault. Moderately diagnostic criteria associated with diapirs include:

- Tectonic setting
- Evidence of local erosion or deposition
- Low levels of seismicity
- Structural relations
- Scale of features

Tectonic Setting

Criterion: If a fault is present in an area underlain by salt or shale substrata, and the area is currently or formerly characterized by an extensional tectonic setting, then it may be related to salt diapirism.

Vendeville (Appendix A) and Jackson and Vendeville (1994) show that diapirs are observed most commonly in two kinds of extensional tectonic settings (Section 2.6.3):

- divergent continental rifts or basins affected by thick-skinned extension, where salt or shale was deposited before rifting began

- passive continental margins and basins where salt or shale was deposited after rifting ceased. Extensional tectonics in these regions is thin-skinned, affecting only the salt/shale layer and overlying rocks. Examples include the Gulf of Mexico and the southern Atlantic margins of the United States

Characterizing the current and former tectonic setting of the site is key for assessing a region's susceptibility to diapirism. For example, the Gulf Coast region of the United States underwent an earlier episode of tectonic rifting and graben development, followed by salt and shale deposition, which collectively provide favorable conditions for salt/shale-induced deformation (Vendeville, Appendix A). Salt-related deformation may span tens of millions to hundreds of millions of years, and thus is commonly evidenced in the stratigraphic record.

Evidence of Local Erosion or Deposition:

Criterion: If sedimentary deposits above salt deposits have substantial thickness variations, these variations may be related to active diapirism, and thus faults may be related to active diapirism.

Patterns and thickness variations of surficial deposits and sedimentary rocks are useful for delineating areas undergoing salt/shale diapirism. For example, where salt and shale diapirs are ascending at rates that exceed local depositional rates, sedimentary deposits are thin at the region of maximum uplift and are thick near the topographic base of the structure. Where variations in stratigraphic thickness across a domed region are less distinct, the diapir may be deforming at rates similar to or less than sedimentation rates in the region. In this case, it may be difficult to attribute thickness variations to diapir-related ground deformation. Conversely, evidence of local erosion and deposition should be interpreted with caution in regions where salt bodies are undergoing regional tectonic shortening. In these settings, thickening and thinning relationships might be incorrectly attributed to diapiric processes, rather than to tectonic folding or faulting.

Low Level of Local and Regional Seismicity

Criterion: If moderate to large magnitude earthquakes occur below the depth of salt or shale within a region of active diapirism, the seismicity is not related to salt or shale diapirism.

If a fault is being investigated in a region with known salt/shale deformation, the regional seismicity pattern is useful in assessing if the fault is tectonic or nontectonic. In general, diapirs are incapable of producing moderate or high levels of seismicity and generally are seismically quiescent structures. In regions with high rates of salt tectonism (i.e., Gulf of Mexico), salt-related deformation is occurring with little or no seismicity. If moderate to large magnitude earthquakes with epicenters deeper than the depth of the salt or shale substrata (e.g., about 7 km in the Gulf of Mexico region) are present within a region of active diapirism, the seismicity probably is associated with the regional tectonic setting rather than an active salt or shale diapir.

Structural Relations

Criterion: If fold axes in a region underlain by salt or shale are randomly oriented or inconsistent with the regional and local tectonic setting, then they may be related to salt or shale diapirism.

Extensive diapirism may produce folds that may appear to be related to tectonic deformation. However, diapir-related folds are dissimilar to tectonic folds. For example, folds induced by rising salt diapirs should show a pattern consistent with an underlying piercing structure (i.e., doming) or a random orientation of fold axes similar to soft-sediment deformation. In contrast, syntectonic folding of similar-aged deposits usually yields consistent fold axes orientations. Furthermore, when warped deposits can be traced to the deformation source (i.e., contact between the salt diapir and wall rocks) the presence of salt at the fault contact provides additional evidence that the deformation is related to diapirism.

Scale of Features

Criterion: If the length of a fault in a region underlain by salt or shale is more than 2 km, then it more likely is related to seismogenic-tectonic *faulting.*

Structures produced by diapirs typically are less than 2 km in length. The shorter length of salt and shale diapir-related faults is controlled partly by the shallow depth of the structures and the thickness of the sedimentary package. Where faults with lengths greater than 2 km are identified in regions characterized by salt/shale deformation, an investigator should consider these faults as non-diapiric structures that are related either to other nontectonic phenomena (i.e., gravity glides) or seismogenic-tectonic phenomena.

3.3.5 Postglacial Faults and Glaciotectonic Deformation Features

Criteria for recognizing postglacial faulting and differentiating these faults from glaciotectonic or other nonseismogenic-nontectonic faults have been described by various authors. Based on criteria established by Mohr (1986) and revised by Adams et al. (1993) and Fenton (1991, 1994), Fenton (in Appendix A) describes the following general criteria to identify postglacial faults:

- Geologic and Stratigraphic Criteria
    1. Timing of Faulting
    2. Evidence for Continuous Deformation versus Episodic Deformation
    3. Style of Faulting
- Geomorphic Criteria
    1. General Geomorphology
    2. Scarp Aspect Ratio
- Structural Criteria
    1. Dip of Fault Plane and Depth of Deformation
    2. Orientation with Respect to the Contemporary Tectonic Stress Field
    3. Rupture Complexity and Secondary Deformation
- Associated Deformation
    1. Contemporaneous Association with Seismically-Induced Features
- Geophysical Criteria
    1. Association with Contemporary Seismicity
    2. Spatial Association with Areas with High Rates of Glacio-Isostatic Uplift.

Fenton (in Appendix A) notes that in combination several diagnostic criteria can be used to differentiate between tectonic and nontectonic faults in former glaciated regions, but that a more difficult problem is that of differentiating between faulting that results from transient loading/unloading stresses, i.e., glacio-isostatic (postglacial) faulting, and tectonic faulting that results from long-term regional lithospheric stresses. Documented cases of postglacial faulting have involved reactivation of pre-existing shear zones, faults and fractures. Movement on the larger of these reactivated faults was seismogenic, as shown by the spatial and temporal association with seismically-induced deformation, including liquefaction and landsliding. In contrast to tectonic faults in more seismically active regions, these faults show evidence for only a single event. In some cases, displacement occurs along shear zones that are characterized by primarily ductile shear deformation and there is little or no evidence of a history of brittle fault displacement.

Demonstrating the lack of a history of brittle deformation and evidence for a single event in the contemporary tectonic regime, provide possible criteria for differentiating faulting related to transient glacioisostatic processes from tectonic faults that may be characterized by long recurrence intervals. Differentiating between these two types of faults is key to addressing the capability and seismic hazard posed by the fault.

Other researchers (e.g., Croot, 1987; Banham, 1988; Broster, 1991), who have focused their research on glaciotectonic structures, have developed the following criteria to identify these types of nontectonic structures.

- Time/space association with glaciogenic deposits and landforms
  1. Termination at base of overlying till
  2. Infilling of glacially-derived sediment
  3. Glaciogenic hydraulic dikes
  4. Associated drumlins, eskers, terminal moraines
- Formation in an ice-marginal environment and absence of evidence for a structural prime-mover apart from ice
- Development directly below (usually within 30 m) of the contemporary ground surface
- Structural orientation relating to applied or relaxed stresses consistent with local direction of glacial movement
  1. Upslope vergence of structures
  2. Regularity of orientation of structures and dominant sense of vergence
  3. Laterally progressive development of structural style/intensity of deformation

In general, differentiating nontectonic faults from tectonic or potentially tectonic faults is most effective when a suite of characteristics are considered. The following discussion, which is organized to address the question of whether surface deformation and faulting in a formerly glaciated region represents postglacial faulting (i.e., displacement along a pre-existing tectonic fault or fracture) or nontectonic faulting (e.g., glaciotectonic, periglacial), expands on some of the criteria outlined above.

Downdip geometry and extent

Criterion: If faulting and related deformation can be demonstrated to occur only in the upper tens of meters of the stratigraphic section, it is not the result of primary tectonic deformation.

Criterion: If a fault exhibits a listric geometry and shallows into a bedding plane or décollement at shallow depths (within 10s of meters), it may be a shallow stress-release feature or a glaciotectonic structure.

In cases where the downdip extent and geometry of the faulting can be assessed with some confidence, the above criteria may be used to demonstrate a nontectonic origin. Tectonic faults by definition are the result of crustal scale processes. Nontectonic mechanisms typically involve only the upper tens of meters of the stratigraphic section. Glacially-transported blocks, which can be a few meters to several tens of meters in thickness and of the order of hundreds of meters to over a kilometer in length (Schroeder et al., 1986), may appear to be intact bedrock at the outcrop scale. However, structural and/or stratigraphic relationships at a more local or regional mapping scale that can be used to show their limited downdip and lateral extent. Where observed in cross section, shallow pop-ups typically are listric and converge with shallow décollements at depths of a few meters. Glaciotectonic faults also commonly sole into shallow basal thrusts. The underlying undeformed strata confirm that these features are not caused by deep-seated crustal deformation. Deformation related to periglacial processes and ice-melting also are shallow, being related to the depth of freezing and the thickness of remnant blocks of ice, respectively.

Regional Context and Timing of Faulting

Criterion: If observed faulting occurs within the former ice limits of the region and if faults have demonstrable movement since the disappearance of the last ice sheet within an area of concern, they may be postglacial (glacio-isostatic) faults.

Criterion: If fault scarps show evidence of glacial modification (e.g., moulding or plucking of the scarp face), they may be the result of glaciotectonic processes.

Criterion: If faulting occurs in regions characterized by significant glacio-isostatic rebound as evidenced by uplifted shorelines, geodetic or gravity measurements, glacio-isostatic postglacial faulting should be considered.

By definition, postglacial (glacio-isostatic) faulting may occur following the disappearance of ice cover within the region of concern. The faults generally displace glacial and late-glacial deposits, glacial surfaces or other glacial geomorphic features. Postglacial glacio-isostatic faulting commonly displaces postglacial stratigraphy and/or geomorphic features, but does not need to cut younger features. Fault scarps and rupture planes expressed in bedrock commonly show no signs of glacial modification, such as striations or ice-plucking. Limited glacial modification, however, may be present on scarps that are of late-glacial or interglacial age. Glaciotectonic deformation, however, can only occur while ice is present and actively advancing or retreating. Establishing the timing of faulting, therefore, may prove useful in distinguishing between postglacial and glaciotectonic faulting.

Evidence for Continuous Deformation versus Episodic Deformation

Criterion: If there is evidence for discrete, single-event displacements, coseismic postglacial faulting may have occurred.

Criterion: If there is structural or stratigraphic evidence for continuous deformation, nontectonic fault mechanisms, such as glaciotectonic or ice stagnation processes, should be considered.

Tectonic fault scarps are formed by episodic, sudden events. Limited paleoseismic trenching investigations along a few postglacial fault scarps in Fennoscandia, Scotland, and eastern Canada, show that these scarps were produced by a discrete, single postglacial event. Trench exposures revealed only one colluvial wedge and trench stratigraphy showed uniform offset regardless of age. Structural or stratigraphic evidence for continuous deformation commonly indicates that the scarp did not form coseismically as the result of surface fault rupture. However, 'event scarps' could also have been produced by nontectonic, nonseismogenic processes (e.g., landslides, sackungen, collapse due to ice melt).

Glaciotectonic features, in contrast, are more likely the result of continuous deformation and, although they may result from pulses or surges of glacial movement, they do not produce the stratigraphic and structural relationships that are indicative of sudden coseismic rupture. They may exhibit a laterally progressive development of structural style and intensity of deformation that is consistent with the advancement and retreat of the ice sheet.

Map Pattern and Style of Faulting

Criterion: If a fault is steeply dipping with reverse displacement in a glaciated region, postglacial glacio-isostatic faulting should be considered.

Criterion: If both compressional and extensional deformation are observed within the same general area of a glaciated region, glaciotectonic processes should be considered.

Criterion: If a fault is related to other faults of different senses of slip that are consistent with an ice-push model, then the fault may be related to ice push rather than seismogenic-tectonic or isostatic postglacial faulting along a pre-existing fault.

Postglacial faults that have been exposed in trenches generally are steeply dipping and exhibit nearly pure dip-slip displacement. Glaciotectonic faulting, in contrast, is more variable, showing a wide range of sense of displacement and dip. Models have been developed for glaciotectonic faulting (e.g., Banham, 1988; Croot, 1987; Schroeder et al., 1986; Derbyshire and Jones, 1980) that show the expected relationship between compressional and extensional features that would result from ice push or the entrainment of a block of frozen ground or bedrock slab by an advancing glacier.

General Geomorphology
Criterion: If scarp faces show signs of glacial modification, it cannot be formed by postglacial faulting.

Criterion: If glacial geomorphic features, such as flutes or striations, are displaced across a fault scarp, the fault scarp may have formed by postglacial faulting.

Criterion: If a fault in glaciated terrain is a kilometer or more in length and has a roughly continuous surface trace, it is more likely to be a postglacial fault than a glaciotectonic fault.

Criterion: If surface deformation parallels the former ice front, it is likely to be the result of glaciotectonic processes.

Criterion: If surface deformation is spatially associated with drumlins or a nearby drumlin field, a glaciotectonic process is more likely.

Geomorphic modification of a scarp provides evidence for the timing and surficial processes that may have influenced the development of a scarp. Differential erosion, ice plucking, or meltwater erosion may result in the development of scarps that subparallel pre-existing faults, fractures or bedding/schistosity planes. Careful examination of stratigraphic and structural relationships may be required to document erosional versus tectonic origins. The continuity and sense of deformation that can be inferred from the geomorphic expression of a feature of concern provide key information related to scale, style of deformation, and deformation mechanisms that may be inferred. Studies of the morphology and internal structure of drumlins (e.g., Zelčs and Dreimanis, 1997) provide evidence of multiphase glaciotectonic and possibly subglacial glaciofluvial genesis.

Feature-Specific Scarp Aspect Ratio
Criterion: If a postglacial fault exhibits anomalously low scarp length to scarp height ratios, the fault movement may be due to glacio-isostatic rather than tectonic processes.

For most tectonic faults, the ratio of overall length of the feature to displacement, which is a function of the strength of the rock prior to fault rupture, is between 10^4 and 10^5 (Scholz, 1990). Glacio-tectonic features commonly exhibit ratios of less than 10^3.

Structural Criteria
Criterion: If a fault expressed in bedrock is steeply dipping with reverse displacement, and is characterized by a consistent amount and sense of throw along strike, it may be a postglacial fault.

Criterion: If a fault is listric with depth or dies out in a shallow décollement, it is more likely to be a glaciotectonic or stress release pop-up structure, rather than a major postglacial fault.

Criterion: If a bedrock fault can be traced as a discrete fault into overlying unconsolidated materials plane (rather than more distributed fold deformation), it is more likely to be related to postglacial faulting than glaciotectonic processes.
Postglacial faulting commonly utilizes pre-existing faults, fractures, and shear zones, many of which may have been inactive for considerable periods of time (Eliasson et al., 1991). Most postglacial faults exhibit relatively simple geometries and a

consistent sense of displacement along the entire fault length. Structural complexities that are observed along some of the larger postglacial fault ruptures are a function of the geometry of pre-existing faults and fractures with secondary deformation occurring at fault bends.

In contrast to postglacial faults, glaciotectonism results in more heterogeneous deformation that is characterized by a lack of consistency in the sense and amount of throw along strike. Glaciotectonic deformation typically results in broad zones of deformation comprising multiple fault planes, with both normal and reverse displacements.

Where faulting involves both bedrock and overlying sediments, postglacial faults typically can be traced as a discrete plane in both materials. In contrast, glaciotectonic faults commonly can not be traced upwards into the unconsolidated sediments as a discrete plane, but may be expressed as folding within these deposits.

Orientation with Respect to the Contemporary Tectonic Stress Field
Criterion: If a reverse fault scarp is oriented orthogonal to the direction of the present maximum horizontal compressive stress, it may be a postglacial or a shallow stress relief structure, such as a pop-up.

Criterion: If the orientation and style of faulting is not consistent with the contemporary tectonic stress field, a nontectonic process is more likely.

Criterion: If different stress directions are evidenced by structural data at a single locality, a glaciotectonic origin is more likely.

It is important to evaluate the sense of fault slip and fault orientation with respect to the regional tectonic framework. Comparison to stress indicators in the region, including earthquake focal mechanisms and borehole breakout data, may provide information on whether a feature is likely to have slipped in response to tectonic stress. Tectonic structures should show a consistent relationship to the contemporary regional tectonic stress field. Glaciotectonic structures develop in response to local stress conditions that result from ice movement and resulting hydraulic conditions. Understanding the glacial history and particularly the direction and locations of ice movement is equally important to evaluating faulting mechanisms that result from ice-push, hydrostatic jacking, or ice-melting. Evidence for varying strain directions within single outcrops or exposures at a single locality is may be indicative of shifts in ice movement direction during either a single ice advance or multiple ice advances. Detailed kineto-stratigraphic analysis (Berthelsen, 1978; Hicock and Dreimanis, 1984) is a tool for evaluating the strain history at a site.

3.3.6 Stress Release Faults

The release of stress may produce a range of faults, including large, crustal-scale fault displacements. Stress release may result from the formation or melting of ice sheets to human engineering activities. A summary of the characteristics of the large-scale postglacial faults is given in Section 2.6.4, and diagnostic criteria for the identification of these faults is given in Section 3.3.5 above. Small to moderate earthquakes may be triggered by human-induced stress release mechanisms, such as quarrying and reservoir drawdown (Section 2.6.5). The added effects of these human activities serve mainly to augment or disrupt the pre-existing tectonic conditions (Yeats at el., 1996), and may produce surface deformation such as fault rupture and/or folding. Previous summaries of triggered seismicity related to human engineering activities are provided by Judd (1974), Milne (1976), and Simpson (1986). There are at least three moderately diagnostic criteria that can be used to differentiate human-induced stress release faults from seismogenic-tectonic features:

- Spatial and temporal associations
- Geologic framework
- Seismologic associations

Spatial and Temporal Associations

Criterion: If the spatial pattern of stress release faults coincides with the pattern of human engineering activities (e.g., mining, quarrying), the features may be nontectonic.

Criterion: If the temporal pattern of stress release faults coincides with the timing of human engineering activities (e.g., mining, quarrying), the features may be nontectonic.

Earthquakes triggered by human activities have a spatial and temporal association with pre-existing geologic discontinuities and the timing of the activities (i.e., rapid water-level changes in a reservoir). Human-induced stress release faults involve quarrying, surface and subsurface mining, and filling or drawdown of large reservoirs. The location and timing of these activities, as determined by reviewing historical land use maps, photographs (aerial and land-based), and/or written records, may provide a basis for assessing the origin of stress release faults. If the faults are present in areas of historical mining, then there is a possibility that the faults are related to mining activities. Given sufficient information on the timing and locations of historical mining activities, it may be shown conclusively that the faults are related to historical land use activities. Thus, assessment of historical records and subsurface conditions may provide highly diagnostic information for evaluating a fault's origin.

Geologic Framework

Criterion: If a fault is within a region having a pre-existing natural level of stress that is close to failure, and if pre-existing faults are within the influence of a stress-inducing agent (e.g., mining, reservoir-filling), then the features may be related to human-induced stress release.

As noted by Simpson (1986), a pre-existing state of stress that is both high and close to failure is a prerequisite for the occurrence of triggered seismicity (Gough and Gough, 1970; Simpson, 1976). This assumes that induced stress changes will be small in comparison to both the level of ambient stress at hypocentral depth and the failure strength of rock. More important than the absolute level of stress is the difference between the magnitude of the induced stress and the stress recovery required to initiate failure. Thus, in areas of relatively low seismicity, where the natural rate of strain accumulation is slow and the Earth's crust may remain at stress levels near failure for long periods of time, it is not unreasonable that induced stress changes on the order of a few bars or less can trigger seismicity on pre-existing faults. Simpson (1986) notes also that the failure of intact rock requires a much higher level of shear stress (to overcome the cohesive strength) than is needed to overcome frictional forces on pre-existing fault surfaces. Except for seismicity in deep mines, where induced stress changes on the order of kilobars may occur due to pre-existing high lithostatic stress conditions, it is unlikely that significant triggered earthquakes will occur in the absence of pre-existing faults.

Seismologic Associations

Criterion: If the pattern of microseismicity in an area coincides with the area, timing, and rate of human activities, then the seismicity may be related to these activities rather than tectonic processes.

Historical microseismicity also may provide evidence of a nontectonic origin for faults in areas of engineering activities. Simply, the pattern of (nontectonic) seismicity produced by these activities (e.g., mining, quarrying, reservoir-filling) typically reflects the influence of the local stress changes, whereas tectonic seismicity may correlate with tectonic structures (i.e., faults and fold belts). Also, the timing of microseismicity may also suggest a nontectonic origin for surface deformation features. If the onset of seismicity coincides with the onset of human engineering activities, a nontectonic origin for surface deformation features is suggested.

3.3.7 Subsidence and Collapse Structures

In general, subsidence refers to the gradual downward settling or sinking of the Earth's surface due to consolidation of underlying sediments or rock units, whereas collapse refers to the removal of an underlying support and consequent failure or collapse of an overlying body at rates that exceed subsidence (Section 2.6.6). As a result of a variety of natural and man-made processes (e.g., withdrawal of oil, gas, groundwater), permeable sandy alluvial deposits commonly undergo grain-packing rearrangements that result in ground subsidence (Section 2.6.6.1). Collapse may occur as a result of natural dissolution of a substratum or man-induced excavation of underground material (Section 2.6.6.2). The following section discusses criteria that permit differentiation between nonseismogenic subsidence and collapse phenomena from seismogenic-tectonic structures.

3.3.7.1 Criteria for Subsidence-Induced Faults

Criteria useful for differentiating subsidence-induced faults from tectonic faults include:

- Spatial and temporal associations
- Geologic structure
- Stratigraphic associations
- Seismologic associations

Spatial and Temporal Associations

Criterion: If the spatial pattern of subsidence faults coincides with the pattern of fluid extraction, the faults may be nontectonic.

Criterion: If the temporal pattern of subsidence faults coincides with the timing of fluid extraction, the faults may be nontectonic.

Criterion: If the rate of subsidence faults varies with the rate of fluid extraction, the faults may be nontectonic.

Surface fissuring, faulting, and low levels of regional seismicity may be associated spatially and temporally with natural gas/oil reserves and groundwater aquifers that have been depleted. If the spatial pattern of subsidence faults coincides with the pattern of fluid extraction, the faults may be nontectonic. However, a spatial mismatch may occur if there are lateral variabilities in the subsurface materials and/or fluid-flow regimes. In addition, subsidence-induced surface fissuring and faulting commonly coincide with the timing of natural or man-induced fluid-level changes, such that temporal coincidence between the fluid extraction and surface deformation is evidence of a nontectonic origin. The onset of surface deformation, evaluated perhaps via historical maps, records, or aerial photographs, may correlate with or closely follow the onset of fluid extraction. Similarly, if temporal changes in the rates of subsidence coincide with changes in the rates of fluid extraction, then a nontectonic origin for the faults is suggested.

Geologic Structure

Criterion: If a fault is within a series of concentric faults and/or borders a circular or semi-circular topographic depression, then it may be related to nonseismogenic-nontectonic subsidence.

Criterion: If a fault extends from the ground surface to depths of unconsolidated material in the shallow subsurface, but not beyond, then it may be related to nonseismogenic-nontectonic subsidence.

Criterion: If a fault extends from the ground surface to depths that are greater than that of unconsolidated material in the subsurface, then it may be related to seismogenic-tectonic processes.

Subsidence induced by excessive fluid withdrawal commonly produces oval or circular subsidence basins with reverse faults concentrated near the center of the depression, and arcuate fissures and high-angle normal faults around the depression. Map patterns of surface deformation features typically reflect subsurface geologic conditions and the dimensions of the subsidence basin. Furthermore, fissures and faults tend to originate near the zone of maximum curvature of basin margins, or above buried structural or stratigraphic discontinuities (faults and bedrock highs). Surface fissures typically do not extend to depths greater than several hundred meters, whereas tectonic faults extend at least several kilometers in depth.

Stratigraphic Associations

Criterion: If a fault is present in an area underlain by thick, unconsolidated granular materials, a subsidence origin should be considered.

Groundwater withdrawal and oil extraction-induced surface deformation features are restricted to sedimentary basins or fluid reservoirs, where thick, unconsolidated granular alluvium may be compacted at depth. Differential subsidence (and surface deformation) is favored where different thicknesses of compressible layers are juxtaposed, or where extensional strains are concentrated. These conditions may be associated with the margins of a sedimentary basin, buried faults, bedrock highs, buried stream channels, areas of differing groundwater conditions, or wherever alluvial deposition has created adjacent stratigraphic sections with markedly different compressibilities.

Seismologic Associations

Criterion: If the pattern of microseismicity in an area coincides with the area, timing, and rate of fluid withdrawal, then it likely is related to fluid extraction rather than tectonic processes.

Historical microseismicity may also provide evidence of a nontectonic origin for faults in areas of fluid withdrawal. Simply, the pattern of (nontectonic) seismicity produced by fluid withdrawal typically reflects the local influence of oil or groundwater extraction, whereas tectonic seismicity may correlate with tectonic structures (i.e., faults and fold belts). For example, Grasso and Wittlinger (1990) show that 95% of the epicenters of earthquakes attributable to development of a gas reservoir occurred within the boundaries of the production field. Concentric normal faults bordering the margins of areas of hydrocarbon exploration commonly are associated with microearthquakes having normal focal mechanisms, and microearthquakes near the areas of greatest extraction may show reverse focal mechanisms. In addition, the hypocenters produced by fluid withdrawal typically are shallow and located at the depths of fluid extraction. Lastly, the timing of microseismicity may also suggest a nontectonic origin for surface deformation features. If the onset of seismicity coincides with the onset of oil extraction and development, a nontectonic origin for surface deformation features is suggested.

3.3.7.2 Criteria for Collapse Structures

Collapse structures are caused by natural and human-induced processes, including subsurface dissolution and underground mining (Section

2.6.6.2). Moderately diagnostic criteria available for identifying features related to collapse deformation:

- presence of fractured soluble rocks and circulating groundwater
- karst-like topography
- Spatial and temporal associations with historical mining activities
- scale of karst-like features
- geometry of karst-like features

Presence of Fractured Soluble Rocks and Circulating Groundwater

Criterion: If karst-like geomorphic features are present in areas containing fractured, soluble rocks and circulatory groundwater movement in the shallow subsurface, then a dissolution collapse origin should be considered for observed faults.

Naturally induced collapse structures occur most commonly in carbonate rocks and evaporites (i.e., salt) where dissolution processes have removed underground support for the overlying rock. The formation of karst topography requires the presence of (1) soluble materials in the subsurface, (2) fractures or voids in the soluble rocks, and (3) circulatory movement of groundwater. Thus, the presence of these three characteristics in a given area raises the possibility that a fault is a result of (nonseismogenic-nontectonic) karst processes. Conversely, the absence of one of these characteristics suggests that karst processes probably are not responsible for the karst-like fault, and thus that the fault may be related to a seismogenic-tectonic process, or to a nonseismogenic-nontectonic process other than solution collapse.

Karst-like Topography

Criterion: If a region is characterized by sinkholes, closed depressions, and other evidence of karst collapse, then a fault may be related to nonseismogenic-nontectonic collapse.

Karst and karst-like geomorphic features include sink holes, closed depressions, and elongate subsidence basins, and are present in areas characterized by dissolution of soluble sedimentary rocks (Section 2.6.6.2). In addition, man-induced collapse of the ground surface over mined adits and rooms may create a karst-like topography. These areas are characterized by abundant small to moderate-sized closed depressions that may be connected by linear depressions above subsurface mining adits or large collapsed fractures. Where formerly active karst processes have ceased, and deposition within the depressions has subdued the topography, it may be more difficult to identify karst deformation features. Nevertheless, the presence of active or relict karst topography within a region or area raises the possibility that a given fault is a result of nonseismogenic-nontectonic collapse processes.

Spatial and Temporal Association with Historical Mining Activities

Criterion: If collapse-like features are present in areas of subsurface mining, then they may be related to nonseismogenic-nontectonic collapse.

Criterion: If collapse-like features developed during or after the time period of subsurface mining, then they may be related to nonseismogenic-nontectonic collapse.

Human-induced collapse structures involve subsurface mining of coal, ore-bearing rocks, and other resource materials. On a worldwide basis, underground coal mining has subjected more land to surface deformation than mining of any other mineral resource (Hasan, 1996). However, deformation is also widespread in many parts of the world where limestone, ore-bearing minerals, and other rock and mineral resources have been mined underground. The location and timing of underground mining activities, as determined by reviewing historical land use maps, photographs (aerial and land-based), and/or written records, may provide a basis for assessing the origin of collapse-like features and associated faults. If the collapse-like faults are present in areas of historical subsurface mining, then there is a strong possibility that the faults are related to mining activities. Given sufficient information on the timing and locations of historical mining activities, it may be shown conclusively that collapse faults are related to historical land use activities. Other subsurface information also may provide evidence of the origin of collapse-related faults, including geophysical and/or drilling data showing the presence of mining adits or rooms beneath or associated with collapse structures. Thus, review of historical records and subsurface conditions may provide highly diagnostic information for evaluating a fault's origin.

Scale of Karst-like Collapse Features

Criterion: If the ratio between the length (L) of topographic scarps or faults and the height (H) is small (L:H <10^3), then the feature may be related to nonseismogenic-nontectonic collapse.

Collapse-related faults commonly extend over a limited distance as a result of localized collapse as sinkholes and/or dolines. In addition, topographic scarps associated with collapse faults may have substantial heights over their local extent, also as a result of localized, but significant downward movement of the ground surface. Thus, scarps associated with collapse faults commonly are relatively short and high compared to topographic scarps produced through tectonic surface rupture. In general, a 1-m-high topographic scarp should be considered a possible tectonic scarp if it extends for more than 1 km. This suggests that a ratio between the length (L) and height (H) of a topographic scarp of 10^3 is a reasonable value for distinguishing between tectonic and collapse-related scarps, with collapse-related scarps having L:H ratios of less than 10^3.

Geometry of Karst-like Collapse Features

Criterion: If a fault is within a series of concentric faults and/or borders a circular or semi-circular topographic depression, then it may be related to nonseismogenic-nontectonic collapse.

Criterion: If a fault extends from the ground surface to depths of soluble rock in the shallow subsurface, but not beyond, then it may be related to nonseismogenic-nontectonic collapse

Criterion: If a fault extends from the ground surface to depths that are greater than that of soluble rock in the subsurface, then it may be related to seismogenic-tectonic processes.

The downdip extent and map pattern of faults provide information on the origin of collapse-related faults in karst topography. In karst or karst-like terrain, topographic scarps may be associated with normal faults that extend to the depth of soluble formation or mined rock. In areas containing circular or semi-circular karst sinkholes or dolines, normal faults associated with collapse may be concentric and coincident with the topographic depressions.

3.3.8 Volcanic-Related Structures

Most volcanic-related extensional faults have a clear geologic relationship to volcanic activity and other magmatic sources (Section 2.6.7). Volcanic-related faults that potentially could be mistaken for tectonic faults include collapsed calderas and dike-related surface deformation.

Several *moderately diagnostic* criteria are useful for differentiating volcanic-related faults from non-volcanic faults:

- geologic setting likely to contain volcanic features
- geophysical setting likely to contain volcanic features
- geomorphic characteristics

Geologic Setting Likely to Contain Volcanic Features

Criterion: If a fault is in volcanic terrain, then it may have a volcanic origin.

Criterion: If a fault is not in volcanic terrain, then it cannot be related to volcanic processes.

Determining the geologic setting of a fault is critical for evaluating whether the fault is related to seismogenic-tectonic processes or volcanic processes. If the geologic setting is, or is likely to be, characterized by volcanic processes, then a volcanic origin for a fault should be considered. The presence of volcanic deposits is the primary means for identifying the presence of volcanic processes. Caldera-collapse structures are typically associated with ignimbrites and volcanic ashes, and volcanic rift zones usually are associated with basalts and less viscous igneous rocks (Section 2.6.7.1).

Geophysical Setting Likely to Contain Volcanic Features

Criterion: If a fault is in a region of high heat flow, then it may be volcanic in origin.

Criterion: If a fault coincides with shallow magma intrusions, then it may be volcanic in origin.

Calderas and volcanic rift regions commonly are associated with high heat flow originating from an underlying magmatic body. The magmatic bodies may be identified using geophysical data based on differences in densities and magnetic susceptibilities between the magma and surrounding country rock. The density and magnetism of the magmatic intrusion typically is greater than the surrounding country rock. For dike intrusions, the geophysical signature typically is linear or elongate, whereas a caldera typically is identified by a roughly circular pattern of geophysical anomalies.

Geomorphic Characteristics

Criterion: If a fault is in an area of abundant volcanic fissuring, then it may be volcanic in origin.

Criterion: If a fault scarp is developed on volcanic deposits with little or no colluvium, then it may have a volcanic origin.

Criterion: If a fault is active for 10^6 years without substantial topographic relief, then it may be volcanic in origin.

Volcanic-rift zones are diffuse zones of extensional surface deformation that are dominated by (1) tensional fissures and normal faults with small vertical displacements, (2) shallow dike swarms, and (3) eruptive fissures. Scarps formed by dike-injection represent the cumulative effect of numerous decimeter-scale displacements. Fissures and normal faults formed by magmatic processes are similar in appearance to tectonic structures. However, in volcanic settings, fissures are more common than normal faults, indicating that most of the deformation is purely dilational (Hackett et al., 1996). Conversely, extensional tectonic faults are associated with fault scarps, small- and large-scale graben, topographically high mountain ranges separated by deep intermontane sedimentary basins, and considerable topographic relief. Furthermore, at the local scale, fault scarps associated with volcanic for rift zones produce little or no colluvium between successive surface faulting events (Smith et al., Appendix A).

Postseismic colluvial wedges usually are not observed in these extensional terrains, due to the small displacements along the normal faults and the resistance of the volcanic deposits to weathering and soil development processes. Lastly, normal faults within volcanic rift zones may be characterized by little or no significant topographic relief, even when they have been active for millions of years.

3.4 Criteria for Assessing Seismogenic Potential

Making a judgment regarding the seismogenic potential of a fault or tectonic feature is an important component in characterizing faults for seismic hazard assessments. In this section, we provide a discussion of the criteria and approaches that are commonly used to make this assessment. In seismic hazard assessments it is useful to define explicitly the criteria that are used to assess whether or not a tectonic feature or lineament is seismogenic, and to define the relative value that each criterion has in making the evaluation. "Relative value," in this sense, is an expression of the resolving power that a particular criterion--and its associated data--has in determining whether or not a tectonic feature is seismogenic. A criterion that provides a high resolving power (i.e., provides a strong indication that a feature is seismogenic) is therefore assigned a high relative weight. For example, if a tectonic feature is spatially associated with several M>5 earthquakes in the historical record, it would have a high potential for being seismogenic (that is, capable of generating M>5 earthquakes in the future). Some criteria may be less diagnostic and have a relatively low resolving power for distinguishing whether or not a tectonic feature is seismogenic. An example of such a criterion might be evidence that the tectonic feature has undergone multiple episodes of reactivation during its geologic history. Studies of earthquake occurrence within stable continental regions worldwide (Johnston and others, 1994) have suggested that observed moderage to large earthquakes commonly have been associated with multiply-reactivated structures. However, these studies also have concluded that there are many more such tectonic features that are not associated with observed earthquakes than have been associated--suggesting that this is not a very diagnostic criterion. Hence, this criterion would be assigned a relatively low weight.

Suggested criteria and their relative weights (on a scale of 1 to 10) are summarized in the following table, and discussed below.

Seismogenic criterion	Relative weight (W), 1-10
Spatially associated with large-magnitude (M>5) seismicity	7
Spatially associated with small-magnitude (M<5) seismicity	6
Extends through seismogenic crust (at least 10 km)	2
Displays evidence for brittle fault slip in present regional stress/tectonic regime	2
Displays geologic evidence for multiple episodes of reactivation	2
Sum of weights	19

The criteria summarized above are judged to be diagnostic types of information that would indicate whether or not a fault is capable of generating M>5 earthquakes in the future. These criteria are best suited for potential sources with unclear evidence of activity, but would not be used if

certain evidence is well-documented. For example, if a fault has been causally (not just spatially) associated with large-magnitude historical earthquakes or shows unequivocal evidence of repeated late-Quaternary displacements (as would an active fault in a plate-boundary tectonic setting), then that fault would be considered to be seismogenic with a probability of 1.0. If, however, these most diagnostic criteria are not present for a particular tectonic feature, then there exist uncertainties in the seismogenic potential of that feature. This uncertainty is expressed by the probability that the feature is seismogenic and will be less than 1.0.
The relative weight assigned to each criterion expresses the degree to which that particular criterion is diagnostic of a tectonic feature being seismogenic. Of course, it is recognized that any given feature will display all or some of these characteristics to varying degrees, depending on the available data. If data do not exist regarding a particular criterion, then that criterion is not used in the assessment, and the sum of the weights is adjusted accordingly. In practice, the complete absence of data—direct or indirect—about particular tectonic features is rare. In most cases, some type of judgment can be made (albeit highly uncertain) regarding virtually all of the criteria. The suggested criteria and the relative weights assigned to each are explained in detail below.

Spatial Association with M>5 Seismicity

The first criterion is the spatial association of a tectonic feature (e.g., fault or lineament) with observed moderage to large earthquakes, which have occurred historically or during the instrumental period. Note that this is not a causal association (which, clearly, would indicate that the feature is seismogenic), but is merely the spatial distribution of observed earthquakes relative to the feature of interest. We define spatial association as being an alignment of seismicity along the length of a substantial portion of a given feature (i.e., a fault or lineament). For general application, the spatial association is made in two dimensions only (map view), because reliable hypocenter data commonly are not available. Uncertainties associated with this criterion stem from the fact that, in low activity environments, the historical record is short relative to the recurrence intervals of large earthquakes. As a result, the occurrence of only a single historical event or the absence of large observed events is not uncommon. In addition, it is difficult to associate older historical events with a particular feature because of uncertainties in epicenter location, and sometimes even instrumentally-located events are not easily associated with known faults. This criterion is given a relatively high weight in assessing activity (7 out of 10), indicating that--assuming that a spatial association with M>5 seismicity can be made, the criterion is diagnostic of the potential to generate future M>5 earthquakes.

If available, other information including focal depth and focal mechanisms are considered in the final assessment of the seismogenic potential of a given feature. This is applied on a case by case basis, because these data commonly are not available. (See discussion below on using feature-specific data sets to "update" the assessment of seismogenic potential).

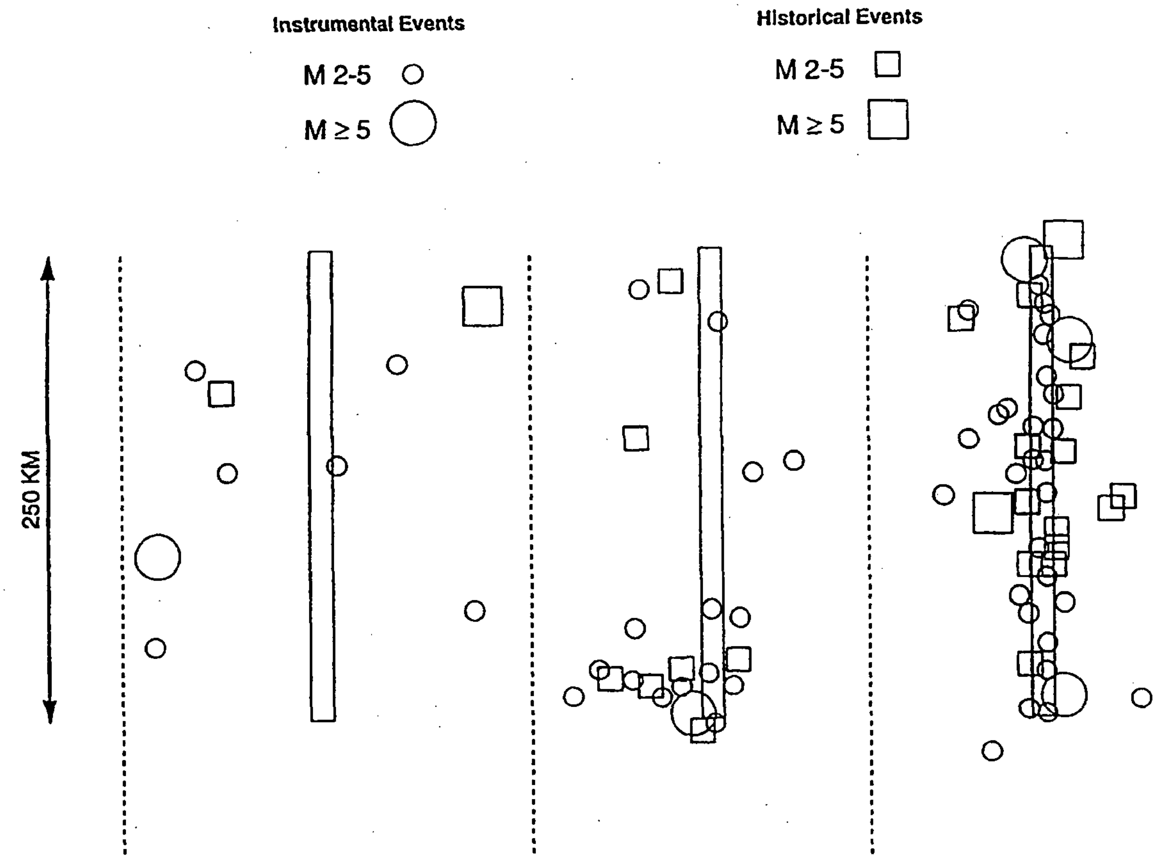
Spatial Association with M<5 Seismicity

The second criterion is the spatial association of a tectonic feature with small-magnitude (2<M<5) seismicity. Seismic networks capable of recording microseismicity (M<2) commonly are not present in many areas precluding the use of microseismic events in the definition. Again, we define spatial association as an alignment of seismicity along the length of a substantial portion of a given tectonic feature. It is common for seismogenic faults to be associated with small-magnitude seismicity, however, the association of a fault or lineament with such events does not necessarily indicate that the feature can also generate larger events. For this reason, this criterion is given a moderate-to-high rating, but is less diagnostic than evidence for spatial association with moderate to large magnitude earthquakes.

To illustrate the notion of a "spatial association" of seismicity with a tectonic feature, we present in Figure 3.1 a range of assessments for an idealized linear tectonic feature and idealized "observed" seismicity. The feature C displays a clear association with seismicity aligned along its entire length, and is assessed to be associated with seismicity with a probability of 1.0. In the example, both historical and instrumental events are present, suggesting a persistence through time and little chance that the spatial association with older historical events is merely a matter of the coincidence of poor locations. The feature A in Figure 3.1 has a single recorded event associated with it, and although other earthquakes occur in its vicinity, they show no tendency to align themselves along the feature. Thus, feature A in Figure 3.1 is assessed to be associated with seismicity with a probability of only 0.1.
The other example presented in Figure 3.1 further illustrates the definition of spatial association. Seismicity in the vicinity of feature B on Figure 3.1 occurs primarily near one end. This cluster, however, is part of a trend of seismicity that is perpendicular to the feature and extends away from the feature. Seismicity that occurs near the remaining portion of the feature is scattered, and cannot be distinguished from random background activity. This feature is assessed to have a probability of 0.2 of being spatially associated with seismicity. For two or more features that are proximal to each other, or intersect, we assume that the seismicity in their vicinity can be associated with any of the features. That is, a single seismicity cluster that is proximal to, say, three features, is considered in the individual assessments of each of the features.

Crustal Extent

Because of the rupture dimensions associated with moderate to large earthquakes, most seismogenic faults extend through the seismogenic crust (i.e., depths of 10-20 km). The third criterion expresses the degree to which a knowledge of the downdip crustal extent of a fault provides an indication of whether or not it is seismogenic. Evidence for expression in the deep crust might include large geophysical anomalies and gradients, the identification of a feature in deep seismic data, and geologic/tectonic interpretations of the tectonic role that a tectonic feature represents (e.g., continental rift faults). The relatively low weight assigned to this criterion (2 out of 10) reflects the judgment that crustal-scale extent is not an exclusively sufficient condition for assessing seismogenic potential. That is, seismogenic tectonic features would be expected to extend to seismogenic depths, but the mere fact that a feature extends to these depths does not provide much assurance that it is in fact seismogenic in the contemporary tectonic environment. This conclusion is supported in many geologic environments, such as much of the eastern United States, where abundant large, crustal-scale faults exist but few are believed to be presently seismogenic. These features are merely the vestiges of previous tectonic deformation episodes. The low weight assigned to this criterion also reflects the observation that some moderate size events can be generated at crustal depths shallower than 10 km. An example is the 1986 M 5 Leroy, Ohio, earthquake, which occurred at a depth of about 6-km in the Precambrian basement (Seeber and Armbruster, 1993, 1995). The 1989 Ms 6.3 Ungava earthquake in northern Canada also nucleated at a relatively shallow depth of 3 km (Bent, 1994).



	A Assessment	B Assessment	C Assessment
Spatial Association	0.10	0.20	1.00
No Spatial Association	0.90	0.80	0.00

Figure 3.1 Examples of spatial association with seismicity.

Brittle Slip in Present Stress Regime

The fourth criterion is geologic evidence for brittle slip that is kinematically consistent with the present tectonic stress regime. Clearly, faults and other features associated with ductile deformation reflect a deformation episode that occurred at a tectonic level that is below the seismogenic zone of the crust. We are, therefore, only considering tectonic features associated with brittle deformation. Seismogenic faults, by definition, are favorably oriented relative to tectonic stresses, because earthquakes are a manifestation of the release of tectonic stresses. However, in the absence of other evidence, determination of how favorably a fault is oriented involves a knowledge of the three-dimensional geometry of the fault, the orientations of the principle stress directions, and the rheological properties of the rocks (Zoback, 1992). This criterion appears to be a necessary component, but it is not sufficient alone for assessing seismogenic potential. For this reason, a relatively low weight (2 out of 10) is given to this criterion.

Multiple Episodes of Reactivation

The final criterion is geological evidence for multiple episodes of reactivation. A fault that exhibits evidence for brittle slip during distinctly different geologic time periods/tectonic phases might be an indication that that structure is a persistent zone of weakness in the crust. For example, in their global study of the association of M>4.5 earthquakes with various tectonic features, Johnston and others (1994) found that several of the moderage to large earthquakes that have occurred in stable continental regions have been associated with intra-cratonic rifts that have experienced multiple episodes of reactivation. Admittedly, they also found, however, that there are many rifts displaying such evidence of reactivation that have not been associated with seismicity. Uncertainties in applying this criterion arise from the lack of, or uncertainties in cross-cutting relationships of brittle structures, and recognizing different types of cogenetic displacement (e.g., normal faults in the hanging wall of a thrust fault). A structure reactivated in the present day stress field would also show evidence of activation during a previous phase of deformation. Evidence for multiple episodes of reactivation is given a relatively low weight (2 out of 10), because although it may suggest an enhanced potential for localizing future deformation, it is not judged to be a diagnostic indicator of that future potential.

Additional Criteria

Additional data of potential use to the evaluation of seismogenic potential that may be available locally, perhaps due to special studies. For example, local seismic networks may provide information on the spatial distribution of microseismicity (M<2), or geologic studies may identify evidence for geologically-recent faulting or evidence for paleoseismic shaking (e.g., paleoliquefaction evidence). These observations must be taken into account in arriving at the probability of the feature being seismogenic. The additional criteria are discussed here and the procedure for "updating" the assessment to include feature-specific additional data is discussed in the following section.

As defined above, the "spatial association with seismicity" criteria are based on a two-dimensional (map view) association between observed seismicity and the tectonic feature. In some cases, additional seismologic data may exist that could affect the assessment. These data include information on the hypocentral distribution (such that the three-dimensional geometry of the feature can be correlated with seismicity and/or the crustal extent of the feature assessed), focal mechanisms that suggest that the orientation of focal planes are consistent (or not) with the feature of interest, stress tensors that may or may not be consistent with the kinematic indicators on the feature of interest, the spatial pattern of microseismicity (M<2), and the spatial pattern of aftershock sequences that define the geometry of coseismic rupture planes.

An additional consideration is geologic evidence for Quaternary tectonic displacement. The Quaternary geologic record spans a much longer period than the historical record, and typically records evidence of large (M>6) earthquakes that rupture the surface. In plate-boundary tectonic environments, this criterion is the primary means of assessing whether or not particular faults are active. Uncertainties in using this criterion in tectonically less active regions come from uncertainties in dating the age of slip and discerning whether displacements are tectonic or other in origin. Given the possible occurrence of geologically young nontectonic deformation as described in Section 3.3, the identification of young tectonic deformation requires considerable care. In evaluating this criterion for a given fault or lineament, both the geologic evidence for recency of deformation associated with that feature, and the likelihood that the evidence represents seismotectonic or other deformation should be considered.

3.4.1 Procedure for Calculating Probability of Being Seismogenic

An example procedure for calculating the probablility that a fault or tectonic feature is seismogenic is discussed below. The feature-specific assessment evaluates the degree to which each criterion characterizes that particular feature (e.g., the probability that the feature is spatially associated with M>5 earthquakes might be assessed as 0.3, as shown below). To arrive at the probability that a given feature is seismogenic, the feature-specific assessments are multiplied by the relative weight or value (W) of each criterion, these products are then summed, and divided by the sum of the criteria weights (19). Assessments for a (hypothetical) feature might be the following:

In this example assessment, the probability that the feature is seismogenic is assessed to be about 0.5. By disaggregating the assessment this way into its component parts, the technical basis for the assessment is made clear. It should be noted, however, that although the assessment is made more explicit by dissecting it into its component parts, the assessment of the probability of the feature being seismogenic is still one that involves professional judgment.

As discussed above, in some cases additional data may exist for certain tectonic features (e.g., focal mechanisms, geologic evidence for Quaternary tectonic slip). In those cases, the probability of being seismogenic that is calculated for the feature—based on the five general criteria—is "updated" to account for the additional information. For example, in the illustration given above the tectonic feature is assigned a probability of 0.5 of being seismogenic based on the five criteria. Suppose that, in addition, a local seismic network has shown that microseismicity is aligned along a significant portion of the tectonic feature, and focal mechanisms for several earthquakes in proximity to the feature indicate that one nodal plane is parallel to the feature. This additional information would lend further support to the seismogenic potential of the feature. As a result, the assessment of 0.5 would be "updated" to a value of 0.6 to account for the additional data. Such an update is consistent with the concept of a Bayesian assessment of the probability, whereby the probability assessed from the five criteria is the "prior" assessment. The prior assessment is updated using new data to arrive at a "posterior" assessment of the probability.

Criterion	Probability	Weight	Product
Association M>5	(0.3)	7	2.1
Association M<5	(0.6)	6	3.6
Crustal Extent	(0.7)	2	1.4
Brittle Slip	(0.5)	2	1.0
Multiple Reactivation	(0.5)	2	1.0
		Sum of products	9.1
		Divided by sum of weights	19
		Probability (seismogenic)	0.48 or 0.5

4 CONCLUSIONS

4.1 Developing an Internally Consistent Interpretation: Does It All Make Sense?

Determining the tectonic or nontectonic origin of a fault is critical for properly identifying the fault as a potential seismic source. Unfortunately, as shown in this study, very few highly diagnostic criteria are available to define conclusively whether a fault is tectonic or nontectonic, and, if tectonic, whether the fault is seismogenic (i.e., capable of producing a moderate- to large-magnitude earthquake) or nonseismogenic. In addition, with rare exception, it is difficult to assess the origin of a fault based on information developed at the outcrop scale. In a few instances, individual fault parameters may be diagnostic of fault origin. For example, if a fault is associated with historical or instrumental seismicity, it must be considered tectonic and potentially seismogenic. If a fault dips gently down slope and daylights at or near the base of the slope, it likely is related to slope failure and is nontectonic (and thus nonseismogenic). If a fault extends to crustal depths of 5 to 7 km or more, it probably is tectonic, and its seismogenic potential must be evaluated. Other than these few diagnostic criteria, an integrated analysis of all fault parameters and the geologic and tectonic setting within which the fault occurs is needed to make a reliable assessment of fault origin.

Because of the paucity of highly diagnostic criteria, evaluating fault origin is subject to interpretation. Sound technical judgement based on all available data is required. In essence, evaluating a fault's origin requires not only the application of several criteria (as given in Sections 3.2 and 3.3 above), but also the development of an interpretive model that "makes sense" considering all geologic, structural, stratigraphic, kinematic, and other relevant data. Such a model should be internally consistent and should take into account all known characteristics of the feature and its surroundings.

Part of developing an internally consistent model of a fault's origin is understanding its geologic context. As noted in Section 3.1, the geologic context of a feature encompasses many data sets at local and regional scales (Table 3.1), and is important for determining whether a particular fault is susceptible to various tectonic or nontectonic processes. For example, solution collapse faults are not likely to be present in areas without soluble substrata, and volcanic-related faults are not likely to be present in areas without a geologic history of volcanism. Thus, understanding the geologic context of a feature provides the basis for developing a reasonable and defensible interpretation of fault origin and an internally consistent model.

In addition, the interpretation of fault origin can support judgment on whether the fault (if tectonic) is seismogenic or nonseismogenic. This interpretation is heavily dependent on the scale, geometry, and kinematics of the fault (*See* Section 3.1). For example, if a tectonic fault is several kilometers or more in length and penetrates 5 km or more in depth, it should be considered potentially seismogenic. If a tectonic fault shows repeated brittle displacements of several tens of centimeters or more, it should be considered potentially seismogenic. But if a tectonic fault extends in length a few hundreds of meters or less and displays only a few centimeters of displacement, then it may be reasonably considered a nonseismogenic feature. In other words, if the scale of the fault is large enough such that it could produce a moderate to large earthquake, based on the occurrences of historical earthquakes throughout the world, then it should be considered a potential seismogenic source. Likewise, if the geometry and kinematics of the fault are consistent with other potential seismic sources, then it should be considered a potential source as well. In short, the development of an internally consistent model, which takes into account the geologic context, scale, geometry, and kinematics of the fault, is critical to assessing the fault's origin and seismic potential.

If a fault is interpreted to be nontectonic, then an internally consistent model must be developed to explain its origin by a nontectonic process. For regulatory purposes, it generally is inadequate merely to ascribe a fault to a nontectonic process without providing an interpretation of the fault's origin. The specific nontectonic process (or set of alternative processes) should be identified with supporting rationale and documentation. Competing alternatives should be identified, where appropriate, and the relative degree of confidence that each alternative correctly explains the fault origin should be provided. As with tectonic faults, developing an internally consistent model that "makes sense" with all the available data is necessary to support an interpretation that a fault is a nontectonic (and therefore nonseismogenic) feature.

4.2 Application to Siting of Nuclear Power Plants

Satisfying the definitions and criteria used by the Nuclear Regulatory Commission to assess whether a fault is a "capable tectonic source" requires differentiating tectonic faults that are capable of generating earthquakes from nontectonic faults and tectonically induced secondary ground deformation features that are not capable of generating significant (i.e., > M 5.0) earthquakes (Figure 4.1). As stated in the regulatory guidelines (Regulatory Guide 1.165, Appendix D), "if questionable features cannot be demonstrated to be of nontectonic origin, they should be treated as tectonic deformation." It is therefore incumbent on the applicant to conduct sufficient investigations to adequately characterize the nature and extent of surface deformation such that the mechanism of formation can be determined.

It is recognized that in many cases faults and associated surface deformation may be ambiguous with regard to the exact cause of their formation, which means that when faced with questionable structures, the geologist must assess the cumulative weight of several imperfect criteria. Applying this approach in a regulatory environment requires that both the relative merits of the varying diagnostic criteria for varying types of faults are well understood, and that the uncertainties or lack of knowledge of the characteristics of a specific feature with respect to those criteria are explicitly acknowledged.

It is important to clearly examine alternative hypotheses and to provide that the relative credibility assigned to any specific criterion reflects the full range of uncertainty and diversity of views. The technical interpretations, their logic and basis in the available data, and all uncertainties should be documented in sufficient detail to support the final assessment of fault origin. By explicitly identifying the key parameters or diagnostic criteria and the extent of knowledge or available data for each, it is possible to better focus further investigations and help evaluate their potential value for reducing uncertainty in the assessment of fault origin.

The approaches used to mitigate potential hazard(s) posed by a fault depend on whether the fault is both tectonic and seismogenic. In all cases, faults that may experience movement in their present geologic or tectonic setting need to be evaluated with regard to their potential to produce surface rupture. Seismogenic faults (i.e., those capable of generating >M 5.0 earthquakes) require additional study to evaluate their potential to generate significant vibratory ground motion.

In conclusion, the evaluation of the potential hazard of an identified or postulated fault is an iterative process that requires collecting and

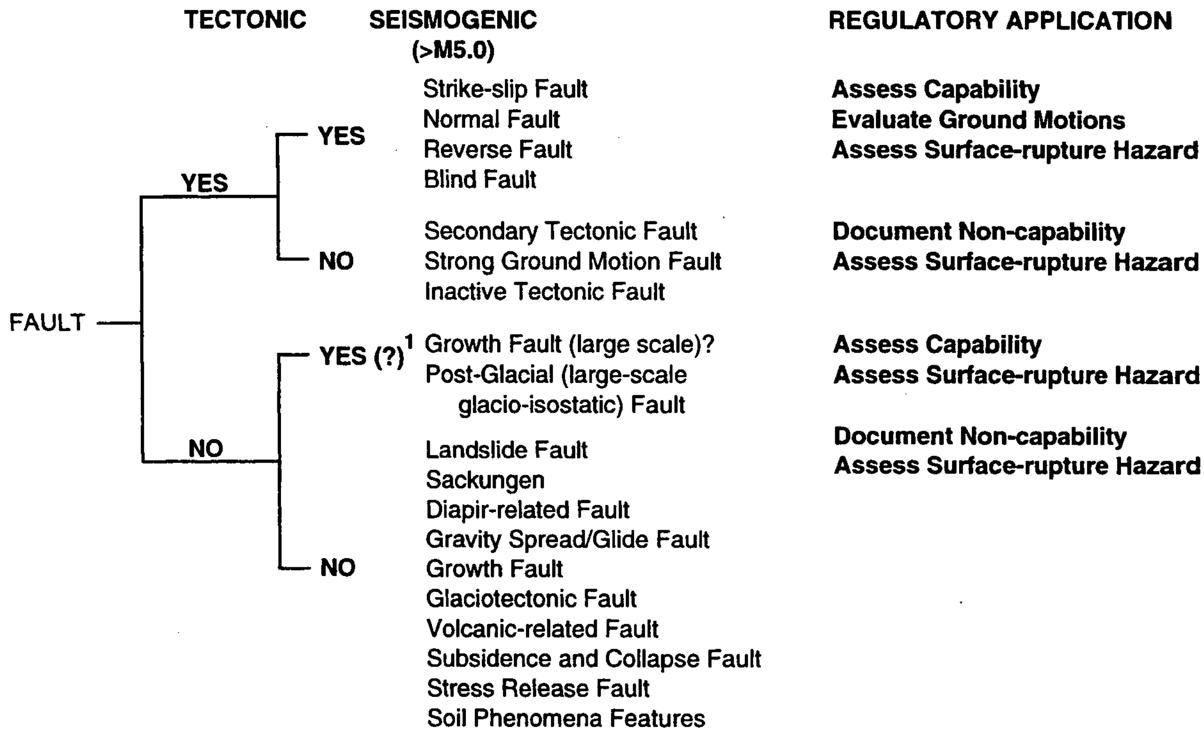


Figure 4.1 Diagram showing fault classifications and appropriate regulatory requirements.[1]

evaluating a variety of data sets. For nuclear power plant sites, a multidisciplinary approach involving the assessment of geologic, geophysical, seismologic, and geodetic data generally is needed to adequately identify and characterize tectonic faults. Differentiating tectonic from nontectonic faults, and tectonic-seismogenic faults from tectonic-nonseismogenic faults requires: (1) an understanding of the types of tectonic and nontectonic processes that can produce faults; (2) an understanding of the geologic and tectonic setting (or context) within which a fault occurs; and (3) detailed characterization of fault-specific attributes, in particular a fault's scale, geometry, and kinematic behavior. Typical steps in the identification and characterization of a potentially significant fault are described below.

Step 1: Surface fault and/or ground deformation of uncertain origin is identified.

Step 2: Compare characteristics of fault or surface deformation to characteristics of tectonic and nontectonic faults discussed in Sections 2 and 3.

Step 3: Make a preliminary assessment of the fault's origin and document uncertainties in data or interpretations of the available data. This requires

[1] Examples of seismogenic-nontectonic faults are rare to nonexistent. Moderate- to large-magnitude earthquakes may have occurred due to postglacial faulting (probably reactivation of pre-existing tectonic structures). Some large-scale growth faults may have sufficient dimensions to be associated with moderate-magnitude earthquakes.

testing alternative hypotheses to develop an integrated model that best fits the available data and is kinematically, mechanically, and structurally viable. The assessment of fault origin very seldom will be made with 100-percent certainty. Sources of uncertainty may include (1) having inadequate data to properly assess geologic context or fault parameters; (2) having two or more alternatives that may equally (or unequally) explain the fault origin; and (3) scientific uncertainty in understanding the tectonic or nontectonic process itself. An explicit treatment of uncertainty in the assessment of fault origin should be provided. Uncertainties in characterization of specific fault attributes (e.g., scale, geometry, and displacement history) should be noted.

Step 4: Conduct additional characterization investigations focusing on those activities that can be used to test alternative hypotheses and that will significantly reduce uncertainties in the assessment of diagnostic criteria.

Step 5: Based on the results of Steps 2, 3, and 4, assess whether fault or feature is tectonic or nontectonic.

Step 6: Based on the results of Steps 2, 3, and 4, assess whether fault is seismogenic (i.e., may generate M 5 or larger earthquakes) or is nonseismogenic.

Step 7: If the fault or surface deformation is judged nontectonic, the potential for surface deformation hazard (e.g., site surface collapse, subsidence, uplift, or differential movement) should be evaluated. Engineering solutions should be provided, or the site should be deemed unsuitable. If engineering solutions appear to be practical, a detailed description of the scale and geometry of the feature, and subsurface conditions obtained through reliable methods of investigation, should be developed to derive a design basis.

Step 8: If the fault or surface deformation is judged to be tectonic, the capability of the fault or feature should be determined. The assessment of fault capability requires an understanding of the timing of recent movement and the scale or seismogenic potential of a feature. A complete description of the data and judgments required to assess fault capability is not included in this study.

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APPENDIX A

PANEL MEMBER SUMMARY REPORTS

APPENDIX A

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INVESTIGATION OF SUSPECT LIQUEFACTION FEATURES AT THE SAVANNAH RIVER SITE, SOUTH CAROLINA

John C. Abbott[1]
Robert L. Gelinas[1]
David C. Amick[1]

1 INTRODUCTION

Seismically-induced liquefaction features can be used to constrain the paleoseismic history of a region. Within the past several years, several organizations have conducted extensive investigations of liquefaction features along the South Carolina coast. Whereas many of these studies focused on seismically-induced liquefaction associated with a large earthquake in 1886 near Charleston, South Carolina, others documented older liquefaction features interpreted to have been caused by large prehistoric earthquakes of magnitudes similar to the 1886 event. Within the past two millennia, large events may have occurred in coastal South Carolina about every 500 to 600 years (Amick and Gelinas, 1991; Amick et al., 1990a; Obermeier et al., 1986, 1990).

The Savannah River Site, located approximately 145 km (90 mi) west-northwest of Charleston, is a nuclear facility managed by the Department of Energy (Figure JA.1). From the 1950s to the 1980s, the primary mission of the Savannah River Site was to produce nuclear materials for the United States' nuclear weapons arsenal. More recently, the Savannah River Site has also provided materials for medical, industrial, and research purposes and the space program. The production and storage of nuclear materials at the Savannah River Site has prompted detailed investigations into possible geologic hazards and the paleoseismic history of the region.

Although no clear evidence of Quaternary seismicity has been observed in the area, several features in outcrop on the Savannah River Site have been interpreted by various geologists as "clastic dikes" or other features suggestive of liquefaction. These suspect features were investigated to assess if they are evidence of neotectonic activity. Hundreds of these features were described in the field and evaluated to assess if they have the diagnostic characteristics of seismically-induced liquefaction. The results of this field investigation are summarized herein.

This report was adapted from a larger investigation performed by Science Applications International Corporation (SAIC) for Westinghouse Savannah River Company under Subcontract C001015P Task 30. The full investigation is provided in *Paleoliquefaction Assessment of SRS, Fiscal Year 1996 Status Report* (WSRC, 1996a).

2 REVIEW OF LIQUEFACTION

Liquefaction is the transformation of a granular material (usually sand) from a solid state to a fluid state due to an increase in pore-water pressure (Youd, 1973). Liquefaction may be induced by either seismic or aseismic mechanisms. Much of our understanding of eastern United States

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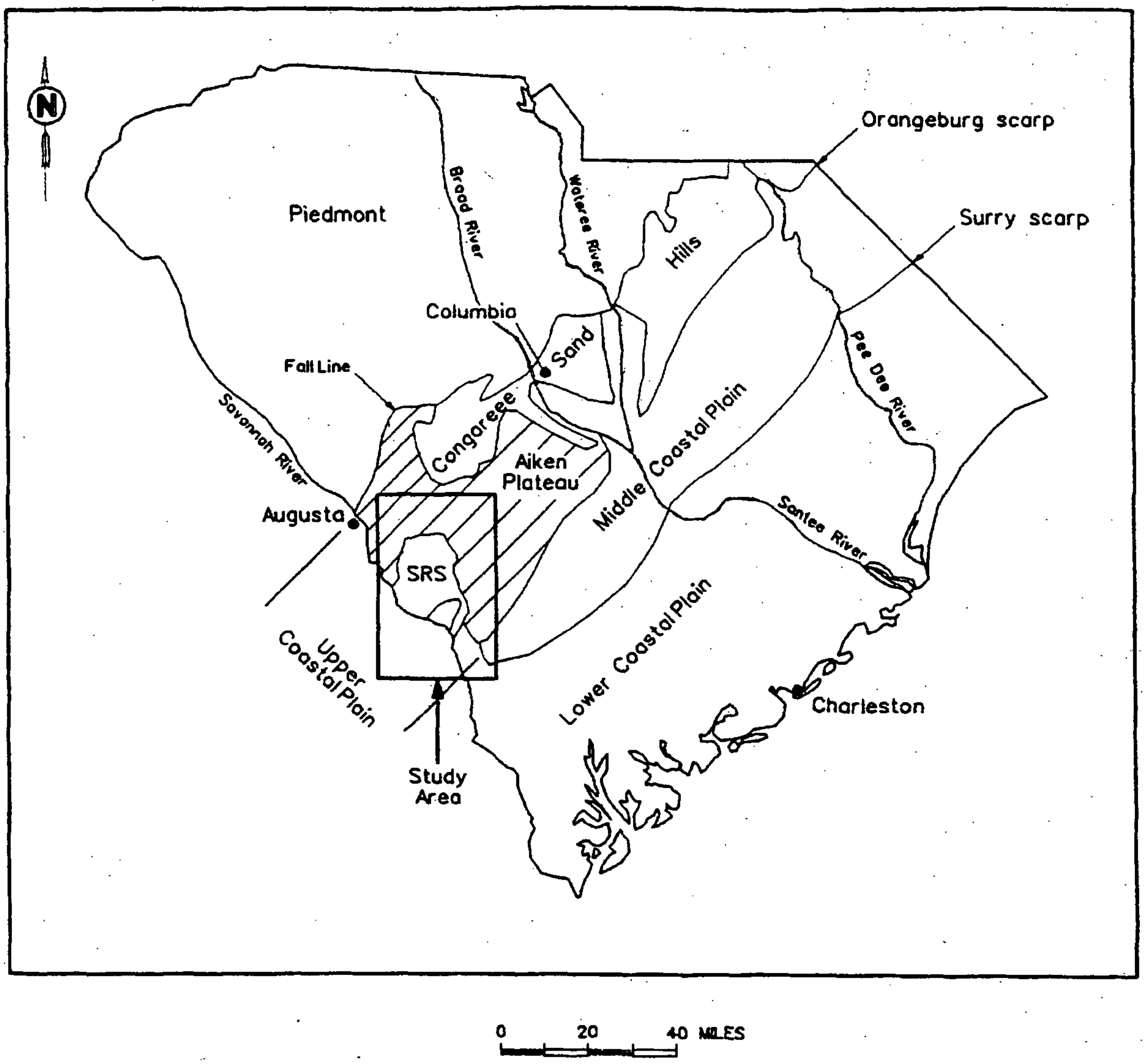

Figure JA.1 Location of the Savannah River Site (SRS) and map of the major physiographic regions of the South Carolina Coastal Plain. Used with permission of the South Carolina Department of Natural Resources: Aadland, R.K., U.A. Gellici, and P.A. Thayer, *Hydrogeologic Framework of West-Central South Carolina*, Water Resources Division Report 5, 200 p., 1995.

liquefaction phenomena stems from eyewitness accounts of liquefaction during the 1886 Charleston event and recent studies along the South Carolina coast, within the New Madrid seismic zone, and in the Wabash Valley of Indiana and Illinois. Key references about paleoliquefaction include Amick and Gelinas (1991), Amick et al. (1990a, b, c), Amick et al. (1992), Munson et al. (1992), and Obermeier et al. (1986 and 1992). The following discussion draws heavily from those documents.

2.1 Seismically-Induced Liquefaction

Liquefaction that results from cyclic shear strain of earthquake ground motions is referred to as seismically-induced liquefaction. As reported by 19th Century investigators, the 1886 Charleston event produced numerous seismically-induced liquefaction features over a 1500 km^2 (577 mi^2) area. The most spectacular seismically-induced liquefaction features associated with the 1886 Charleston earthquake were sand-blow explosion craters. Craters within the meizoseismal area were about 0.3 to 1.2 m (1 to 4 ft) deep and up to 5.5 m (18 ft) across. The largest craters measured approximately 7.3 m (24 ft) in diameter. An extensive ejection blanket of sand up to 0.6 m (2 ft) thick extended for tens of feet outward from many features. The formation process of sand blow explosion craters is known from eyewitness accounts during the 1886 earthquake and from studies of the internal morphology of exhumed, prehistoric craters. Following the onset of seismic loading and the development of a water interlayer, four sequential phases have been identified in the development of sand-blow explosion craters: an explosive phase, a flowage phase, a collapse phase, and a filling phase. This progressive development of a sand-blow explosion crater is illustrated in Figure JA.2. In the Charleston area, sand-blow explosion craters are observed primarily in beach deposits, and are generally absent in fluvial settings.

In addition to crater-like seismically-induced liquefaction features, eye-witnesses also reported numerous fissures and cracks in the meizoseismal area. Whereas some of these features were described as "dry", many emitted large volumes of water laden with sediment. This type of liquefaction feature is referred to as a sand-vent/fissure. Based on a review of historical accounts, as well as recent field investigations, sand-vents/fissures primarily occurred adjacent to rivers and streams in the meizoseismal area. Based on laboratory studies and internal morphology, four sequential phases have been postulated in the development of sand vent/fissure features: development of a water interlayer, lateral flowage, confining cap rupture, and sand extrusion. The progressive development of a sand vent/fissure is illustrated in Figure JA.3.

2.2 Aseismic Liquefaction and Similar Features

Liquefaction is not related exclusively to seismicity. For example, liquefaction may be induced by a rapid increase in groundwater level associated with a flood or storm surge. Liquefaction may also result from the natural settling and compaction of loose, saturated sands that are isolated within less permeable stratigraphic units.

A wide variety of soft sediment and soil structures resemble seismically-induced liquefaction features, but are unrelated to past earthquakes. Paleoliquefaction studies in the southeastern and northeastern United States encountered a variety of these features (Obermeier et al., 1986, 1990; Amick et al., 1990a). They fall into two general categories: (1) penecontemporaneous soft sediment deformational features, and (2) post-depositional features.

Penecontemporaneous soft-sediment deformation features are primary features generated in sediment during or shortly after deposition. Typical primary deformation features include spontaneous liquefaction unrelated to seismicity, dish structures, slumps and slides, and convolute bedding. Penecontemporaneous soft-sediment deformation features are common in unconsolidated, Quaternary deltaic deposits of the southeastern United States.

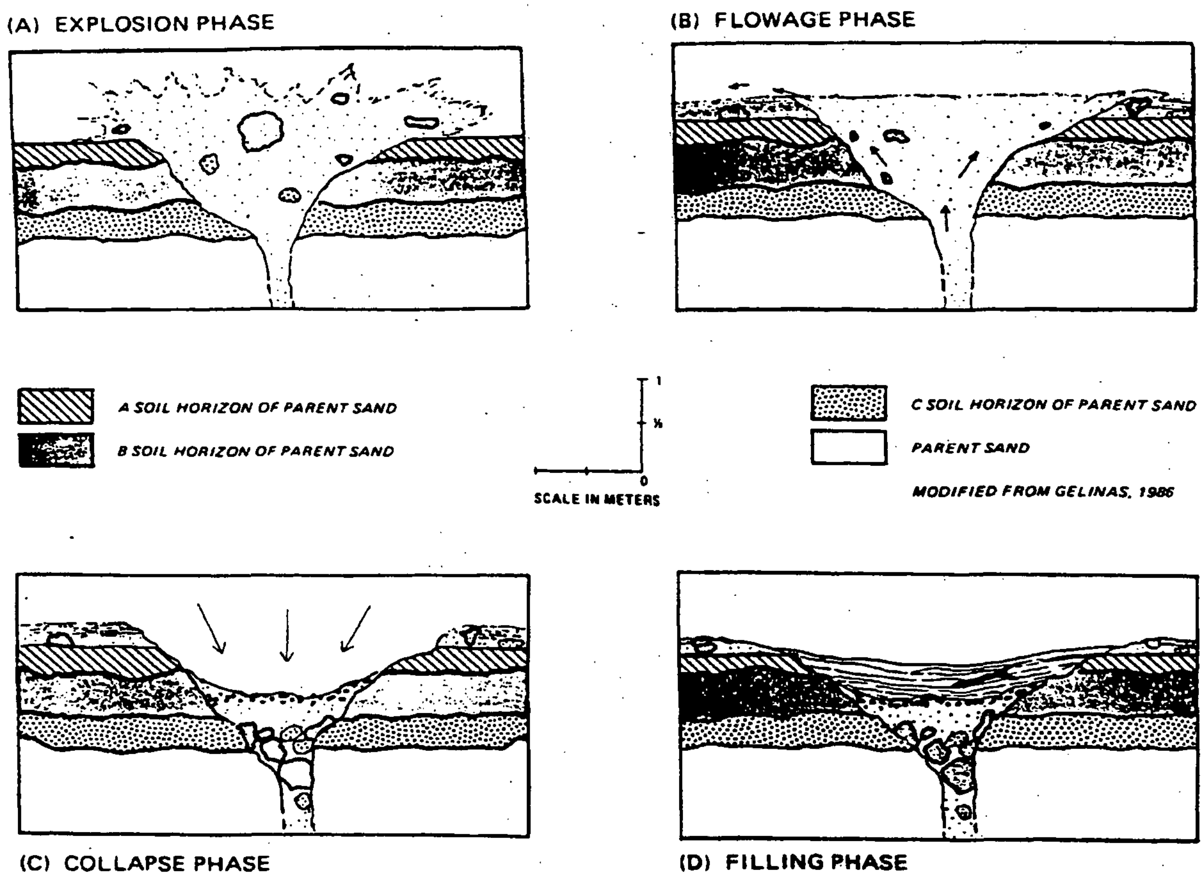

Figure JA.2 Schematic representation of the phases in the development of a sand-blow explosion crater proposed by Gohn et al. (1984). (A) Explosive Phase: cyclic seismic loading results in the reduction of void spaces and an associated increase in pore pressures. A water interlayer forms in parent sand with pore pressures great enough to explode and excavate a crater. (B) Flowage Phase: flow of sand-laden waters continues after ground motion has ceased, and stops only when the pore pressure of the source sands equals the confining pressure. (C) Collapse Phase: collapse begins when pore-water pressures decrease to nearly the confining pressure of the source sands. During this phase, clasts settle according to size and density, resulting in clast segregation into two zones (the large clast zone near the bottom of the crater and within the central vent and the small clasts zone near the top of the crater). As pore-pressures continue to decrease, upward transport of fine grained material stops and the crater begins to collapse. At this time, small-scale dewatering structures may develop as well as local gravitational faulting along the sides of the crater. (D) Filling Phase: filling of the crater probably takes place in the days, weeks and months following the earthquake, as materials from the crater rim eventually fill the crater by sedimentary and eolian processes (from Gelinas, 1986).

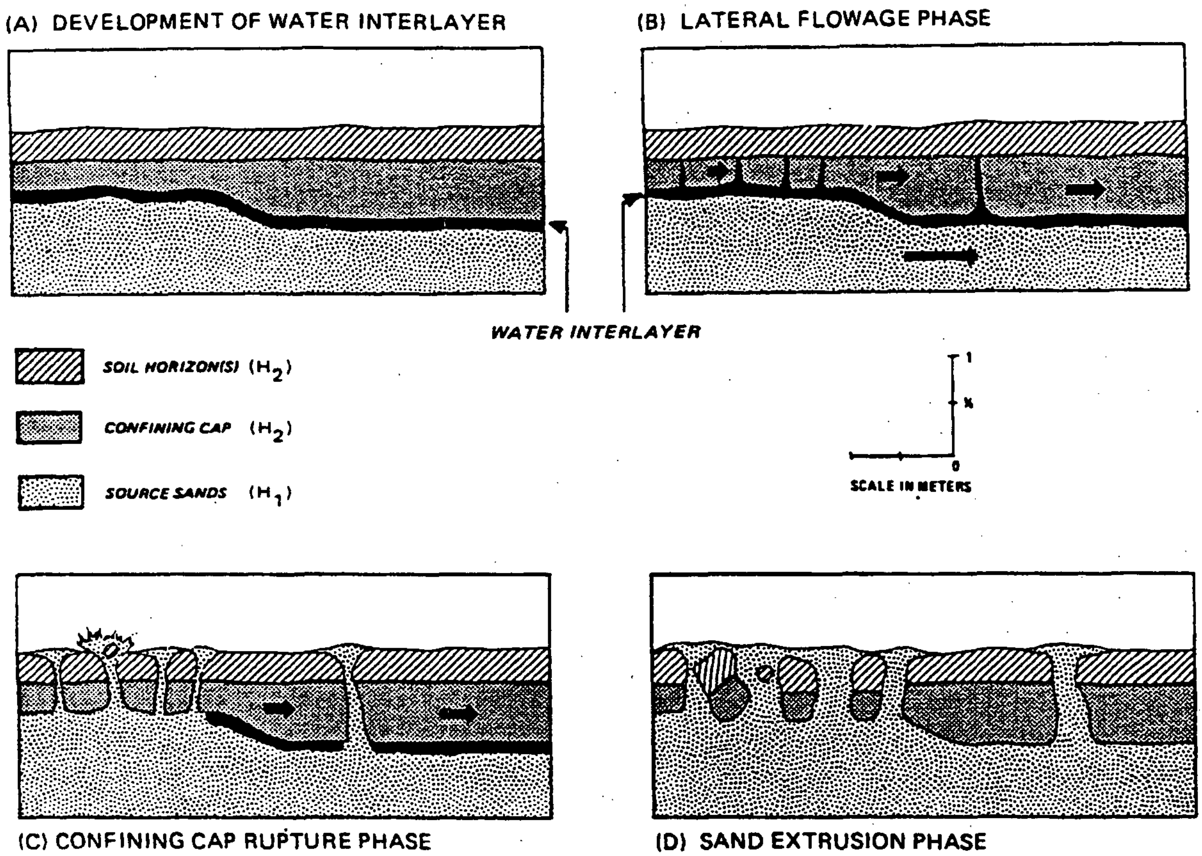

Figure JA.3 Schematic representation of the phases in the development of a sand vent/fissure. (A) Development of water interlayer due to cyclic shear strain and dewatering under a more impervious confining cap. (B) With the liquefaction of underlying sands and the development of a water interlayer, the friction at the contact between H_1 and H_2 is reduced to the point where the non-liquefied cap begins to move laterally in response to local gravitational forces. (C) As the cap (H_2) is transported, it begins to break apart, resulting in the formation of tension fractures that are filled by the underlying liquefied sands. Sands may vent to the ground surface. In general, vents/fissures are more closely spaced where the cap material is thinner. (D) Sands vent to ground surface in greater quantities. With the venting of the sands, pore-water pressures decrease, the coefficient of friction at the sand/cap boundary increases and lateral spreading ceases. Although downslope mass transport has stopped, flow of sand-laden waters continues until the pore pressure of source sands equals the confining pressure (from Gelinas, 1986).

Post-depositional features are secondary features generated in sediment after deposition has taken place. Typical post-depositional features include geochemical alteration features, tree throws, root casts, animal burrows, and ice wedge features. Geochemical alteration features include pedogenic features such as cutans. Cutans are a pedogenic modification of the texture, structure, or fabric of the host material, either by concentration of particular soil constituents or in-situ modification of the matrix.

2.3 Previous "Clastic Dike" Assessments in the Coastal Plain of South Carolina

Features commonly referred to as "clastic dikes" are widespread in near-surface Coastal Plain sediments of South Carolina (Siple, 1967; Colquhoun, 1969; D'Appolonia, 1982; McDowell and Houser, 1983). The term "clastic dikes" has been applied to features interpreted to have formed from various origins, both tectonic and nontectonic. Some of these features exhibit the diagnostic characteristics of liquefaction, and have been interpreted to be the result of seismically induced liquefaction (Amick et al., 1990a, b, c). Others were interpreted as weathering, shrinkage, or mass wasting features (Johnson and Heron 1965; Siple, 1967; McDowell and Houser, 1983). Some "clastic dikes" were initially interpreted to have resulted from upward injection of water-saturated sand (Siple, 1967; Zupan and Abbott, 1975; McDowell and Houser, 1983; McCarten et al., 1990), but were later re-interpreted as weathering features that resemble deformation structures (Geomatrix, 1993).

3 SAVANNAH RIVER SITE GEOLOGY

The Savannah River Site is underlain by a seaward-thickening wedge of unconsolidated and semi-consolidated sediments of the Atlantic Coastal Plain. The Coastal Plain sequence is an essentially undeformed package of marine and fluvial sediments consisting of clay, limestone, sand and gravel. In the vicinity of the Savannah River Site, these sediments are approximately 320 m (1,050 ft) thick and range in age from Late Cretaceous to Holocene (Figure JA.4) (Aadland et al., 1995).

3.1 Stratigraphy

The following discussion of stratigraphy is limited to the two units in which the suspect liquefaction features were investigated, the Tobacco Road Sand and the overlying Altamaha Formation. These two units are the youngest Tertiary deposits of the Coastal Plain sequence at the Savannah River Site (Figure JA.4), and account for the vast majority of surface exposures.

The Tobacco Road Sand and Altamaha Formation are interpreted to be fluvial and transitional marine in origin. The lithologies of both units are suggestive of lower delta plain environments. Recent studies suggest that the Altamaha Formation and Tobacco Road Sand are similar in texture and composition, indicating that they might be similar genetically (i.e., that they are part of the same transgressive/regressive depositional cycle, with the Altamaha Formation being the most continental end member lithofacies) (Colquhoun et al., 1994). Fossil evidence is scarce, so it has not been established whether or not there was a significant hiatus between the time of deposition of the two units (Siple, 1967; Logan and Euler, 1989; Nystrom et al., 1991; Fallaw and Price, 1992).

3.1.1 Tobacco Road Sand

The late Jacksonian (late Eocene) Tobacco Road Sand is the upper unit of the Barnwell Group (Nystrom and Willoughby, 1982; Nystrom et al., 1986). It consists of moderately to poorly sorted, red, brown, tan, purple, and orange, fine to coarse, clayey quartz sand.

Pebble layers are common, as are clay laminae and beds. Ophiomorpha burrows are abundant in parts of the formation. The base of the formation

LITHOSTRATIGRAPHIC UNITS
(Fallaw and Price, 1995)

Age	Group	Formation	Member / Unit
Miocene	Barnwell	Altamaha Formation	
Late Eocene	Barnwell	Tobacco Road Sand	
Late Eocene	Barnwell	Dry Branch Formation	Irwinton Sand Member
?	Barnwell	Dry Branch Formation	Griffins Landing Member
Middle Eocene	Barnwell	Clinchfield Formation	Albion Member
Middle Eocene	Barnwell	Clinchfield Formation	"Orangeburg District Bed"
Middle Eocene	Barnwell		Riggins Mill Member
Middle Eocene	Barnwell		Utley Limestone Member
Middle Eocene	Orangeburg	Tinker Formation	Santee Limestone
Middle Eocene	Orangeburg		"Blue Bluff Unit"
?	Orangeburg	Warley Hill Formation	
Early Eocene	Orangeburg	Congaree Formation	
Early Eocene	Black Mingo	Fourmile Branch Formation	
Late Paleocene	Black Mingo	Snapp Formation	
Late Paleocene	Black Mingo	Lang Syne Formation	
Early Paleocene	Black Mingo	Sawdust Landing Formation	
Late Cretaceous		Steel Creek Formation	
Late Cretaceous	Black Creek		
Late Cretaceous		Middendorf Formation	
Late Cretaceous		Cape Fear Formation	
Late Triassic	Newark Supergroup	Sedimentary Rock (Dunbarton Basin)	
Paleozoic-PreCambrian (?)		Crystalline Basement Rock	

Figure JA.4 Generalized lithostratigraphy for the Savannah River Site. Permission to use this copyrighted material is granted by Southeastern Geology.

is marked in places by a coarse layer that contains flat quartz pebbles. Sediments have the characteristics of a shallow marine deposit. The top of the Tobacco Road Sand is chosen where comparatively well-sorted sand is overlain by more poorly sorted sand, pebbly sand, and clay of the Altamaha Formation. Contact between the units is difficult to pick on geophysical logs because the upper surface of the unit is very irregular due to fluvial incision that accompanied deposition of the overlying Altamaha Formation (Aadland et al., 1995; Fallaw and Price, 1995; WSRC, 1996b). The thickness varies considerably because of the eroded upper surface, but is at least 18 m (60 ft) in places (Fallaw and Price, 1995).

The unit has been traced in outcrop from its type locality in Richmond County, Georgia, to the Savannah River Site (Huddlestun and Hetrick, 1978, 1986; Nystrom and Willoughby, 1982; Fallaw and Price, 1995). The "Barnwell Formation" of Siple (1967) seems to correlate roughly to the Tobacco Road Sand (Fallaw and Price, 1995). The Tobacco Road Sand is unit O1 and perhaps part of unit E8 of Prowell et al. (1985) (Fallaw and Price, 1995).

3.1.2 Altamaha Formation

Deposits of poorly sorted, silty, clayey sand, pebbly sand, and conglomerate of the Miocene Altamaha Formation cap many of the hills at higher elevations over much of the Savannah River Site. Clay clasts, weathered feldspar, and muscovite are abundant in places. Cross-bedding is prominent locally. The color is variable, and facies changes are abrupt. In general, the Altamaha Formation has poorer sorting, larger and more common feldspar grains, more abundant and thicker clay beds, more argillaceous and indurated sands, larger pebbles, and in places, more muscovite than the underlying Tobacco Road Sand (Fallaw and Price, 1995). Thicknesses up to 18.3 m (60 ft) have been documented, although the thickness changes abruptly because of channeling into the underlying Tobacco Road Sand during Altamaha deposition and subsequent erosion of the Altamaha Formation itself. According to Siple (1967), a significant characteristic of the Altamaha Formation which distinguishes it from the underlying Tobacco Road Sand is the presence of numerous "clastic dikes". These features are substantially more common in the Altamaha than in the underlying deposits.

The type locality of the Altamaha Formation is in southeastern Georgia (Huddlestun, 1988; Nystrom and Willoughby 1992; Fallaw and Price, 1995). Previously, the strata of the Altamaha Formation at the Savannah River Site were assigned to the Hawthorn Formation (Siple, 1967). The informal term "Upland Unit" has also been applied to these strata (Nystrom and Willoughby, 1982; Nystrom et al., 1986; Nystrom et al., 1991; Colquhoun et al., 1983; Steele, 1985; McClelland, 1987; Logan and Euler, 1989). Other terms applied to these strata are "Lafayette" (Sloan, 1908) and "Citronelle" (Doering, 1960, 1976; Smith and White, 1979). The Altamaha Formation is unit M1 of Prowell et al. (1985) (Fallaw and Price, 1995).

3.2 Structural Geology

Several different kinds of deformational features have been described from outcrops of the Altamaha Formation and Tobacco Road Sand, including small-scale faults, closed circular depressions, joints, and "clastic dikes". D'Appolonia (1982) determined that the small-scale faults and joints within the Savannah River Site area were not capable (i.e., they do not represent a potential seismic hazard).

No map-scale faults are present in surface exposures at the Savannah River Site, although the presence of subsurface faults have been documented by numerous studies. The most comprehensive study to date is that of Stieve and Stephenson (1995), who interpreted seismic reflection data, potential field data (aeromagnetic, ground magnetic, time domain electromagnetic, and gravity surveys), and drilling results. Their study evaluated the style and timing of seven significant fault structures that underlie the Savannah River Site (Figure JA.5). The timing of the most recent deformation on each of the major faults is poorly constrained (Stieve and Stephenson, 1995; Geomatrix 1993). None of the faults studied by Stieve and Stevenson (1995) could be traced upwards through the Coastal Plain section and shown to offset late Tertiary materials. No surface expression of these faults has been identified. Apparently unrelated small-scale faults are present in a few areas. These faults generally have displacements of less than a few feet, and are widely accepted to have formed by aseismic processes such as dissolution-related subsidence and/or mass wasting.

Joints are common on the Savannah River Site and regionally. Bartholomew et al. (1995) described joint fillings and orientations in the area and constrained the relative sequence of fracture development. They document that joints in the Coastal Plain sediments are high angle (70 to 90 degrees), and commonly have fillings of clay, and coatings of clay, limonite, or manganese. Preliminary data indicates the joints have strikes that vary from 0 to 360 degrees, with several local, but no well-defined regional preferred orientations.

4 EVALUATION OF SUSPECT FEATURES

4.1 Field Methods

Approximately 30 exposures in the Tertiary sediments of the Savannah River Site uplands were investigated (Figure JA.5). All exposures are manmade railroad or road cuts, and consequently tend to be located at the crests of hills. Most exposures reveal profiles 1.8 to 3.0 m (6 to 10 ft) deep, with the largest cuts up to 6.1 m (20 ft) deep. Many of the exposures previously were described as containing "clastic dikes" (D'Appolonia, 1982). Some of the exposures are located proximal to suspected subsurface faults (Stieve and Stephenson, 1995), whereas others are not (Figure JA.5).

The features at each outcrop were described in terms of their geometry, width, depth, filling material, and host material. Particular attention was given to their orientation and to the relationship between the features and the soil profile. The orientation of planar to slightly curviplanar features were recorded. Where a good three-dimensional exposure of the feature did not exist, it was partially excavated to obtain an accurate measurement. The spacial distribution of the features with respect to topography and known subsurface faults was also recorded to assess any possible geomorphic or structural associations.

4.2 Field Observations

Three types of post-depositional features were identified in the field: cutans, joints, and small-scale faults. Two types of cutans were identified: irregular and structurally controlled. The cutans are believed to be the features previously interpreted as "clastic dikes". The following discussion demonstrates that the features interpreted as cutans are not clastic dikes or other features characteristic of liquefaction.

4.2.1 Irregular cutans

These features are highly curviplanar in shape (Figure JA.6). They are invariably less than 5 cm (2 in.) wide, but may be traced for several feet. They exhibit no preferred orientation, though locally they are polygonal in plan view.

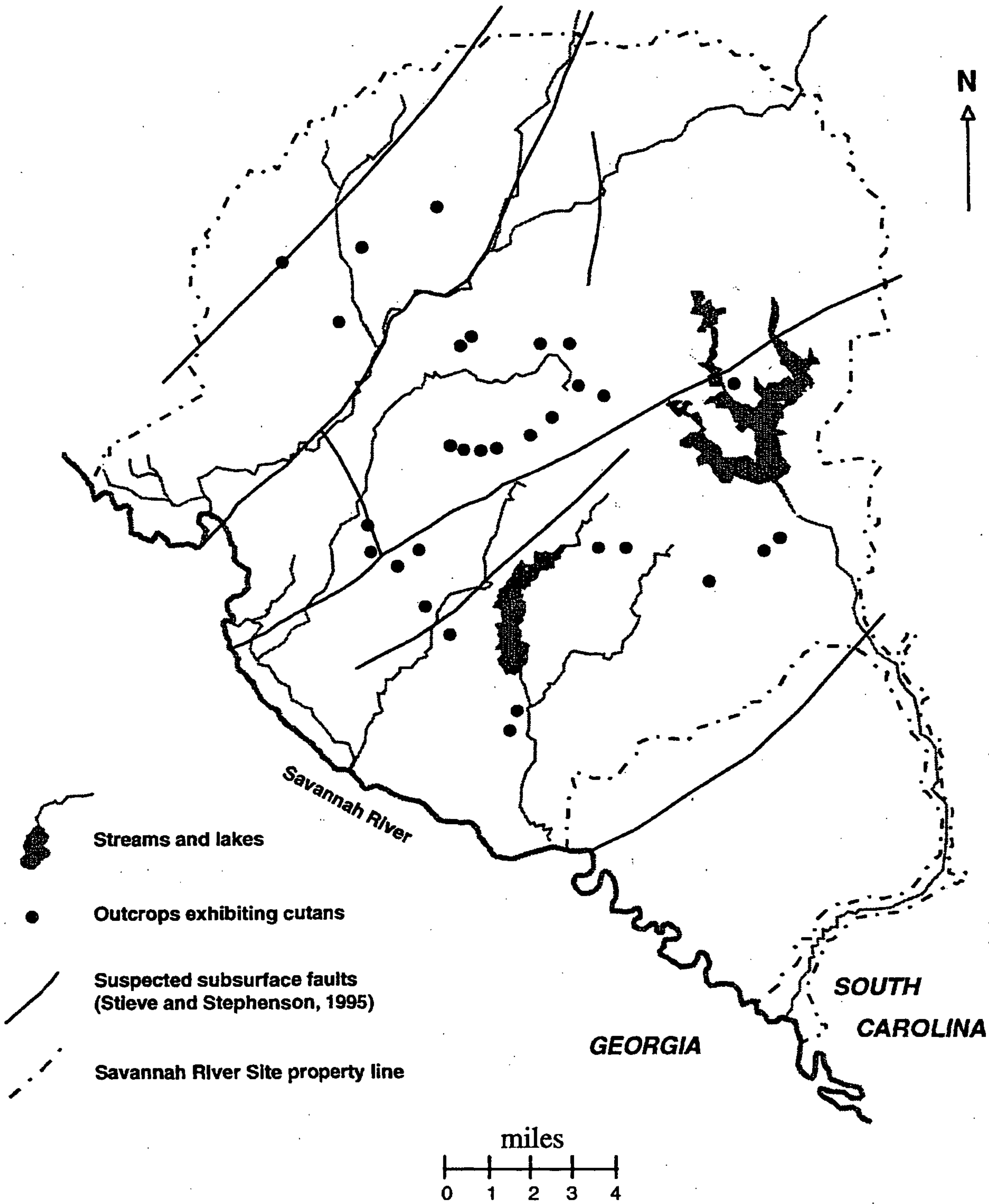

Figure JA.5 Locations of exposures of Tertiary strata exhibiting cutans at the Savannah River Site.

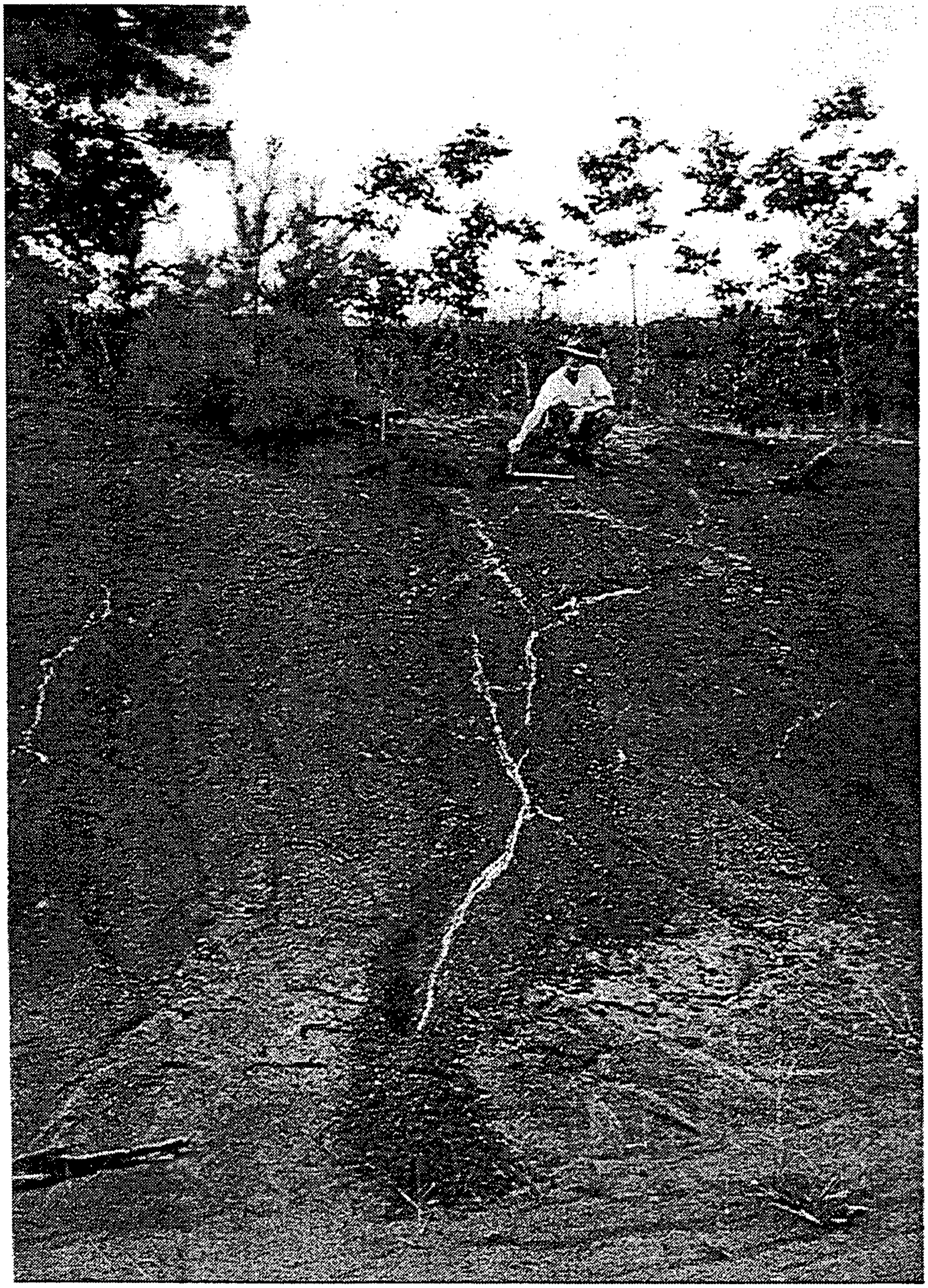

Figure JA.6 Irregularly shaped cutan. Shovel handle marks the top of the Teriary strata; geologist sits on fill material.

They are generally evident as bleached zones within more resistant limonite cemented rinds, and are very similar in appearance to a zone of more intense geochemical alteration that is present at the top of most exposures. Pedogenic processes at the top of this zone appear to overprint and erase both the fabric of the alteration zone and the highly curviplanar features. Pebbly horizons are not disrupted or offset by the features, clearly illustrating that the development of the features did not disrupt original bedding.

These features are ubiquitous in most exposures of the Altamaha Formation and Tobacco Road Sand. They are most pronounced in highly mottled, poorly sorted, clayey sand, and are not present in pebble conglomerates. They are most common within 1.5 m (5 ft) of the ground surface. The features are not more common or better developed near the upward projection of subsurface faults. Additionally, they are not more common in outcrops that contain faults.

The higher density of these features near the ground surface, and their similarity in appearance to the zone of more intense geochemical alteration that is present at the top of most exposures, suggests that these features are pedogenic in origin. The absence of offset eliminates the possibility that they are faults, and the undisrupted bedding within and across the features eliminates the possibility that they are seismically-induced liquefaction features. They are interpreted as an in-situ, pedogenic modification of the texture, structure, and fabric of the host material, and therefore are referred to as irregular cutans.

4.2.2 Structurally Controlled Cutans

4.2.2.1 Description

Structurally controlled cutans are sub-vertical and generally 7 to 15 cm (3 to 6 in.) wide, with the largest up to 35 cm (14 in.) wide (Figure JA.7). The lateral length is not well constrained, though two well exposed features in plan view extend for a minimum of 9 m (30 ft). Larger features commonly have splays that pinch out along strike. In almost all locations, the features extend to depths greater than the depth of the exposure (>2.4 to 4.6 m [>8 to 15 ft]), with no more features at the top of the exposure than at the base. In a few exposures, however, the features clearly and consistently thin downwards. Vertical pinch-outs are uncommon in the exposures, but do exist locally (Figure JA.8).

The features are composed of anastomosing bleached zones that generally have a slightly higher concentration of clay than the surrounding host material. Thin, discontinuous sheets of gray clay are locally evident near the center of the bleached zones. The features have limonite cemented rinds that are more resistant to weathering and make ridges that protrude up to 15 cm (6 in.) from the face of the outcrop (Figure JA.7). Internal rinds form between anastomosing bleached zones (Figure JA.7). The features neither displace nor disrupt marker horizons such as channels or bedding (Figure JA.9).

4.2.2.2 Relationship to Lithology and the Soil Profile

These features, like the irregular cutans, are ubiquitous on the Savannah River Site. They are present in most exposures of the Altamaha Formation and Tobacco Road Sand. They are present in sand, silt, clay, and coarse pebble conglomerate, though their appearance and density varies somewhat with lithology. The features are generally well defined and planar in sand, but branching networks of thinner features are typical in overlying or underlying clay layers. They are most pronounced in exposures of poorly sorted pebbly sand of the Altamaha Formation. They are least common in clay and clast-supported pebble conglomerates. The features are not more common or better developed near the upward projection of subsurface faults. Furthermore, they

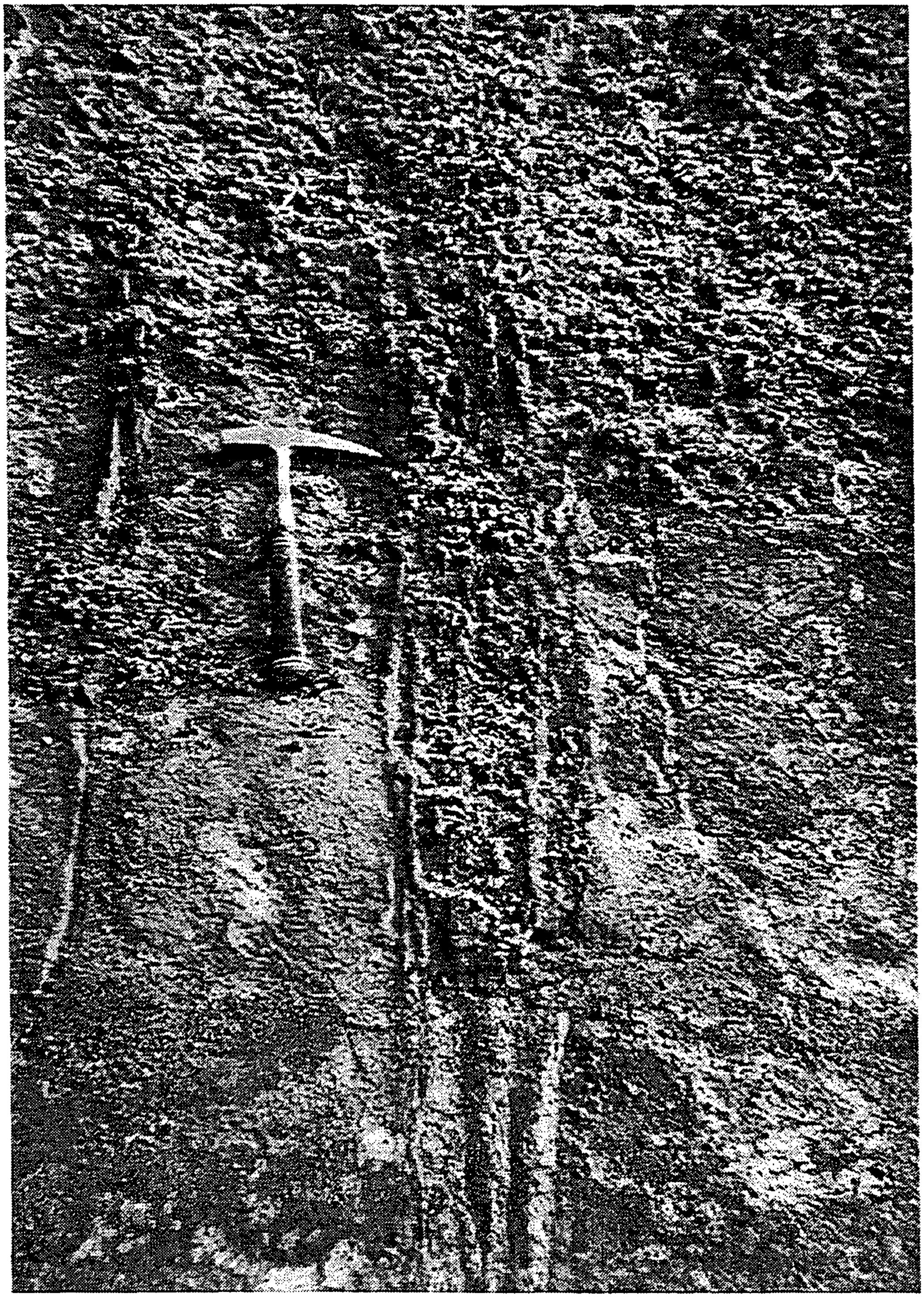

Figure JA.7 Large, structurally controlled cutan, with well developed limonite cemented rinds and vertical internal fabric.

Figure JA.8 Cutan (center) that pinches out at a depth of 3.0 m (10 ft). Note more intense geochemical alteration near the top of the exposure (under shovel) that is similar in color and texture to the cutan.

Figure JA.9 Cutans do not offset or disrupt bedding (note undisrupted pebbly horizon sub-parallel and below 15.2 cm [6 in.] ruler).

Figure JA.10 Cutans do not exhibit a clear cross-cutting relationship with the sub-horizontal fabric of the zone of more intense geochemical alteration that is present at the top of most exposures (above shovel).

are not more common in outcrops that contain faults. In the few exposures that do contain faults, each fault is generally associated with a cutan.

At the top of most exposures is a zone of more intense geochemical alteration (Figures JA.8 and JA.10). The thickness of the zone is generally 0.3 to 0.9 m (1 to 3 ft). The zone generally has a strong sub-horizontal fabric defined by orange and light gray mottled sheets (Figure JA.10). The sub-horizontal fabric of the zone is similar in appearance to the cutans. Some cutans terminate abruptly at the base of the sub-horizontal zone, whereas others clearly extend into the zone with no clear cross-cutting relationship. The two relationships exist in similar cutans of the same orientation in the same outcrop. Both the horizontal fabric of the zone of more intense geochemical alteration and the sub-vertical fabric of the cutans have been locally erased by pedogenesis in the upper 0.6 m (2 ft) of the soil profile (Figure JA.10).

4.2.2.3 Preferred Orientations

The strikes of the features at any given outcrop were observed to have one, two, or three preferred orientations (Figure JA.11). The preferred orientations vary laterally over several hundred feet, as illustrated in long exposures. In some locations, the orientation of the features changes gradually with position along the exposure (Figures JA.12 and JA.13). Several locations illustrate that the features are parallel to joints. At one location, the features have developed along small-scale faults.

As the preferred orientations of the features vary considerably around the Savannah River Site, the possibility was entertained that their orientations are controlled by the local topography. The local slope direction was determined from 7.5-minute topographic quadrangles, and plotted on lower hemisphere stereographic projections of the features at each outcrop. In all cases, the local slope direction is oblique to any preferred orientation (Figures JA.12 and JA.13). The obliqueness varies from about 30 to 70 degrees, and in no case is a preferred orientation normal to the local slope direction. Most exposures in virtually flat terrain have one or more preferred orientations (Figure JA.11). Furthermore, the preferred orientations on steep slopes are not consistently stronger than those on gentle slopes. There is no clear or consistent relationship between the preferred orientations of the features and the local slope direction or magnitude.

4.2.2.4 Interpretation

The similarity in appearance between the features and a sub-horizontal zone of more intense geochemical alteration that is present at the top of most exposures, plus the inconsistent cross-cutting relationship between them, suggests that the features and the zone of alteration formed by the same process. The position of the zone of more intense geochemical alteration at the top of the soil profile, and the overall downward thinning and pinchout of the features, suggests that they formed by pedogenesis.

There is no evidence of rapid injection of material into the features. Clay within the features is in discontinuous sheets, suggesting that it did not migrate rapidly into the features. The similarity of the material within the features and that of the host material, as well as undisrupted bedding within and across the features, demonstrates the features are not ice wedges, seismically-induced liquefaction features, or any other feature formed by deformation. The absence of offset across most (over 95%) of the features demonstrates that they did not develop as faults.

Strong preferred orientations at most exposures, parallelism with adjacent joints, and their occurrence along fault planes at one locality, suggests that the orientations of most of the

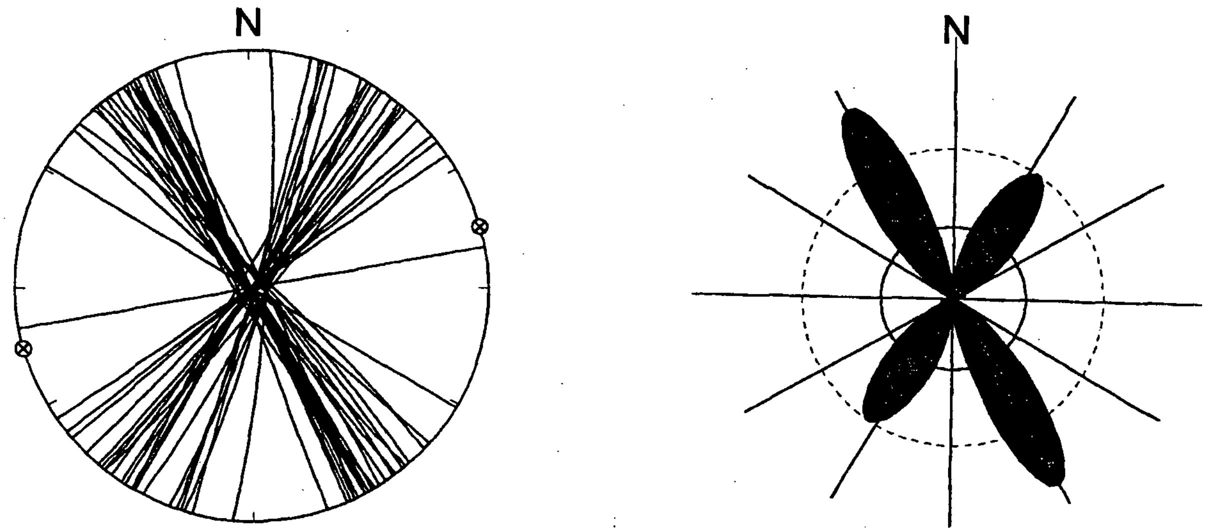

Figure JA.11 Example I of preferred orientations of structurally controlled cutans. Left: equal area, lower hemisphere, stereographic projection of cutans. Local topography is flat. Small circles containing an 'x' indicate the strike of the outcrop. Right: rose diagram of strikes of cutans, using a Gaussian smoothing function computed at 2.5° intervals. Solid circle shows the expected frequency for a uniform distribution of the number of data values. Dashed circle indicates the standard deviation of the frequency.

features are controlled by pre-existing structures. The features therefore are referred to as structurally controlled cutans.

5 CONCLUSION

Several features in Tertiary outcrops at the Savannah River Site have been described as "clastic dikes" by previous investigators. This investigation evaluated these features to assess if they have the diagnostic characteristics that recently have been documented for seismically-induced liquefaction.

No features indicative of liquefaction were observed. The "clastic dikes" referred to in previous reports are interpreted herein as cutans, which represent modification of the texture, structure, or fabric of the host material by pedogenic processes, either by concentration of particular soil constituents or in-situ modification of the matrix. The similarity of the material within the features and that of the host material, as well as undisrupted bedding within and across the features, demonstrates the features are not seismically-induced or aseismic liquefaction features. Furthermore, the undisrupted horizons demonstrates the features are not tree throws, root casts, animal burrows, or ice wedges. The absence of offset across most of the features demonstrates that they did not develop as faults. The features are interpreted to have developed through pedogenic processes based on:

- their similarity in appearance of the features to a sub-horizontal zone of geochemical alteration in the soil profile that is present at the top of most exposures.

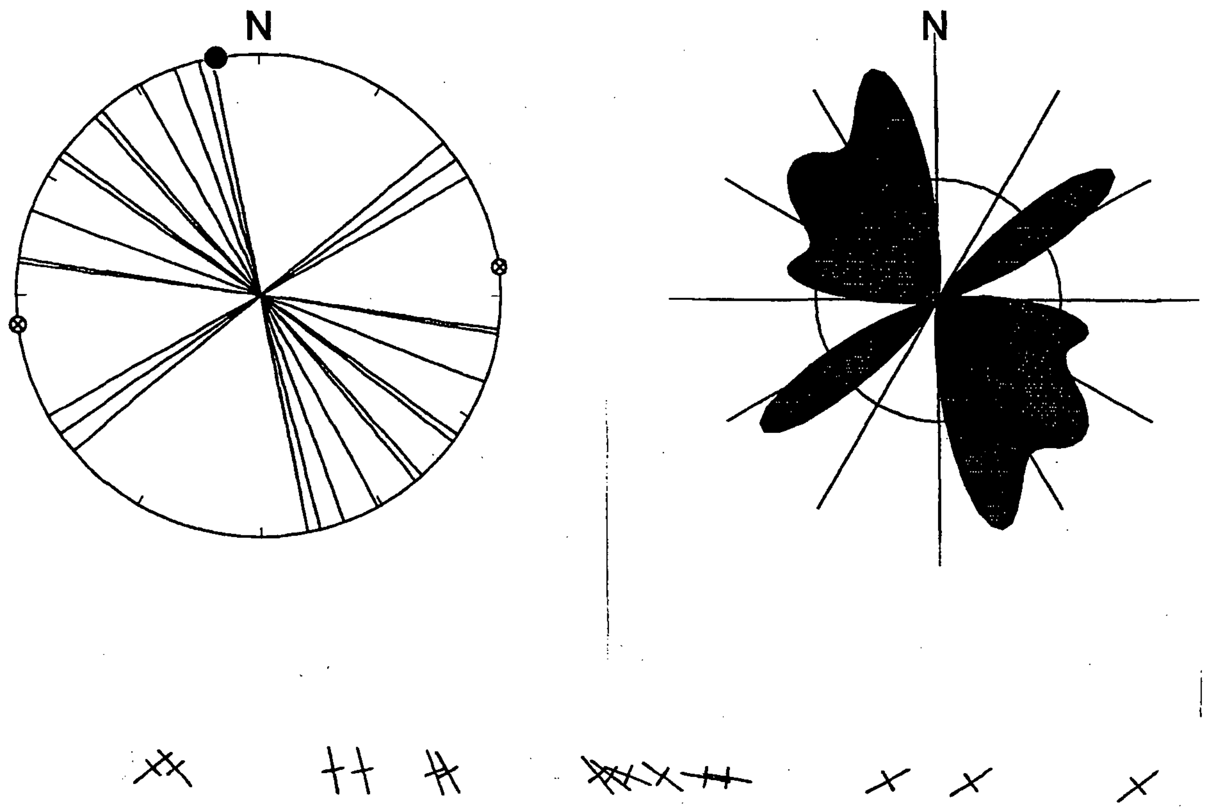

Figure JA.12 Example II of preferred orientations of structurally controlled cutans. Left: equal area, lower hemisphere, stereographic projection of cutans. Solid dot indicates the local topographic slope direction. Small circles containing an 'x' indicate the strike of the outcrop. Right: rose diagram of strikes of cutans, using a Gaussian smoothing function computed at 2.5° intervals. Solid circle shows the expected frequency for a uniform distribution of the number of data values. Dashed circle indicates the standard deviation of the frequency. Bottom: linear map illustrating a gradual and consistent change in orientation of the structurally controlled cutans with position along the exposure. Distance along map is 42.7 m (140 ft).

- an overall thinning and pinch-out of the features with depth, and their common occurrence in the upper parts of exposures.

Two types of cutans were identified: irregular cutans and structurally controlled cutans. Irregular cutans are irregular or polygonal in shape, and do not appear to be controlled by pre-existing structure. Structurally controlled cutans are planar to curviplanar and generally exhibit locally strong preferred orientations. At most exposures, structurally controlled cutans share a common orientation with adjacent joints. At one locality, cutans are present along fault planes. These relationships suggest that the position of the cutans is controlled by pre-existing structures.

The age of the cutans was not established under this investigation. Two hypotheses are identified.

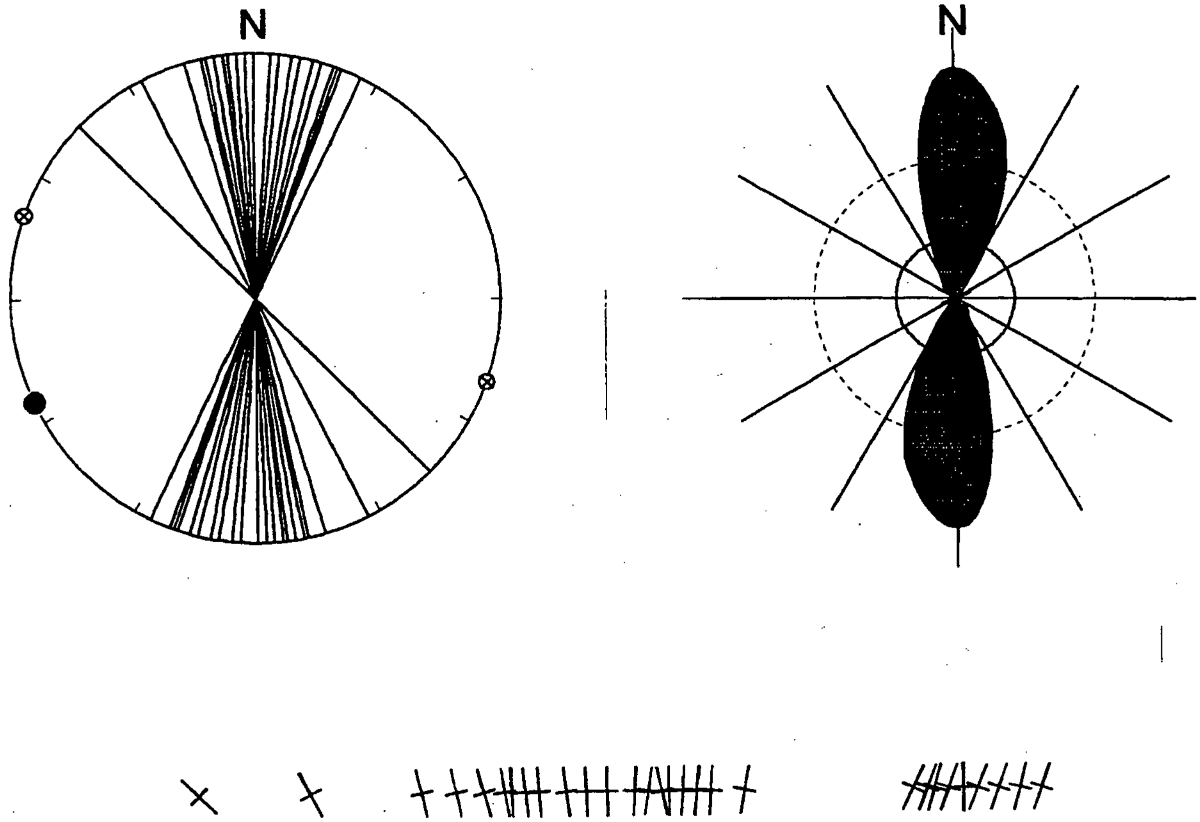

Figure JA.13 Example III of preferred orientations of structurally controlled cutans. Left: equal area, lower hemisphere, stereographic projection of cutans. Solid dot indicates the local topographic slope direction. Small circles containing an 'x' indicate the strike of the outcrop. Right: rose diagram of strikes of cutans, using a Gaussian smoothing function computed at 2.5° intervals. Solid circle shows the expected frequency for a uniform distribution of the number of data values. Bottom: linear map illustrating a gradual and consistent change in orientation of the structurally controlled cutans with position along the exposure. Distance along map is 76.2 m (250 ft).

The cutans may have formed in the Holocene, and are developing at present with the current soil profile. Alternatively, the features may have developed with a former (Tertiary) soil profile that has since been dissected by erosion. The fact that the features are more common in the younger Altamaha Formation (which crops out at higher elevations) than in the underlying deposits may support the latter hypothesis.

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FAULTS AREN'T ALWAYS WHAT THEY'RE CRACKED UP TO BE

William R. Cotton[1]

1 INTRODUCTION

The ability to distinguish between a fault that is capable of generating a large damaging earthquake and one that is non-seismogenic is of primary concern to the siting of engineered facilities. Although this task seems to be a simple academic exercise based largely on regional studies of historic seismicity, structural geology and paleoseismology, it becomes considerably more difficult when a fault is observed in a field exposure where the regional framework is missing. As we get closer to the geologic feature that is under investigation, we have an increasingly difficult time defining its regional significance. Paleoseismic studies require detailed analysis of the faulted microstratigraphic record, thus demanding that the field geologist work in close quarters with the fault in question. The more "micro" the field studies become, the more difficult it is to establish that the faulted structure exposed in a trench is, indeed, tectonic in origin. In most cases, experienced paleoseismologists select their exploratory sites in areas that have a clear pattern of historic seismicity and well-defined geomorphic evidence of young faulting, leaving little doubt that the faults encountered are tectonic. Demands of project siting, however, often require that exploratory trenches be placed in localities that may be influenced by landslide processes, which are frequently capable of developing a variety of physical features that mimic those produced by tectonic faulting.

In this discussion, the term **"fault"** is used in a descriptive sense and does not imply a particular mode of origin. Faults in the geologic record that are of interest to us are fractures in rock and soil along which shearing and dislocation have taken place. Faults that result from tectonism are referred to as "tectonic", while "nontectonic" faults have gravity as the principal driving force responsible for their origin. The tectonic faults that we search for in the microstratigraphic record are those that result from large-magnitude surface faulting events. Potentially damaging moderate-magnitude (M 6±) earthquakes commonly do not significantly rupture the earth surface and thus are not well defined in the microstratigraphic record.

Landslides, which are capable of producing nontectonic faults, are classified by material types and type of movement (Varnes 1978). Movement includes falls, topples, slides, lateral spreads, flows and complex (i.e., combination of two or more principal types of movement). Of these, only **slides, lateral spreads** and **flows** produce discrete shearing of earth materials, and the development of rupture surfaces both within the body and at the boundaries of the landslide.

Every pattern of tectonic faulting can be duplicated by sliding. Landslides exhibit a wide variety of scales, material types, internal structural complexity and boundary faults. The principal subsurface contact of a sliding mass is the basal shear surface, commonly referred to as the basal rupture surface or the slip surface. In general, it forms a continuous surface of faulting that extends up the side flanks, toe and head of the slide. The fact that slides are surrounded on all sides by more stable hillside terrain requires that tensional, compressional, and shear stresses develop along various segments of the slide boundary, causing strain to be displayed by normal faults in the head area, thrust faults at the toe, and strike-slip faults

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along the flanks. These relationships, where the secondary process of gravity sliding produces well defined rupture surfaces, are a constant and predictable product of sliding land masses.

The scale of the sliding mass does not appear to exert any strong control on the type of faulting found at the slide margins. The type of earth materials, ground-water regime, and the amount and rate of movement appear to be the primary factors that govern how well the boundary faults develop, evolve and are preserved. Some nontectonic shear surfaces may look like faults of tectonic origin because they are within the bedrock record, are regional in extent, and often have different slip histories. It is extremely important for a field geologist not to overlook the fact that gravity-driven sliding processes have taken place on immense lithospheric scales throughout geologic history. Gravity sliding of extremely large size, associated with worldwide mountain systems like the U.S. Cordillera and the Appalachian mountains, is well documented. The concept of early Paleozoic gravity sliding has been suggested as a likely mechanism for the emplacement of Taconite allochthons from the central Appalachians to Newfoundland (Voight and Cady, 1978). Likewise, low-angle Tertiary faults in the Whipple Mountains area of southeastern California have been described as part of an area of major imbricate, normal faulting, accompanied by pronounced rotational tilting of Tertiary strata (Davis, 1980). These normal faults are interpreted to be associated with a low-angle basal detachment surface that sloped to the northeast and accommodated the gravity sliding prior to the domal uplift and warping of the region. Similar examples of massive intra-lithospheric detachment have been described in the U.S. and Canadian Rockies. A classic example is Heart Mountain, a large slab of flat-lying Paleozoic strata and Eocene volcanic rocks that slid on the Heart Mountain detachment fault to its present position 40 to 50 km east of the Wyoming Rockies (Pierce 1987, Prostka, 1978). Alternatively, there are subduction complexes such as the mélange terranes of the Mesozoic Franciscan Complex wherein extensive slabs of coherent bedrock slide material (i.e., olistostromes) are bounded by highly sheared rock of tectonic origin. This condition exists in a vast sequence of rocks that forms the core of the northern California Coast Range, where rupture surfaces of both tectonic and nontectonic origin are intimately associated. Most of these regional detachment surfaces have undergone subsequent deformation and deep erosion, further masking their origins. At a single exposure of one of these faults, a field geologist would be incapable of distinguishing whether the fault is tectonic or not. Only with extensive field studies can these large shear surfaces be recognized with sufficient detail to define their nontectonic origin. Although recognition of extremely large-size gravity sliding is important, the primary focus of this paper is to address those slide masses that range in size from small surficial slumps of a few cubic meters to large deep-seated bedrock landslides that measure several kilometers in their largest dimension and are represented in the modern landscape.

An observation should be made about methods of scientific inquiry and the geologists who attempt to employ them. We tend to see what we are trained to see. Because landsliding and hillside processes often are not emphasized in our academic training, geologists may not consider the products of these processes, thus tending to interpret old landslide ruptures as faults of tectonic origin. A rupture surface viewed in a natural exposure or exploratory trench should always be considered to have either a tectonic or a nontectonic origin. Prudent and proper site analysis should always consider sliding as a possibility. S. A. Schumm, in his book *To Interpret the Earth: Ten Ways to be Wrong* (1991), addresses this topic in his discussion of "**convergence**", a situation wherein different processes and different causes produce similar effects. In his discussion of geologic extrapolation he points out that faulting, which is usually the result of tectonic activity, can also be developed by mass movement processes. He notes, for example, that ridge-top graben-like features (i.e. sackungen) can be formed by lateral spreading away from a topographic high, and that thrust faults situated along the toe of high mountain slopes sometimes result from either tectonic activity or sliding. These are two examples of convergence in the cause of faulting found at the top and bottom of mountain slopes, where tensional or compressional strain caused by sliding is capable of producing tectonic-like faulting.

Multiple working hypotheses should be the standard method of operation for all objective, unbiased paleoseismologists. As a means to overcome the bias of our training, it is important always to seek data that can disprove or falsify a hypothesis. Geologic models that we conceive to explain observations in an exploratory trench should always be tentatively applied. A means we can use to test a paleoseismic model is to interpret faulted field exposures or the faulted microstratigraphic record, as seen in an exploratory trench, within a context of regional seismicity, bedrock geology and geomorphology.

Finally, there is a need to recognize that a fundamental difference in goals exists between the technical responsibilities of a geologic consultant and that of a research professional. Research geoscientists often work at regional scales and attempt to present hypothetical models to explain their conclusions to fellow scientists in the form of published maps and journal articles. In contrast, geologists working at the applied levels, where siting considerations are essential to the safety of engineered facilities, start with consideration of models generated by their research colleagues, but have as a primary goal the development of a site specific data base from which risk is judged and mitigation measures are developed. The technical gap between the products of these two approaches is often huge. At times it is very difficult to correlate geologic data from site assessment investigations with expectations derived from the regional framework developed by research professionals. This kind of discrepancy sometimes requires geologists and engineers who are engaged in the siting of large engineered projects to test their conclusions against speculative geologic models cited in scholarly literature. Such a dialectic frequently bears little fruit.

2 DESCRIPTION OF LANDSLIDE FEATURES

Landslides produce fault features ranging from outcrop-scale shears that disrupt the Quaternary microstratigraphic record seen in exploratory trenches, to regionally deformed detachment surfaces as shown on small-scale maps. Landslides that disrupt Quaternary geology typically are associated with the modern landscape, whereas regional detachment features are commonly the product of much older landsliding events.

The number of landslide features that may cause confusion during paleoseismic research may be somewhat reduced, however, if the field geologist has a good understanding of landslide processes.

2.1 Landslide Terminology

All mountainsides are affected by gravity-driven processes that act to reduce the slope of the mountains and to achieve a state of dynamic equilibrium. The gravitational instability of a mountain is a transient, one-way process that reduces the potential energy stored in the topographic relief of modern landscapes. Tectonic fault rupture, being a product of primary processes, is not controlled by gravity.

Only a few types of landslide movements are capable of shearing rock and soil. These movements include **sliding, flowing** and **lateral spreading** (Figure 1).

Landslide terminology, although well established in the technical journals, is not generally

TYPE OF MOVEMENT			TYPE OF MATERIAL		
			BEDROCK	ENGINEERING SOILS	
				Predominantly coarse	Predominantly fine
FALLS			Rock fall	Debris fall	Earth fall
TOPPLES			Rock topple	Debris topple	Earth topple
SLIDES	ROTATIONAL	FEW UNITS	Rock slump	Debris slump	Earth slump
	TRANSLATIONAL	FEW UNITS	Rock block slide	Debris block slide	Earth block slide
		MANY UNITS	Rock	Debris slide	Earth slide
LATERAL SPREADS			Rock spread	Debris spread	Earth spread
FLOWS			Rock flow (deep creep)	Debris flow	Earth flow
				(soil creep)	
COMPLEX			Combination of two or more principal types of movement		

Figure WC.1 Landslide classification after Varnes, 1978. Shaded areas include those landslide deposits that are commonly associated with shearing of rock and soil and the development of tectonic-like boundary faults. Reprinted with permission from Landslides–Analysis and Control. Copyright 1978 by the National Academy of Sciences. Courtesy of the National Academy Press, Washington, D.C.

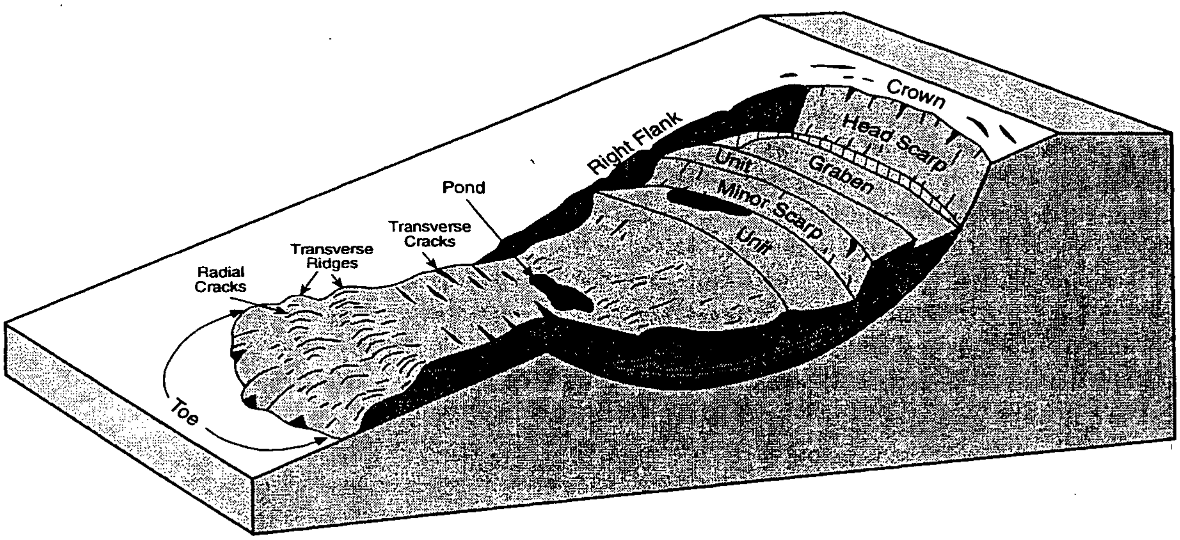

Figure WC.2 Diagrammatic landslide block diagram showing the main identifying geomorphic features produced by sliding. Geomorphic features such as arcuate head scarps, grabens, linear flanks, bulging toes, closed ponded depressions, hummocky topography and flat bench-like unit surfaces are characteristics of most landslides, and appear in marked contrast to the neighboring intact mountainside.associated with the faults. In all three areas, the origin of surface ruptures remains unresolved.

acknowledged and properly used by the geologist involved in active tectonic studies. The basic terms used to describe the geomorphic elements of a typical landslide are displayed in Figure 2..

2.2 Characteristics and Geometry

All of the fault types that we normally attribute to tectonic origin can be created by sliding. It is clear that the boundary of a landslide is delineated by faults that result from the entire spectrum of stress. In the head area of a landslide, where it is moving downslope and away from stable ground, tensional stresses result in extensional features such as horsts, grabens and other terrain features bounded by normal faults. Alternatively, in response to compression, the toe area will develop thrust faults and sinuous, bulging ridges that overrun the intact hillside or valley edge. Strike-slip faults, both left-lateral and right-lateral define the left* and right* margins of the landslide, respectively. The sense of slip changes by gradual transition from normal-slip at the head to left-and right-lateral strike-slip along the flanks to reverse-slip at the toe of the landslide mass (Figure WC.3). Likewise, many geomorphic features (e.g., ridges, sag ponds, deflected drainages) and microstratigraphic features (e.g., colluvial wedges, juxtaposed strata, buried soils) that are considered by paleoseismologists to be indicators of young faulting can also be produced by sliding processes. Although landsliding produces a variety of landforms and ground ruptures within the slide mass, this discussion is confined to the margins of the landslide, where the principal relative displacement with respect to the neighboring intact hillside is concentrated.

Three areas in the California Coast Ranges provide examples of confusion regarding the true nature of sheared rock and soil. These examples represent concerns with the origin of the three basic fault types. One location is associated with the General Electric nuclear reactor site in the Vallecitos Valley, south of Livermore. The second area is in the Summit Ridge area of the Santa Cruz Mountains. The third is situated in the Shelter Cove-Point Delgado area of the north coast of California, where surface faulting from the 1906 earthquake was mapped. Concerns at the G. E. reactor site centered around the origin of thrust faults running along the base of a series of hills and within close proximity to the reactor container. At Summit Ridge, a broad band of extensional ground fissures, many with a component of left-slip, developed along the crest of the ridge during the 1989 Loma Prieta earthquake. At both the Summit Ridge and Livermore areas, extensive paleoseismic studies were carried out in order to define the character of the faulting and to identify the potential hazards associated with the faults. In all three areas, the origin of surface ruptures remains unresolved.

Consultants retained by G.E. excavated a large number of exploratory trenches, drilled many holes, and extensively mapped the mountainsides before concluding that a series of thrust faults mapped at the mountain front were caused by gravity driven processes and represented the toe of a large paleolandslide. Government geologists disagreed, proposing a geologic model that suggested the faults are tectonic. Their newly discovered fault was called the Verona thrust fault, along which the hills were believed to have been elevated and thrust over the adjoining valley floor (Meehan, 1984). Arguments about the origin of the thrust faults and the risk to the reactor from the renewed movement were heatedly debated for an extended period of time. The debate about the fault origin was never resolved, but the risk to the reactor container was considered to be low because the container was judged to have sufficient structural integrity to accommodate a design displacement event.

**The margins of a slide mass are defined as "left" and "right" with the viewer looking down slope from the top or head of the slide.*

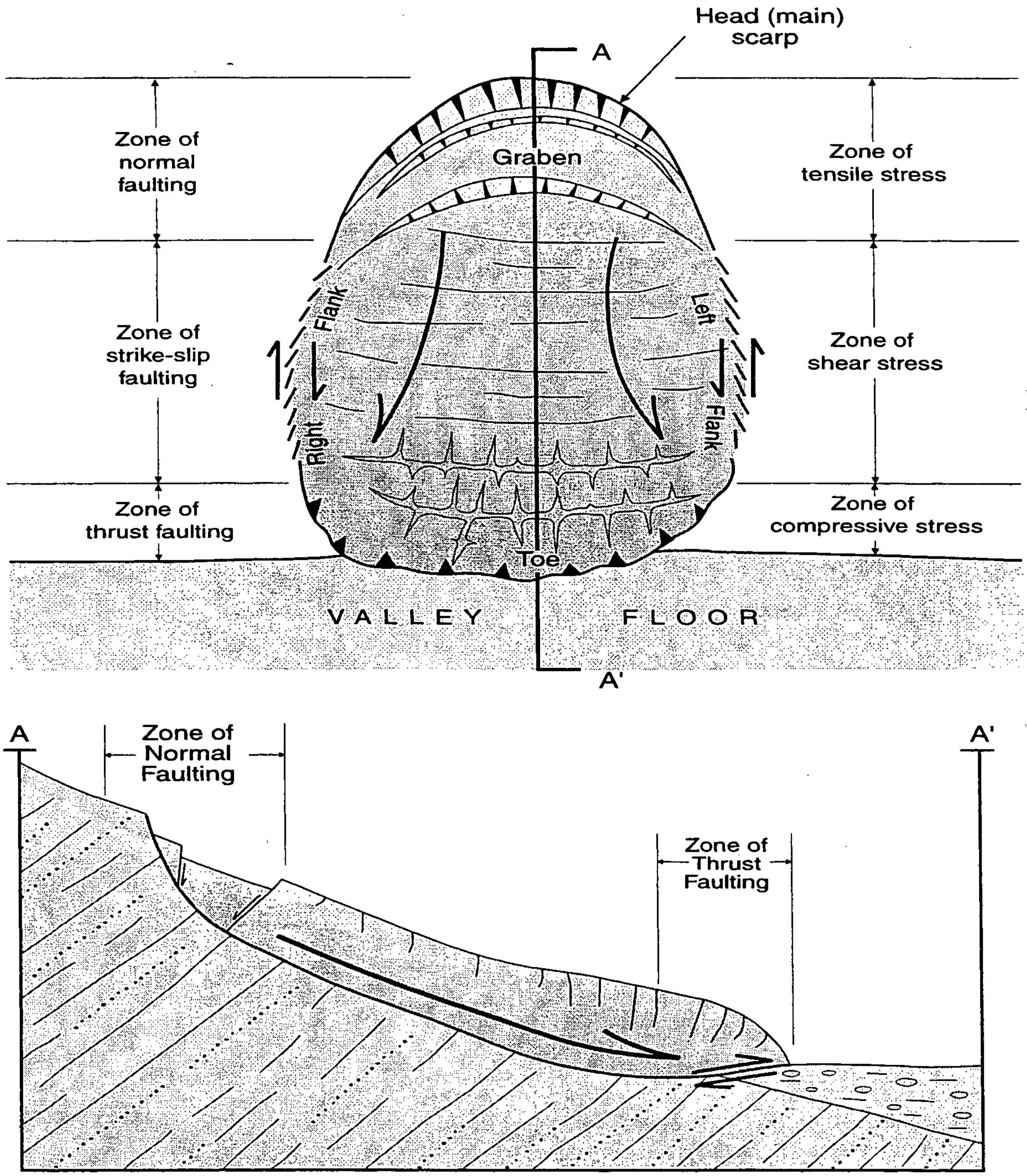

Figure WC.3 Map and cross section views of a typical landslide showing main landslide features, types of stresses and boundary faulting.

Near the northern end of the Loma Prieta earthquake rupture zone the crest of Summit Ridge was extensively cracked and ruptured by a northwest-trending band of coseismic ground fissures. The flanks of the ridge also experienced reactivation of pre-existing landslides, where the distribution and pattern of hillside ground cracking followed geomorphic boundaries of individual slide masses. The ground fissures along the crest of the ridge are attributed to several causes, including:

(1) tectonic extension and shaking-induced gravitational ridge-top spreading (Ponti and Wells, 1990; Hart, et al 1991)

(2) second-order tectonic faulting along weak structures primarily bedding within the Tertiary sedimentary rocks that underlie the ridge crest (Cotton, 1990; Zoback and Reches, 1990); and

(3) right-lateral shear across a broad zone beneath Summit Ridge, which represents the "Summit Ridge shear zone", interpreted to be the surface expression of the fault responsible for the Loma Prieta earthquake (Johnson and Fleming, 1993)

It is clear that two populations of ground fissures were produced during this seismic event. The numerous landslide deposits observed downslope on the steep flanks of the ridge have ground fissures along their margins that clearly indicate partial reactivation. On the other hand, the unique set of ground fissures confined to the ridge crest are not related to pre-existing slides. Instead, they show strong structural association with the strike of the Tertiary bedrock that underlies the ridge crest. Detailed field studies of the pattern, distribution and slip of these ground fissures (Johnson and Fleming, 1993) and paleoseismic studies (Cotton, Fowler and VanVelsor, 1990) strongly indicate that most of the ridge-crest fissures are tectonic. Kinematic models based on the ground fissure orientation and slip direction (Johnson and Fleming, 1993) and the extension of the ground fissures into the subsurface where they can be traced to normal fault displacements within the shale bedrock (Cotton, Fowler and VanVelsor, 1990) are the chief field observations that support a tectonic origin. The shaking-induced gravitational ridge-top spreading model is suspect primarily because preliminary geotechnical analysis indicates that the duration of strong ground shaking was probably insufficient to generate massive area-wide spreading of the ridge crest, but once again an ambiguity of fault genesis persists.

Shortly after the 1906 earthquake along the San Andreas fault, the main trace of ground rupture in the Shelter Cove – Point Delgado area was mapped by F.E. Matthes (Lawson, 1908) for approximately 5 km as a very well-defined, narrow belt of diagonal fractures, raised soil and linear scarps associated with "characteristic fault topography", including elongated ponds, spur ridges and scarps two to three meters high. Another observer at the same time, A.S. Eakle, inspected all of the surface faulting features and their relation to the geomorphic features and concluded that they were the result of a huge landslide that had been apparently reactivated as a result of the earthquake (Lawson, 1908). Recent geologic mapping of the fault zone in this area by the U.S. Geological Survey has also led to the conclusion that the region is probably a very large landslide complex (R.J. McLaughlin, personal communication), but some geologists still interpret the ruptures as tectonic. An obvious problem with assessing fault origins in landslide-prone mountainous terrain is the probability that ground rupture can result from both tectonic faulting and landslide reactivations.

3 CRITERIA AND/OR METHODS TO DIFFERENTIATE TECTONIC FROM NONTECTONIC DEFORMATION

As noted above, landsliding is capable of producing all of the fault types and many of the associated landforms that commonly are attributed to tectonic faulting. Because it is virtually impossible to determine correctly if a fault is tectonic or nontectonic solely on the basis of an exposure in a roadcut, trench wall or natural rock outcrop, we must follow a procedure that will often reduce the uncertainty to tolerable levels. It is required that the field geologist step back from the exposure and consider the regional setting. The first basic task of any paleoseismic investigation is to compile and analyze the regional pattern of historic seismicity and bedrock and Quaternary geology. Once completed, the selection of a field site should be narrowed to segments of the fault that cut late Quaternary deposits that are capable of yielding a productive earthquake history. The best paleoseismic trench sites are those that have all or most of the following field conditions:

1) the fault is sufficiently well defined by geomorphic features that indicate that the slip history is concentrated along a narrow, single fault trace

2) faulting offsets late Quaternary deposits

3) datable horizons are likely to be included within the late Quaternary stratigraphy

4) the deposition rate was such that the ages of the deposits exposed in a four- to five-meter deep trench span enough time to record multiple earthquake episodes, and

5) the trench site appears to be free of landslide processes that may alter or destroy the paleoseismic record

Adhering to these site selection criteria will effectively reduce the chance that hillside processes have altered the microstratigraphic record. It is obviously desirable to avoid hillside regions where gravity driven processes are, or have been, active. In many cases of basic paleoseismic research, the geologist is investigating large active faults that have fault lengths of many hundreds of kilometers. Longer fault lengths usually increase the probability that a research site can be located that avoids hillside processes, yet one that provides all of the basic field criteria needed to conduct potentially fruitful research. This luxury, however, is not always afforded for site investigations pertaining to engineered facilities that are constrained by technical and/or socioeconomic needs.

The discussion below outlines the difficulties and problems encountered when attempting to differentiate between tectonic and nontectonic faulting. Topics begin with the regional setting and end with the microstratigraphic setting, the order by which a geologist should approach this important task of site characterization.

3.1 Seismological and Geophysical Criteria

Very large active slides, such as those associated with both subaerial and submarine volcanic centers, have on occasion recorded seismic activity. However, the density, pattern and historic record of this activity does not match that of active tectonic faults. A long-term record of historical seismicity, microseismicity, and geophysical and geodetic anomalies, so common to active fault systems, is not commonly characteristic of slide processes. Recent work on the sea floor surrounding the Hawaiian Ridge has delineated large landslide deposits that extend upslope to include a large part of the subaerial volcanic edifice (Moore, Normark and Halcomb, 1994). On the southeast flank of the island of Hawaii, the headscarp of one of these landslides is defined by normal faults that have produced the Hilina Pali and Holei Pali escarpments, and may extend as far back as the active volcanic rift zone of the Kilauea Caldera (Figure 4). The subsurface rupture surface may project down 10 km to near the base of the volcanic pile. Abrupt oceanward displacement of several meters of this landslide mass is believed to have produced the largest of Hawaii's historic earthquakes, in 1868 and 1975. The magnitude 7.1 Kalapana earthquake of 1975, centered on the southeast coast of Hawaii, has been attributed to an abrupt failure episode on the flank of the island (Eissler and Kanamori, 1987). It is important to recognize that rare instances exist when megalandslides are capable of generating large earthquakes, and producing geophysical and geodetic anomalies. It helps to remind us of the immense size and scale of some gravity-driven processes. There is little chance, however, that isolated earthquake events such as those in Hawaii will be confused with the dense, commonly linear pattern of historic earthquakes observed in many parts of the world.

3.2 Structural/Geologic Criteria

Large landslides are capable of disrupting geologic structures and earth materials on a variety of scales. Large tracts of bedrock, such as the Heart Mountain and Whipple Mountain systems, have been gravity-transported long distances, subsequently deformed and deeply eroded, thus becoming essential elements of the regional landscape and geologic framework (Figures 5 and 6). Many other bedrock landslides that are smaller and less mobile than these examples, nevertheless, clearly have modified the local geologic structure. In recent years, landslide specialists armed with new remote sensing techniques have described "megalandslides" as essential elements in the regional geologic setting of mountainous terrains. Most of these major slide masses have been largely overlooked and mistakenly mapped as the bedrock geology by field geologists. Published geologic maps beginning about a half century ago show a gradual evolution of our understanding regarding the presence of large landslides in the modern landscape and the role of gravity-driven processes in structural geologic analyses. For the most part, geologic maps published by state or federal agencies focused on bedrock geology, whereas Quaternary deposits, especially landslides, either were undetected or not mapped in sufficient detail. Very large landslide complexes, and especially paleolandslides, commonly are not recognized, either because of their size or subtle geomorphic expressions.

For example, the Big Rock Mesa area on the south slope of the Santa Monica Mountains of southern California until recently was not recognized as a large paleolandslide. In the late 1950s and early 1960s, the County of Los Angeles and the U.S. Geological Survey were engaged in a cooperative geologic study of the Malibu coastline along which Big Rock Mesa is centered. The mesa stands approximately 70 meters above sea level, extends 680 meters inland, and is approximately 1000 meters wide. At the back of the mesa a steep arcuate ridge surrounds its landward edge. The front edge of the mesa is characterized by a steep rocky cliff which descends to the Pacific Coast Highway and parallels the shoreline. This area was mapped by a team of highly experienced field geologists with many years of excellent field work in coastal California. Their final map, first published in 1974, shows the Big Rock Mesa to be intact rock with a relatively thin cover of landslide debris (Yerkes and Campbell, 1980). The surface of the mesa contains hummocky topography with closed depressions, all situated below a high, steep escarpment, suggesting that landsliding played a role in shaping the surface geomorphology of the mesa. The hypothesis that the mesa was a marine terrace remnant was rejected because little or no terrace deposits were observed to support this model. Also rejected was the notion that the mesa represented a deep-seated landslide that involved the entire mesa including the 70-meter-high cliff that defined the mesa's front. The physical character of the bedrock exposed in the steep cliff was clearly disturbed and structurally different from the intact bedrock on either side, but inasmuch as the entire region is structurally complex, the bedrock beneath the mesa was not mapped as a deep, massive landslide. In late 1983, the true nature of the mesa was revealed when the entire mass moved shoreword approximately one to two meters. Big Rock Mesa is now known as the Big Rock Mesa landslide.

A primary reason why the ancient landslide was not recognized was that the field geologists did not envision that there could be deep landsliding of massive, gritty sandstone with dipping beds opposed to the sliding direction. Landslides of such large extent were considered at this time to be rare. Because geomorphology of the mesa strongly indicated landsliding, geologists mapped a verneer of landslide debris on the mesa rather than identifying the entire mesa as a slide block. The message is that large landslide deposits are seen only by geologists who are trained to see them. Most geologic maps available today were not intended to show landslide deposits. Likewise, most field geologists today are not trained for this task. The good news is that field geologists can be trained, in a relatively short period of time, to identify and map landslide deposits.

Subsequent investigations of the Big Rock Mesa landslide determined that the depth of sliding ranges from 100 to 190 meters, that the landslide toes-out near sea level, and that the sliding was a reactivation of an ancient pre-existing landslide. Detailed field mapping, conducted after the 1983 reactivation, demonstrated that the total displacement of the Big Rock Mesa landslide is approximately 55 meters (Cotton, 1986). This was determined by recognition of structural and stratigraphic piercing points mapped along the flanks of the slide.

As demonstrated at Big Rock Mesa, the structural overprint of sliding creates anomalous relationships within the regional structural setting. Careful field mapping and analysis of the structure of late Quaternary deposits and the underlying bedrock can reveal a pattern of deformation that elucidates the neotectonic history of a region. A fundamental task in the early stages of a paleoseismic investigation is the development of a structural framework showing fold trends and fault types, that allows recognition of a consistent pattern of deformation that extends into the geologic past. This structural framework of the region when compared with the deformation seen in late Quaternary deposits, can allow identification of faults that are not compatible with regional tectonic pattern. In many cases, as in the Whipple Mountains and at Heart Mountain, careful regional mapping is required to identify immense slide masses. Faults that are inconsistent with the regional framework of deformation should be considered as suspect structures.

A unique feature of landslides is the listric nature of all of the faults that mark the slide margins. The basal rupture surface of a sliding mass defines the main detachment fault on which sliding takes place. This basal surface commonly is subparallel to the mountainside on which it forms. Thrust faults at the toe of a large landslide dip into the base of the mountain and bend upslope as they merge with the basal shear surface. Tectonic thrust faults, on the other hand, dip into the base of the mountain front and continue that orientation as they extend to seismogenic depths. Techniques to distinguish between these two fault types require costly field programs including subsurface drilling, field mapping and geophysics conducted by geologists having landslide detection and fault mechanics experience.

3.3 Geomorphic Criteria

Geomorphology is one of the most powerful tools for evaluating active tectonics, in general, and the selection of paleoseismic research sites, in particular. Landforms are essential elements in the modern landscape and their origin and modifications are important to the study of prehistoric earthquakes and paleolandsliding.

Tectonic geomorphology usually involves regional reconnaissance studies as well as detailed site specific evaluations. Regional studies are designed to provide insights into the area-wide influence of active tectonic deformation on the erosional and depositional processes of river or hill slope systems (Keller, 1986). These studies include calculation of indices such as the stream-gradient index, mountain-front sinuosity and ratio of valley-floor width to valley height. Site specific studies attempt to establish the chronology of faulting by evaluating temporal relationships between landforms, earth materials, surface processes, and faulting.

Certain landforms are characteristic of active tectonics. For example, tectonic strike-slip faulting commonly produces landforms such as linear valleys, offset or deflected streams, sag ponds, pressure ridges, scarps, shutter ridges, and horst and graben structures. Whereas many of these features can be produced by sliding of mountainsides, it is unlikely that sliding could be responsible for their development at large distances from the front of mountainous terrain. Although some geomorphic features typical of tectonic strike-slip fault zones are produced by shearing along the margins of active landslides, modern landslide-produced features usually are local in extent and are oriented perpendicular to the slope (Johnson and Fleming, 1993). Conversely, major strike-slip ruptures of tectonic origin produce patterns and landforms that are similar to those seen along the flanks of large active landslides (Figure 7).

Geomorphic features produced by dip-slip faulting such as normal faulting and thrust faulting also are developed by tensional stresses in the head of a slide and compressional stresses in the slide toe region. Unlike large strike-slip faults that have near-vertical fault planes and rather linear traces, normal faults and thrust faults have a less steeply dipping fault surface, commonly producing a sinuous map pattern. Because the toes of large mountainside landslides often produce sinuous fronts and bulging masses of slide material, they may look identical to mountain-front, tectonic thrust faults (Figure 8).

It is sometimes difficult to separate tectonic thrust faults from nontectonic thrust faults located near the base of a modern mountain front, especially where massive and deep-seated paleosliding and active tectonic uplift have occurred. This setting is common along active mountain fronts in a compressive tectonic regime. Being able to recognize landslides in the modern landscape is based in part on understanding how mountainsides evolve through time, and the geomorphic expression of various types of sliding masses. Identifying landslides is based largely upon recognition of characteristic geomorphic features. If a slide is large and has moved in the not-too-distant geologic past, it will be readily expressed in the topography, and thus easy to identify. As time passes, surface processes tend to subdue the sharpness of landslide physiography, thereby helping to disguise the slide mass. Large ancient landslide deposits that have been stable for long periods of geologic time are sometimes difficult to identify by geomorphic character and are commonly unrecognized because of their subtle appearance. Once overlooked, they can cause confusion regarding the origin of bounding faults that may be observed at some later time in a paleoseismic excavation or natural exposure.

There is need to develop techniques and criteria to recognize regions that have been modified by sliding. It is clear that at the reconnaissance stage of any regional paleoseismic analysis, all slopes should be considered as suspect terrain. The most common product of a reconnaissance level geomorphic analysis is a simple "**Landslide Inventory Map**". This map normally is prepared without the benefit of detailed mapping to verify its accuracy. This powerful tool provides a

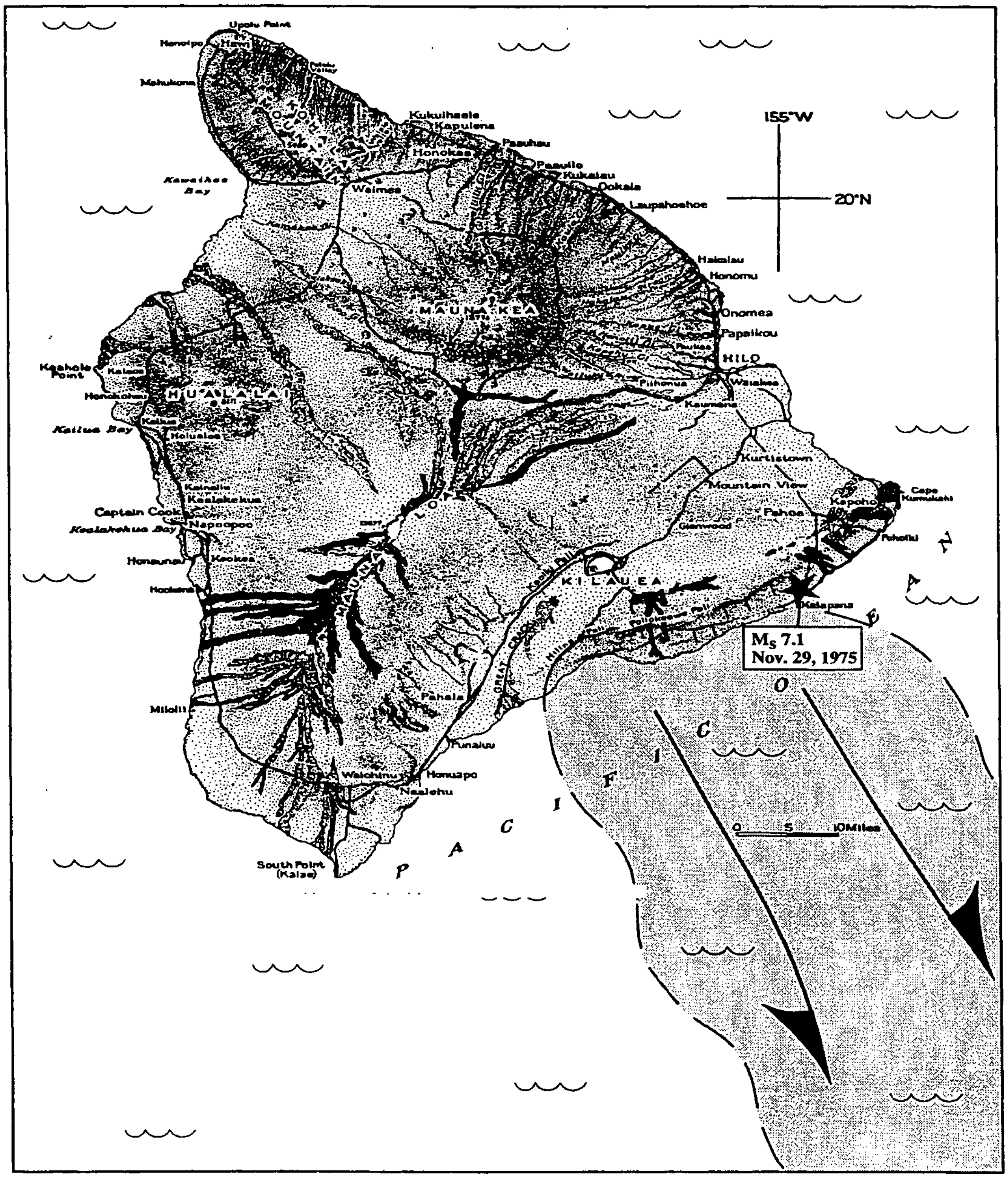

Figure WC.4 Map of the Island of Hawaii showing the probable inland extent of a large submarine landslide (pink area) (modified from MacDonald and Abbott, 1970). The Kalapan earthquake of November 29, 1975 was caused by gravity sliding. Permission to use this copyrighted material is granted by the University of Hawaii .

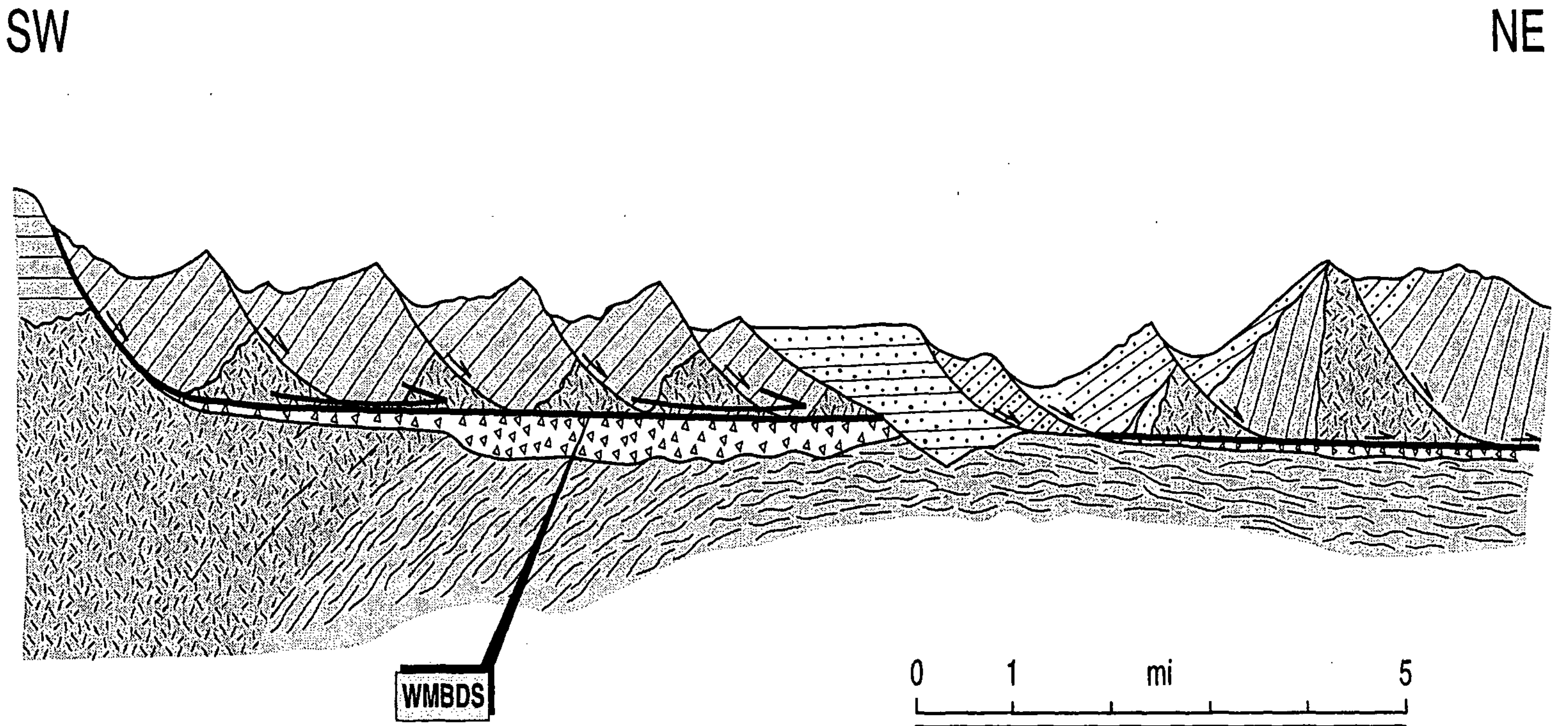

Figure WC.5 Diagrammatic cross section across the Whipple Mountains of Southeastern California illustrating the Whipple Mountains basal dislocation surface (WMBDS) during middle Miocene. Reprinted with permission from G.A. Davis, Problems of Intraplate Extensional Tectonics, Western United States, Continental Tectonics, Studies in Geophysics. Copyright © 1980 by the National Academy of Sciences. Courtesy of the National Academy Press, Washington, D.C.

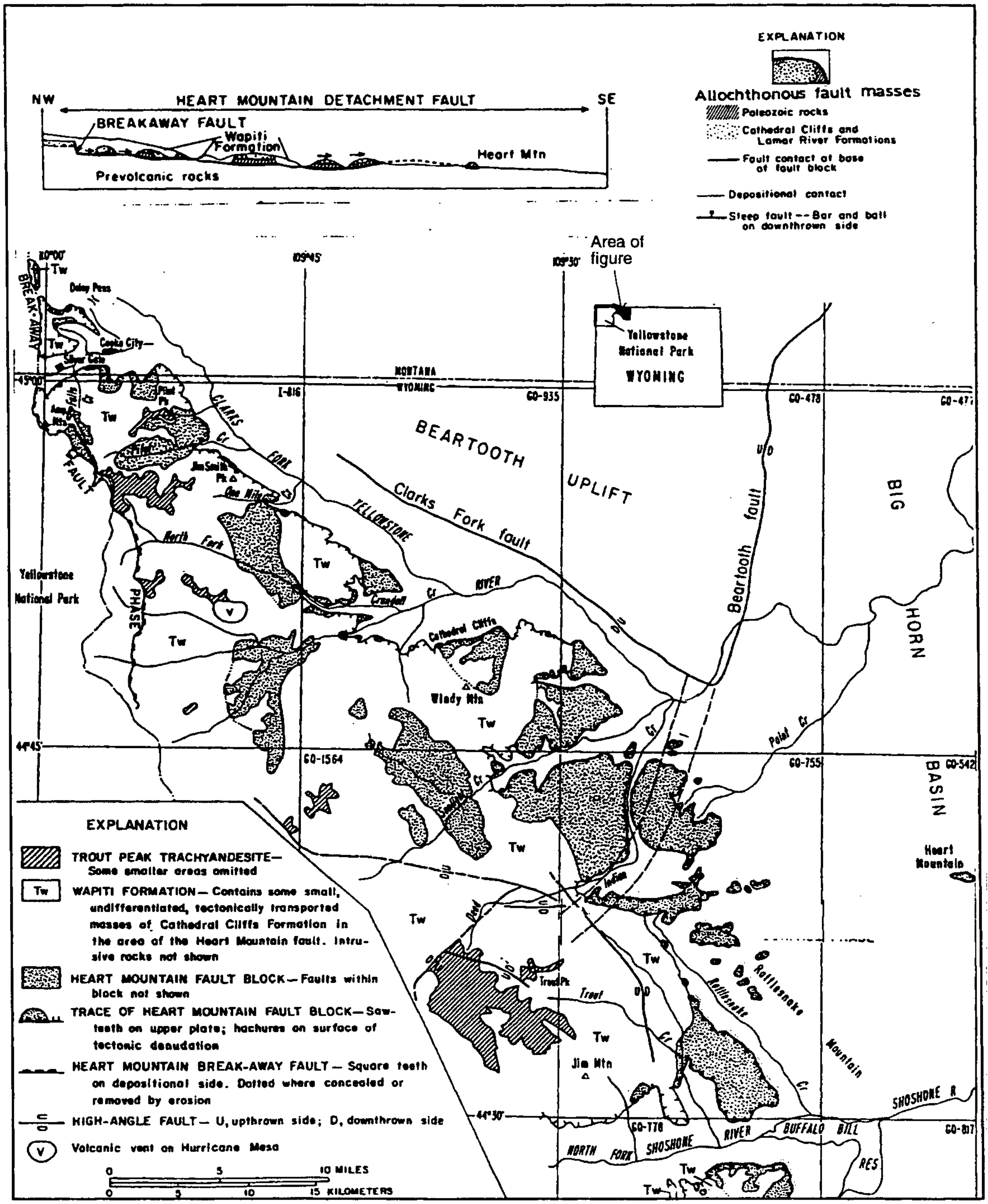

Figure WC.6 Map of the Heart Mountain fault and the distribution of upper plate Paleozoic carbonate and older volcanic rocks, and break-away escarpment. From Geological Society of America Bulletin, Pierce. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1987 Geological Society of America.

preliminary assessment of the distribution of failed mountainsides as well as the likely mechanisms of failure. Production of a landslide inventory map follows a basic two-step process:

(1) inspecting topographic (7.5 minute) maps to delineate suspected landslides on the basis of their geomorphic expression as defined by the topographic contour patterns (e.g. arcuate headscarps, hummocky terrain, ponded water, bulging toe, displaced drainages, etc.), and

(2) interpretation of stereoscopic aerial photographs to verify the existence of the landslide features shown on the topographic mapAs might be expected, such mapping efforts result in maps of highly variable quality and accuracy. Development of high-quality inventory maps is largely influenced by the scale and the accuracy of the topographic map, the scale and vegetation coverage on the aerial photographs, the state of development and the preservation of the landslide deposits and, last but not least, the ability and training of the person making the landslide inventory map. Landslide deposits can be identified on the maps as "Definite", "Probable" and "Questionable" as a way of reflecting the judgment of the person making the map. Once completed, the map can be compared to a geologic map of the same region to determine possible relationships between the location of sliding and the slope, drainage system, Quaternary deposits, bedrock types, and structure. Large landslides are capable of altering all of these elements. When selecting paleoseismic sites in mountainous terrain, development of a landslide inventory map is a critical task that is equally as important as defining the regional seismic and geologic setting.

3.4 Microstratigraphic Criteria

Criteria that can be used to differentiate tectonic and nontectonic faults in the stratigraphic record are based largely on deformed late Quaternary deposits. Being the products of young near-surface processes, these deposits are the targets of both tectonic geomorphologists and paleoseismologists. Both tectonic faulting and landsliding commonly provide a stratigraphic record of faulting that may, or may not, yield subtle information on the origin of the faulting. Deciphering the microstratigraphic record poses a great challenge (Figure 9).

Sedimentary packages identical to those derived from newly-formed tectonic fault scarps can also be produced by sliding. The most dramatic examples are from the headscarp areas and the toe regions of large landslide masses. Normal fault scarps flanking graben structures running along the upper slopes of mountainsides, and sinuous thrust fault scarps developed at the base of the mountain fronts will, over time, produce fault scarp debris on their down-thrown blocks. The resulting colluvial wedge is a record of the scarp-forming event, and repeated ground faulting will commonly produce a series of stacked colluvial wedges (McCalpin, 1995) (Figure 10). Requirements for colluvial wedges to form are degradation of the scarp by erosion, subsequent deposition of scarp debris along the base of the scarp, and the preservation of these deposits. The fault scarp, however, need not be tectonic.

Tectonic faults seen in a trench wall commonly truncate sedimentary layers (Figure 9). These faults frequently have well developed highly sheared surfaces that are slickensided, striated and associated with clay gouge and crushed rock or soil fabric. Although there is essentially no published literature describing similar features formed from sliding, geotechnical investigation reports describing exposures made for engineered or stabilization projects frequently provide clear descriptions of similar nontectonic features. Subsurface investigations of large landslides using large-diameter borings that are entered and geologically logged, describe essentially all of the same features that are generally attributed to tectonic faulting. Subsurface programs commonly encounter basal rupture surfaces characterized by

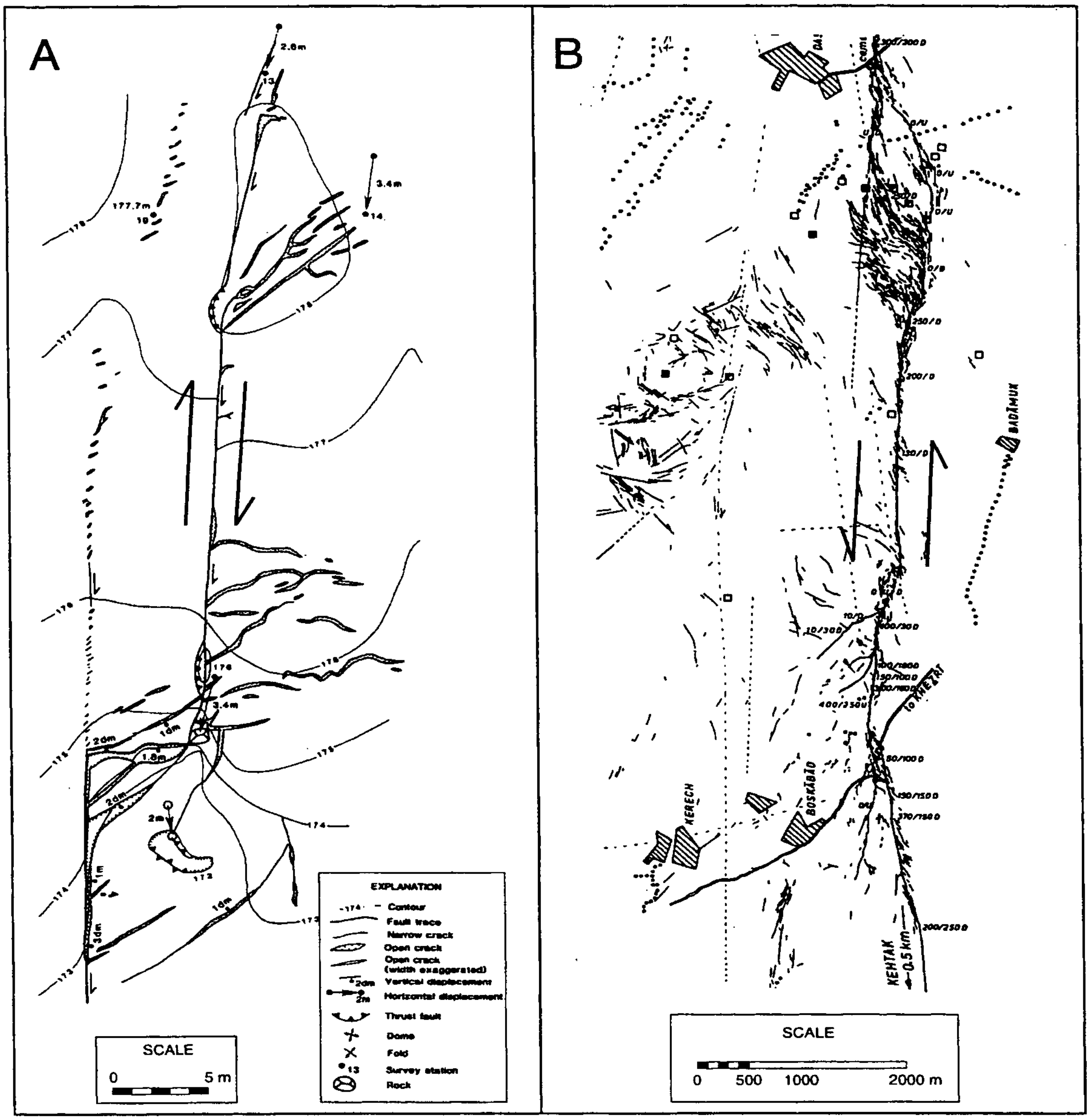

Figure WC.7 Similarity in surface fault pattern between a strike-slip landslide fault and a strike-slip earthquake fault. A. Planetable map of the right flank of the Aspen Grove landslide in 1984. Surface faults and ground deformation resemble those developed along right-lateral strike-slip faults during large earthquakes. Reprinted from Engineering Geology, v. 27, Fleming, R.W., and A.M. Johnson, Structures Associated with Strike-Slip Faults that Bound Landslide Elements, p. 67-70, © 1989, with permission from Elsevier Science. B. Map of the ground fractures produced by the August 31, 1968 Dasht-e Bayaz earthquake in Iran (from Tchalenko and Ambraseys, 1970). Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1970 Geological Society of America. The difference in map scales is 1:200.

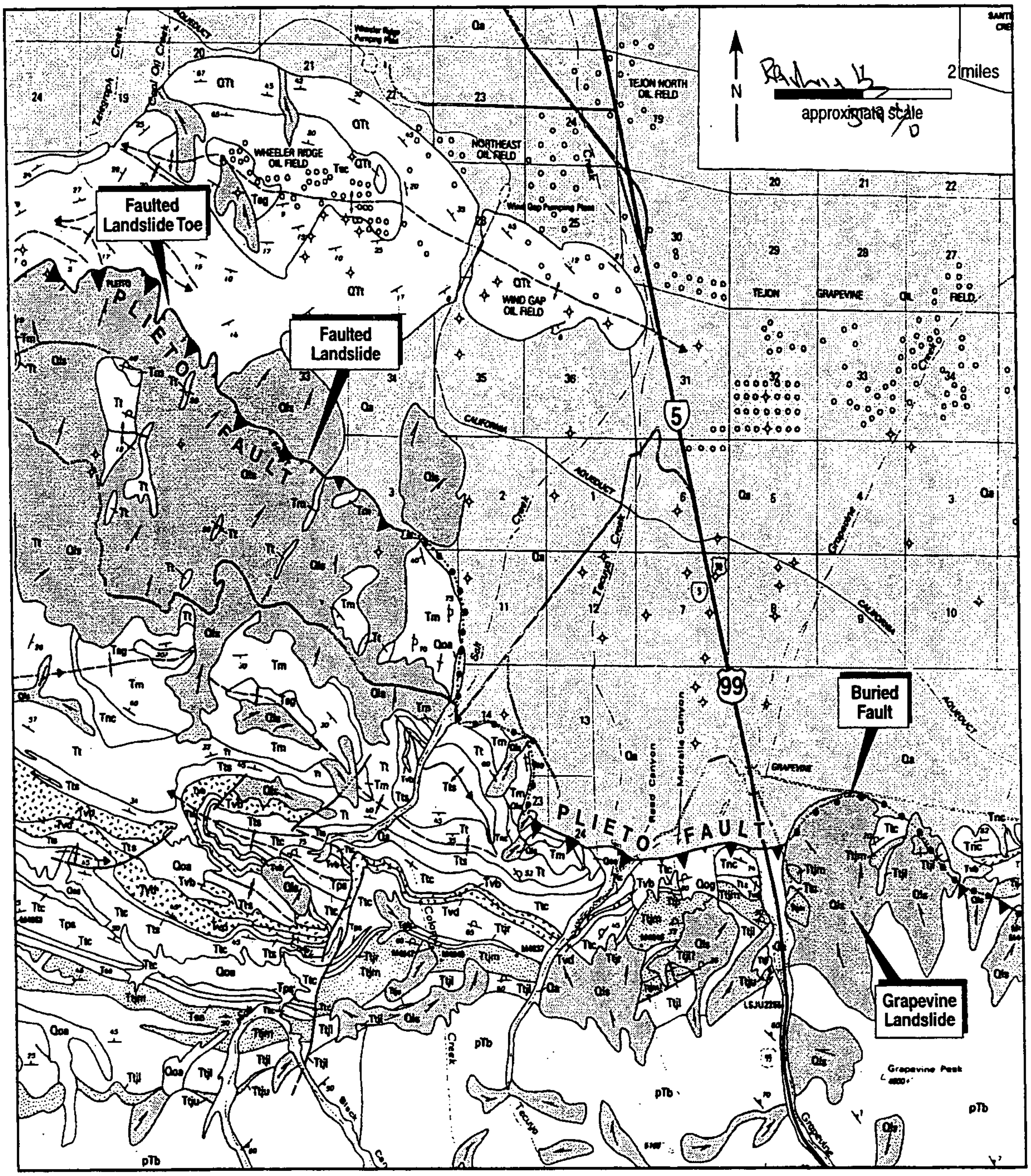

Figure WC.8 Map of the complex relationships between the Plieto thrust fault and massive landslides (pink, Qls) on the north flank of the San Emigdio Mountains, California (from Nilsen, 1987). The Plieto thrust fault is mapped as being locally buried by the Grapevine landslide, cutting Quaternary landslide deposits, or as the toe of a portion of the massive landslide complex.

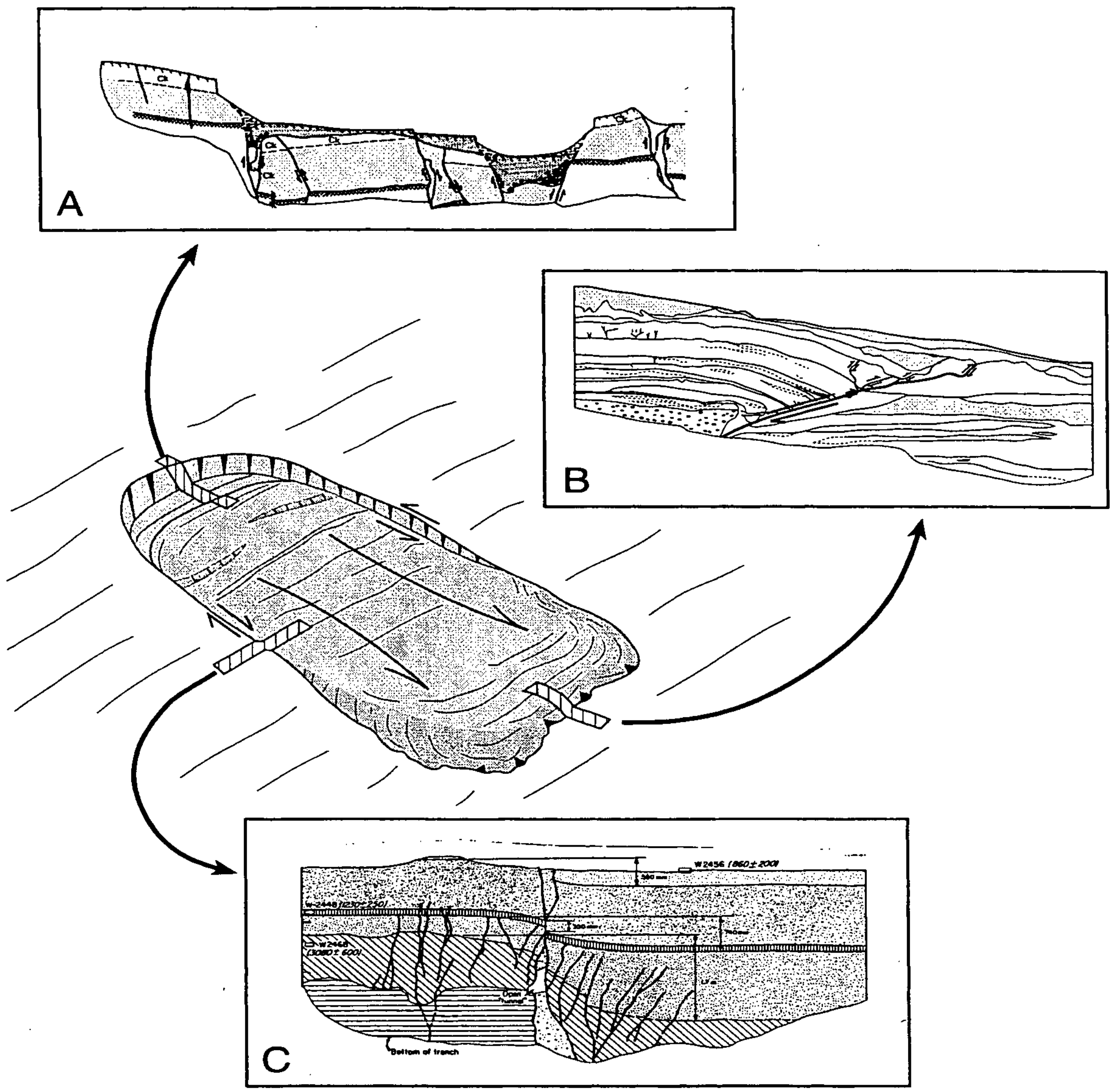

Figure WC.9 Landslide model with trench locations at the head scarp, right flank, and toe, and published paleoseismic trench logs that reflect the type of faulting expected at each locality. (A) Normal fault log. Reprinted with permission from Schwartz and Coppersmith, "Seismic Hazards: New Trends in Analysis Using Geologic Data," in Studies in Geophysics, Active Tectonics. Copyright 1986 by the National Academy of Sciences. Courtesy of the National Academy Press, (B) thrust fault log from Carver 1994, and (C) strike-slip fault log from Clark, Grantz, and Rubin, 1972.

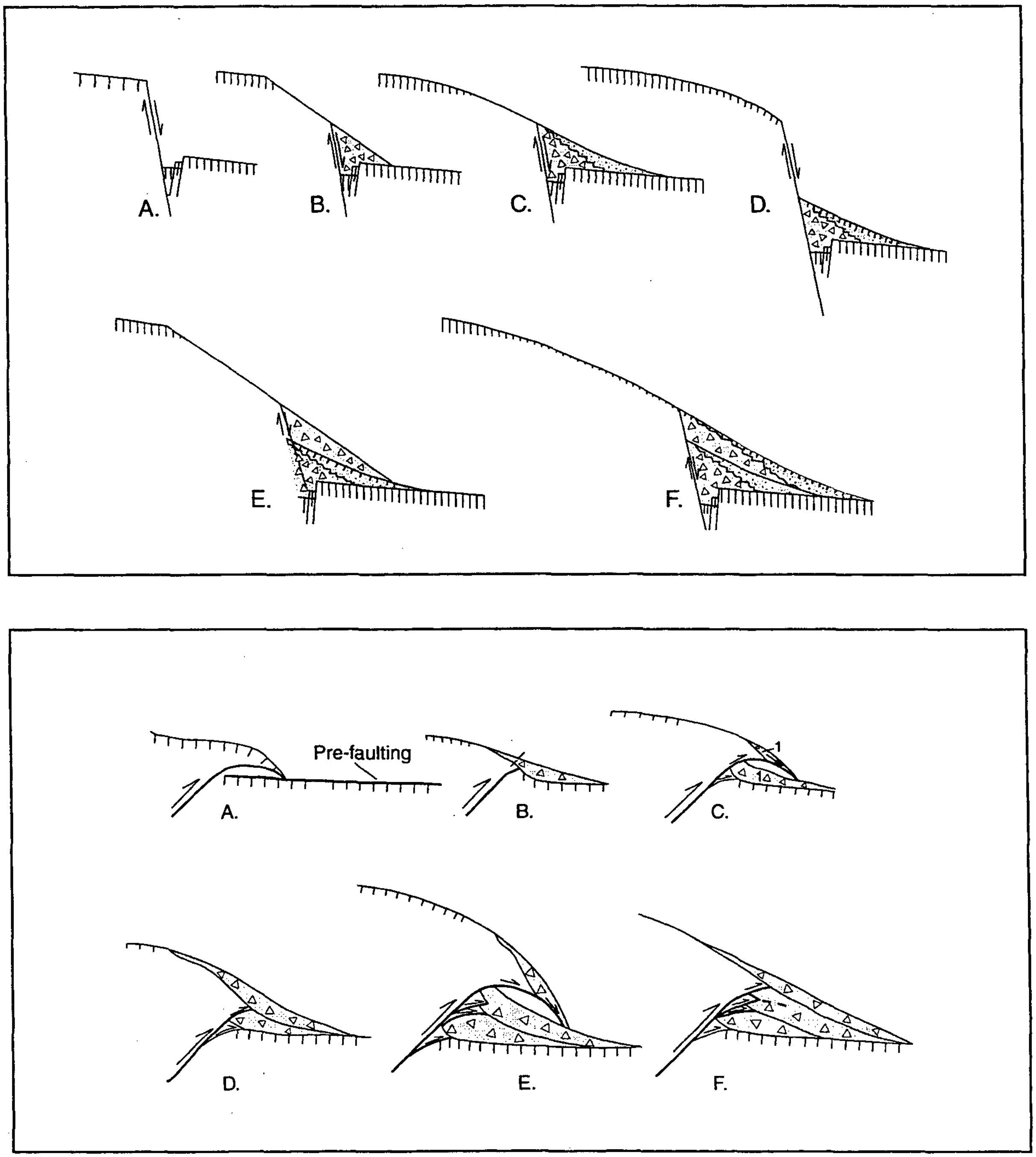

Figure WC.10 Diagrammatic models of the development of colluvial wedges by repeated ground rupturing of a normal fault (top) and thrust fault (bottom) (Modified from McCalpin, 1987). These same records could be developed by normal faulting along the head of a landslide and thrust faulting at the toe.

planar slickensided shear surfaces, striae, clay gouge, and crushed fabric of adjacent earth materials. Careful study of the orientation and juxtaposition of the microstratifications, morphology of the faulted rock and sediments should all be described and compared with trends in the regional tectonic setting. Landsliding is a likely explanation for anomalies identified by such comparisons.

In the final analysis it is impossible to distinguish between tectonic and nontectonic faults as viewed in a trench wall. It is always necessary to step back and gain the benefit of seeing the surrounding geomorphic and geologic setting.

4 SUMMARY

This summary outlines the primary findings and conclusions distilled from this paper:

(1) The tectonic or nontectonic origin of a fault or a series of faults exposed at a field outcrop or a paleoseismic trench cannot be determined in the absence of a three-dimensional assessment of the regional geomorphic, geologic and seismic setting

(2) Both gravity sliding and tectonic forces develop the same set of stresses in brittle earth materials that, in turn, produce faults of normal-, reverse- and strike-slip displacements, and their associated geomorphic features. The origin of all faults should be suspect in the early stages of any paleoseismic investigation

(3) The world-wide geologic record contains excellent examples of faulted rock materials which are the product of large gravity sliding events along detachment surfaces that are regional in extent and have subsequently undergone deformation and deep erosion. Only with extensive field studies can these large fault surfaces of nontectonic origin be accurately recognized

(4) Identifying a fault as seismogenic requires that paleoseismologists follow well-established procedures that are presently in use. The possibility that a paleoseismic study site was affected by gravity-driven processes should force investigators to always consider sliding as a mechanism capable of influencing the microstratigraphic record. This is especially true where, for either technical or socioeconomic reasons, critical facilities are required to be placed in mountainous terrain

(5) Paleoseismologists and tectonic geomorphologists need to improve their understanding of techniques that can be used to recognize large landslide features in the modern landscape. To this end, development of Landslide Inventory Maps is a critical task. It is as important, during the paleoseismic site selection process, as the characterization of the regional geologic and seismic setting

(6) Field research is needed to define better the subsurface microstratigraphic signature of landslide boundary faults. The published literature describing faulted rock and faulted soil geometries and fabrics is essentially nonexistent

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APPENDICES

GLACIO-ISOSTATIC (POSTGLACIAL) FAULTING CRITERIA FOR RECOGNITION

Clark H. Fenton[1]

1 EXECUTIVE SUMMARY

Crustal rebound following the removal of an ice sheet is not always a regionally uniform process. Differences in ice thickness and the response of different rock types to ice loading give rise to differential uplift that is accomplished by reactivation of pre-existing faults. Movement on these reactivated faults is often seismogenic, as shown by the spatial and temporal association with seismically-induced deformation including liquefaction and landsliding. Such faulting is termed glacio-isostatic or postglacial faulting. Postglacial faulting occurs in regions that presently have low rates of seismicity and fault activity. Postglacial faulting is generally reverse or thrust faulting, with displacements ranging from a few millimeters up to several tens of meters on faults that range from a few meters to several hundred kilometers in length. Limited exploratory trenching shows that postglacial fault scarps are the result of single events, at or immediately following the time of deglaciation.

The lateglacial and postglacial environment, however, has a number of other processes that form fault scarp-like features that can commonly be mistaken for postglacial faults. These are glaciotectonism (ice push or ice drag), periglaciation (freeze-thaw), and shallow stress relief (pop-ups).

Distinguishing between postglacial fault scarps and scarps produced by nontectonic processes requires the application of geologic, stratigraphic, geomorphic, structural, and geophysical criteria.

Postglacial faulting is predominantly reverse and shares many geomorphic and structural characteristics with tectonic thrust faulting, including continuity of surface trace, consistency of sense and amount of throw along strike, and an orthogonal orientation to the direction of maximum horizontal compression. Glaciotectonic and periglacial faulting are more random in orientation, sense and amount of throw. Glaciotectonic faults tend to parallel the former ice front. In addition, they also tend to form broader, but more discontinuous zones that result from continuous deformation. Postglacial faulting results from discrete (seismogenic) events. Postglacial faulting, by definition occurs after deglaciation. Glaciotectonic faulting, on the other hand, must occur while ice is still present and actively advancing.

In conclusion, distinguishing between postglacial faulting and nontectonic deformation features in formerly glaciated regions is best accomplished by the application of combined geomorphic, structural, and stratigraphic criteria. Detailed geomorphic mapping, exploratory trenching, accurate age-dating, detailed structural analysis, and mapping of contemporaneous deformation features will all aid in the correct identification of postglacial and nontectonic deformation structures.

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2 INTRODUCTION

Glacio-isostatic faulting, colloquially referred to as 'postglacial faulting', occurs in regions of ice cover in response to changes in the glacial load: either as a result of deglaciation (crustal unloading) or glacial advance (crustal loading). Postglacial faulting has been reported in northwest Europe (Norway, Sweden, Finland, Russia, Eire, and Scotland) and North America (eastern Canada, northeastern U.S., and possibly California and Montana). To date, all examples of postglacial faulting have been recorded in regions of presently low to moderate seismicity, namely passive margin, failed rift, or intraplate/craton environments. With the notable exception of the 1989 M6.1 Ungava surface rupture (Adams *et al.*, 1991), postglacial faults are unique in that they occur in regions where there is no evidence of prior surface faulting during recorded history. In addition, these regions have no historical record of seismicity that approaches the magnitude thresholds for generating surface faulting. To date, all examples of postglacial faulting have involved reactivation of pre-existing faults and fractures.

One problem encountered while trying to identify tectonic faulting in formerly glaciated regions is that many glacial and periglacial processes create geomorphic features that mimic surface fault rupture. These processes include glaciotectonism (ice-push or basal drag), freeze-thaw heave, ice-cave collapse, cryoturbation, and meltwater expulsion (Adams *et al.*, 1993). In order to characterize the seismotectonic behavior of such regions and to identify potentially seismogenic structures, it is important to differentiate between these deformation mechanisms and tectonic surface faulting. Glacio-isostacy is a transient process that endures only as long as the glacial loading/unloading stresses are present (Fenton, 1991). Therefore, it is also desirable to be able to differentiate between faulting that results from transient loading stresses, i.e., postglacial faulting, and tectonic faulting that results from long-term regional lithospheric stresses.

Postglacial faulting was first reported by Mather (1843) at Copake, in New York State, where "*masses of slate had been shifted a few inches in a vertical direction by a slight fault, so that the grooves and scratches on the lower part of the mass were continued quite up to that part that had been elevated; and on the upper mass, the same grooves that had once been continuous, were prolonged in their former direction, with the same breadth and depth. This shift of position, or slight fault must have been subsequent to the period when the scratches were made.*"

This description predated the general acceptance of Agassiz's theory of global glacial episodes, and the "grooves and scratches" described by Mather (1843) are, in fact glacial striations. Several other incidences of postglacial faulting, i.e., faulting that offset glacial deposits or glacial striae, were also reported from the northeastern United States and eastern Canada around the turn of the century (Chalmers, 1897; Hitchcock, 1905; Hobbs, 1907; Matthew, 1894a, 1894b; Miller, 1908; Woodworth, 1905, 1907). Postglacial faulting was not reported in Europe until comparatively recently (Kujansuu, 1964; Tanner, 1930). However, until Lundqvist and Lagerbäck (1976) reported the existence of the 200-km-long Pärvie fault in northern Sweden, these reports languished in relatively obscure and hard-to-get literature sources. Although it was a well-known landmark, the comparative remoteness of the Pärvie fault (Pärvie comes from Lapp and roughly translates as "wave in the ground") prevented its earlier discovery. Subsequent to the work of Lundqvist and Lagerbäck (1976) several other, equally spectacular examples of postglacial faulting were discovered in northern Fennoscandia (e.g., Lagerbäck, 1979, 1988; Olesen 1988; Olesen *et al.*, 1992a, 1992b). The search for a site for a hard rock radioactive waste repository site by SKB AB, the Swedish Nuclear Fuels and Waste Management Company, spurred further work on postglacial faulting in Fennoscandia (e.g., Bäckblom and Stanfors, 1989). Further examples of postglacial faulting have been reported in the Scotland (Fenton, 1991; Holmes, 1984; Ringrose, 1987, 1989a, 1989b; Ringrose *et al.*, 1991; Sissons, 1972; Sissons and Cornish 1982a, 1982b), Ireland (Mohr, 1986), and Switzerland (Jackli, 1965). A number of these claims for postglacial glacio-isostatic faulting have since been shown to be the result of nontectonic processes (Muir Wood, 1993).

Despite the relatively successful search for glacio-isostatic faulting in northwest Europe, to date, relatively few examples have been described from North America (Adams, 1981; Fenton, 1994). Large areas of the formerly glaciated Canadian shield, however, have yet to searched in a systematic manner for evidence of recent faulting. Several notable examples, most prominently, the Aspy fault on Cape Breton Island, Nova Scotia (Grant, 1990), and the Ungava surface fault rupture (Adams *et al.*, 1991), have been described in considerable detail. Recent studies (Fenton, *unpublished*; Ruffman *et al.*, 1996), however, suggest that there may be many undiscovered examples of postglacial faulting in the remoter parts of the Canadian shield. Dyke *et al.*(1991, 1992) reported the spectacular offsets of lateglacial shorelines in the Prince of Wales Island-Somerset Island-Boothia Peninsula region of the Canadian Arctic. Shoreline isobases indicate 60 to 100 m down-to-the-west displacement on an approximately north-striking fault structure. Although the causative fault has not been identified, the shoreline isobase offsets indicate significant crustal dislocation and differential rebound during and after deglaciation

The majority of postglacial faults observed in North America, however, are small-scale (< 1 m) offsets of glacially-striated rock pavements (e.g., Adams, 1981; Fenton, 1994; Oliver *et al.*, 1970).

With the exception of two unconfirmed reports of possible postglacial faulting in the granite batholith of the Sierra Nevada, California (Greene, 1996), and in Archean basement in the area of Clarks Forks valley area of Montana (Hinz *et al.*, 1997), to date, no incidences of postglacial faulting have been reported to the west of the Rocky Mountains.

The purpose of this paper is to summarize the identifying characteristics of postglacial faulting and to provide criteria to enable differentiation between postglacial glacio-isostatic faulting, tectonic faulting *sensu stricto*, and other landforms, created by glacial and periglacial processes, that mimic the geomorphic features associated with tectonic faults. This paper is based on a compilation of postglacial faulting literature and publications detailing deformation processes in formerly glaciated regions. In addition, this is supplemented by the experience of the author and that of his various collaborators (see acknowledgments) gained whilst studying postglacial faults, both in North America and northwest Europe.

3 DESCRIPTION OF GLACIO-ISOSTATIC FAULTING AND THE POSTGLACIAL FAULTING PROCESS

3.1 Terminology

Matthew (1894a) first proposed the term postglacial faulting to describe faulting that offset glacially-scoured surfaces at St. John, New

Brunswick, in eastern Canada. The use of this term has be perpetuated by numerous authors (e.g., Hobbs, 1921) to describe similar fault offsets in the northeastern United States. Fenton (1994), describing intraplate faulting from eastern Canada and northeastern United States, considered the term "postglacial" unsatisfactory, stating that it merely implied a temporal constraint and did not consider the genesis of these faults. Equally unsatisfactory, is that the postglacial period differs markedly in duration in different regions. In this respect, it is preferable to use a term that is reflective of the process that triggers these faults, thus, the terms glacio-isostatic or glacial rebound faulting are considered more suitable. However, since the term postglacial faulting has become established in the published literature as a synonym for glacio-isostatic faulting, it will be used in this paper to denote faulting that results from glacial loading/unloading.

Glaciotectonics, on the other hand, is the process by which glacial movement results in the deformation of underlying substrate, including shallow bedrock as well as unconsolidated deposits. The majority of glaciotectonic deformation is the result of ice-push at the leading edge of advancing ice fronts (Croot, 1988). The dominant style of deformation is compressional, resulting in the formation of folds and thrust faults (e.g., Croot, 1987). However, like tectonic thrust systems, glaciotectonic deformation can also form extensional features as a result of bending moment stresses or stress relaxation (Adams *et al.*, 1993). Another form of deformation resulting from ice movement is basal drag. Underlying substrate or shallow rock can become frozen to the base of the ice mass and dragged along. Like ice-push deformation, this results in compressional deformation at the leading edge, however, the trailing edges of areas that have undergone basal drag commonly display extensional deformation (e.g. Schroeder *et al.*, 1986). Broster and Burke (1990) also used the term 'glacigenic' to describe faults that they considered the result of glacial movement. This term is considered redundant, and the term glaciotectonic will be used to denote any deformation resulting from ice movement.

3.2 Characteristics and Geometry

Postglacial faults and similar-looking landforms have been described almost exclusively in terms of their geomorphic (surficial) expression. This level of study, has unfortunately led to a less than complete understanding of the processes involved. In addition, many landforms created by nontectonic processes (glaciation, periglaciation, and landsliding) have also been mistakenly attributed to postglacial fault activity (e.g., SKB, 1990; Muir Wood, 1993).

Postglacial faults are best described by a number of criteria, from regional seismotectonic setting, through local geologic setting, and finally detailed geometric and geomorphic features (Table CF.1). These criteria are discussed in the following sections.

3.2.1 Regional Context

Postglacial faults, by definition, occur in formerly glaciated regions. The nature of the last glaciation determines the physiography of the region; areas subject to ice-cap or continental-scale glaciation (e.g., Fenoscandia and Canada) invariably comprise broad regions of predominantly subdued topography, whereas regions subject to valley or Alpine glaciation, (e.g., UK) are more dissected, leading to high relief (1000 m+), with oversteepened slopes. Common to both areas is the presence of glacial deposits (tills, eskers, drumlins, etc.) and glacially polished/scoured bedrock (striated pavements, rôche moutonées, flutes, etc.). In some instances, if ice retreat was lateral, rather than purely by downwasting (the latter, however, is more common in areas of ice-cap glaciation), many of these features are further modified by glaciofluvial processes or can be entirely buried by outwash deposits. Thus, the topographic/physiographic characteristics of regions containing postglacial faulting vary from broad open regions with subdued relief to areas of extremely steep, dissected 'Alpine' topography.

The regions where postglacial faulting has been observed are exclusively intraplate craton regions (e.g., eastern Canada, Sweden), passive margin settings (e.g., Scotland, Norway), and failed rifts (e.g., the St. Lawrence rift zone, eastern Canada). Although some of the intraplate regions may have been subject to tectonic activity as recently as the early Cenozoic, as shown by igneous activity (e.g., Muir Wood, 1989a) and sedimentation in the basins formed in the arms of failed rifts. They are essentially regions of contemporary tectonic quiescence.

The geophysical characteristics of both passive margins and cratonic settings for postglacial faulting are similar. Each is a region of presently low to moderate heat flow, with low to moderate, diffuse seismicity. Craton areas are generally regions of over-thickened silicic crust. Heat flow at passive margins may be marginally higher than cratons (57 mWm^{-2} as opposed to 38 mWm^{-2}, Bott, 1982) on account of relatively recent rifting and magmatic activity (e.g., the last rifting/magmatic episode in the UK was during the Paleogene-Eocene, with some minor activity continuing into the Oligocene, Muir-Wood, 1989a). The most visible characteristic of regions undergoing glacio-isostatic rebound is the presence of both uplifted (more common) and submerged marine shorelines (Andrews, 1991). Measurements of raised shorelines have traditionally been used to infer uniform or concomitant uplift of broad crustal regions, with minor perturbations in the elevation data being assumed to be the result of measurement inaccuracies. More recently, these discontinuities in elevation data have become understood as indications of non-uniform behavior of the crust during glacio-isostatic rebound (e.g., Sissons, 1972; Sissions and Cornish, 1982a, 1982b). In addition, where there is good age-control on these shorelines and when there is well constrained elevation data, differential depression of shoreline elevations can be noted during ice-advance stages (e.g., Firth, 1986; Koteff *et al.*, 1993). The well constrained age of uplifted shorelines around the Gulf of Bothnia, Sweden, and the west coast of Scotland in particular, provide ideal time markers with which to date the movement on postglacial faults. The Aspy fault on Cape Breton Island, Nova Scotia is observed to displace the 125 ka late Sangamon rock-cut platform by 15 m (Grant, 1990).

Therefore, at a regional scale, postglacial faults occur in formerly glaciated (recently deglaciated) craton, failed rift, and passive margins settings that are undergoing or have recently undergone glacio-isostatic rebound. These regions are characterized by relative tectonic quiescence with diffuse, low to moderate levels of seismicity, and low to moderate heat flow. Topography is generally subdued in regions of former ice-cap glaciation or can be extremely dissected in areas of valley or alpine glaciation.

3.2.2 Local Geologic Context

Postglacial faults occur in areas of relative tectonic quiescence. This includes highly deformed cratons, Phanerozoic (Paleozoic) fold belts and Mesozoic-Cenozoic rifted margins. On a local scale, with the exception of minor (millimeter scale) faults in eastern Canada expressed in fissile slates and phyllites (e.g., Adams 1981, 1989; Fenton, 1994; Grant, 1980), there appears to be no lithologic control on faulting. Structural control of faulting, however, is important. To date, all examples of postglacial faulting involve rupture of pre-existing faults (e.g., Eliasson *et al.,* 1991). Postglacial faulting activity does not seem to be able to create new faults in the presence of numerous pre-existing fractures in extensively deformed crust. (Note, no postglacial faults have been reported from platform regions (c.f., Hancock *et al.*, 1984) that have undergone relatively little deformation). In addition, at a local scale postglacial faults do not appear to be constrained to any particular geophysical characteristics, merely that they tend to follow pre-existing structure, namely faults and fractures (and even shear zones) (Fenton, 1991).

At a local scale, as a result of the diffuse nature of seismicity, there appears to be little spatial association contemporary seismicity and postglacial faults. A recent study by Arvidsson (1996), however, indicates that microseismic activity in the Lansjärv region of northern Sweden may be correlated with the subsurface projection of the Lansjärv fault. It should be noted that the Lansjärv fault was chosen for special study by SKB AB as part of the site selection process for a hardrock site radwaste repository, therefore, the subsurface extent of the fault is well constrained by detailed seismic data.

The seismogenic nature of the postglacial faulting process itself, is highlighted by the spatial association of the fault scarps with seismically-induced liquefaction features and landsliding (e.g., Lagerbäck, 1990; Ringrose, 1989b).

At a local scale there appears to be no consistent association of postglacial faults with hydrologic conditions. Expulsion of groundwater was noted along the scarp of the Lainio fault during a field visit in 1991. Also, Lagerbäck (1988), noted groundwater expulsion features in trenches excavated along the Lansjärv fault Elevated fracture water pressure may be important in the triggering of postglacial faults (Fenton, 1991, 1992; Muir Wood, 1993; see Section 3.3). Other than liquefaction, no other hydrologic features have been noted in association with postglacial faulting.

3.2.3 Physical Characteristics of Postglacial Faults

From the study of postglacial faulting around the world, a number of characteristics appear to be common to all postglacial fault scarps. In general postglacial faults are reverse faults that are reactivated pre-existing (faults, fractures or shear zones) crustal weaknesses. The scarps themselves are commonly in bedrock. In these instances the faults displace glaciated rock surfaces (striated or polished pavements or landforms). Faults within glacial or postglacial materials are often difficult to identify due to the mechanical instability of these materials in the postglacial environment (Lagerbäck, 1990). The geometry of postglacial fault scarps are roughly continuous, linear to slightly arcuate reverse fault scarps. The geometry, in particular the along-strike orientation of the faults, is a direct consequence of pre-existing bedrock structure. Larger faults, for example the Pärvie and Lansjärv faults in northern Sweden do, in some instances, break along new ruptures over short sections in order to join a number of pre-existing faults. Equally, some postglacial faults may be sinuous where dictated by pre-existing structures.

The scale of postglacial fault scarps range from features that can only be traced for a few millimeters to the 200-km-long Pärvie fault. Shorter fault lengths may be a consequence of relatively poorly exposed faults in areas with considerable glacial deposits. Smaller faults generally occur in large groups in areas of slatey or phylitic rocks, where up to several hundred faults, each with a throw of only a few millimeters, may occur over an outcrop width of a few 10s of meters (Adams, 1981; Fenton, 1994; Oliver *et al.,* 1970). Larger faults are almost exclusively single scarps with few or no branch faults, splays or secondary deformation. A notable exception is the 'big bend' region of the Pärvie fault where a shallow thrust ramp is associated with secondary normal faulting that accommodates bending moment stresses (Muir-Wood, 1989b, 1993). A similar low-angle thrust 'flake' is observed along the Lansjärv fault and does not appear to be associated with any pre-existing brittle structure (Lagerbäck, 1988). The aspect ratio (scarp height to fault scarp length) for postglacial faults is often less than 1:11,000, i.e., less than that for tectonic reverse faults (Scholz, 1990). The scarp heights range from a few millimeters to possibly several 10s of meters. As stated above, the sense of deformation is almost exclusively reverse faulting, however, small components of strike-slip faulting and normal faulting have been reported. The latter almost exclusively accommodate geometric complexities, namely bending moment stresses, in the fault trajectories.

Where paleoseismic data is available, the offsets on postglacial faults are observed to be the result of single events (Lagerbäck, 1988, 1990; Fenton, 1991, *unpublished*). Trenching studies in Sweden, Scotland, and Canada have so far not shown any conclusive evidence for repeated movement on these structures following deglaciation. In some cases, this means that faulting events have involved up to 15 m of vertical displacement. The Lainio fault in northern Sweden has a 30-m-high scarp. Degradation of this bedrock scarp, however, has not allowed any assessment of the number of faulting events that resulted in its formation.

The crustal extent of most postglacial faults is unknown, however, it is obvious that faults with along-strike lengths of a few millimeters and throws of several millimeters are not going to extend to seismogenic depths. Conversely, the large faults in Fennoscandia, exemplified by the Pärvie scarp, with along-strike lengths of 100s of kilometers and offsets of up to 15 m per-event, will extend to seismogenic depths and have been inferred to continue through the entire (40 km) thickness of the brittle lithosphere (Arvidsson, 1996; Muir Wood, 1989b). These fault dimensions (40 x 200 km) have been used to infer paleoevents as large as M_w 8.5 for the Fennoscandian faults (Muir Wood, 1989b, 1993).

A final local characteristic of postglacial faults is their relationship to the contemporary stress field. Although many minor or small faults show no apparent relationship to the stress field (e.g., Fenton, 1991), larger faults are almost always oriented perpendicular to the direction of maximum horizontal compression (Muir Wood, 1993). Also related to the stress field or rather the strain rate (the release of stress) is the spacing between postglacial faults. In Fennoscandia and western Scotland, two areas where postglacial faults have been studied in considerable detail over a broad region, there appears a self-similar or fractal relationship between fault size (length) and interfault spacing. In Fennoscandia, where the faults are several tens to hundreds of kilometers long, the spacing between faults is of the order of about 100 km. In Scotland, where the mapped faults range in length from a few km to several tens of km, the spacing between faults is about 10 km. The offset along these faults scales in a similar manner, with the average being 10 m and 1 m for Fennoscandia and Scotland, respectively. The reasons for this behavior are not clear. The main difference between Scotland and Fennoscandia is that the former was subject to valley glaciation with an ice thickness of ~ 1 km, while Fennoscandia was covered by a 3-km-thick ice-cap during the last glaciation.

3.3 Glacial Loading and the Generation of Glacio-Isostatic Rebound Stresses

Most studies concerning the effects of ice loading consider the lithosphere as a whole and are primarily investigations of the physical properties of the mantle (e.g. Peltier and Andrews, 1976), therefore they do not give much insight into the generation of stress in the upper brittle regime of the lithosphere. Regardless, such approaches give a regional perspective of the behavior of the crust in response to external loads. In such studies, the lithosphere is modeled as a thin (visco) elastic sheet of infinite length subject to an external load. This results in flexure of the sheet giving rise to bending or "fiber stresses". This effect is greatest for loads of a width that is greater than the thickness of the lithosphere. Such "fiber stresses" reach a maximum for loads c. 4.4 times the flexural parameter of the lithosphere (commonly 55-100 km) i.e. load widths of c. 500 km. The maximum horizontal stresses created are c. 6 times the load pressure. Such bending stresses will become negligible compared to the load pressure when then the load width is >50 or <0.5 times the lithospheric parameter (Bott, 1982).

On a regional scale ice-cap loading induces bending stresses of the order of 100 bar (10 MPa) (Quinlan, 1984; Stein *et al.*, 1989). Peltier & Andrews (1976) show that ice-cap melting involved the redistribution of mass in the planetary interior. This is accomplished partly elastically, i.e. there is an instantaneous response to changes in the ice load, and partly in an anelastic manner, occurring by mass transfer of sub-lithospheric plastic material. Such flow gradients acting on the base of the brittle part of the lithosphere will allow the transfer of stress into the upper brittle part of the lithosphere giving rise to longer lived glacially induced stresses than those associated with the elastic response of the crust.

Stein *et al.* (1989), in an attempt to resolve the cause of seismicity at 'passive margins', modeled the stresses due to glacial loading as flexure of an elastic plate. The stresses created are seen to be trivial in comparison to those created by sediment loading. Evidence from earthquake source parameters does not point to the largest earthquakes being correlated with the areas of greatest sediment accumulation. However the limited data base on passive margin earthquakes does point to the largest earthquakes being in the areas that have been formerly glaciated. Thus, glacial loading stresses, although seemingly small in comparison, appear in some way to control the locus of large passive margin seismic events. Quinlan (1984) emphasizes that glacially-induced stresses, although being capable of triggering earthquake faulting, are not of sufficient magnitude to dictate the mode of failure, and therefore need to be considered in conjunction with the (usually poorly understood) ambient stress field. The behavior of the crust responds to the whole stress system that is active at any particular time, and is not controlled by any single mechanism, unless it can be shown that that particular mechanism dominates. The deglaciation model of Stein *et al.* (1989) predicts several hundred meters of rebound in response to the removal of 1 km thickness of ice. This agrees with the shoreline evidence from Fennoscandia (Mörner, 1981). However, the model fails to heed the advice of Quinlan (1984) in that it considers only the stresses due to the removal of the ice-load giving rise to a situation where the area of former ice cover is in extension while the area immediately outside this is in compression. From the styles of post-glacial faulting observed in Fennoscandia and Scotland this is clearly not the case. The model fails to account for the effects of the regional tectonic stress field. Adams (1989), describing the causes of post-glacial faulting in Canada, does consider both the regional stress field along with that created by glacial unloading and shows that the resultant stress field has the ability to trigger reverse faulting in the upper c. 500 m of the brittle crust. He states that the glacially-induced flexural stresses (of c. 20 MPa) may dominate the contemporary horizontal stress in the upper 300-1000 m of the brittle crust, and therefore, may even dominate the regional stresses in this part of the crust. The brittle, fractured nature of the upper 500 m of the crust suggests that such stress is at least partly relieved and not stored elastically as would be the case at depth. Deeper, stored stress may be released suddenly, during or following deglaciation, by seismic activity. The small throw of the faults and parallelism of the fault orientations to the ice margins in eastern Canada point to fault movement being in response to transient stresses in the fractured upper 500 m. A much different response to deglaciation than that noted in Scotland and Fennoscandia where movement on individual faults is of the order of meters and even tens of meters. Bostrom (1984) considered the effects of crustal extension during the residence of an ice load. This was accomplished by the flow of asthenospheric material out of the center of loading giving rise to tensional stresses. Deglaciation causes movement in the opposite direction and creates compressional stresses. Such asthenospheric flow is deemed responsible for the dissipation of isostatic stresses, prior to the yield strength of the quasi-plastic crust being reached, and as such may account for the lack of the expected spectacular post-glacial faulting in eastern Canada. However, recent work in eastern Canada suggests that the comparative lack of postglacial faulting may be due to the lack of an active search to find such features (Grant, 1990; D.R. Grant, Geological Survey of Canada, *pers. comm.*, 1991).

Modeling crustal stresses under conditions of glacial unloading (Appendix A; Fenton, 1991, 1992; Muir Wood, 1993) shows the importance of the creation of a regime of crustal fluid overpressuring to account for the observed fault movements. The model shows that, in the immediate post-glacial period, the sudden release of the vertical stress due to the ice load creates a situation of 'chaotic' fault activity, with the orientation of fault reactivation being only loosely controlled by the regional stress field. As the crustal stress levels are reduced, by a combination of seismic stress drop and the decreasing influence of fluid overpressuring, a second phase of fault movement, utilizing only the most favorably orientated faults, is initiated. With horizontal stress levels enhanced by stress transfer from the flow of sub-crustal material back into the former area of ice loading and localized preservation of high levels of vertical stress, this period is characterized by reverse fault movement.

4 CRITERIA FOR THE RECOGNITION OF POSTGLACIAL FAULTS

Potential methods for differentiating between postglacial faults, tectonic faults and other landforms/structures can be divided into geological, geomorphic, structural, and associated criteria. Mohr (1986) first proposed a list of criteria to differentiate postglacial fault scarps from glacially-plucked features in western Ireland. These criteria were modified and added to by Fenton (1991, 1994) using additional data from Scotland, Sweden, and Canada. These criteria are:

(1) Faults should have demonstrable movement since the disappearance

of the last ice sheet within the area of concern.

(2) The fault should offset glacial and lateglacial deposits, glacial surfaces or other glacial geomorphic features. Preferably, it should be demonstrated that the fault displaces immediately postglacial stratigraphy and/or geomorphic features, though it need not cut younger features.

(3) Fault scarp faces and rupture planes expressed in bedrock should show no signs of glacial modification, such as striations or ice-plucking. Limited glacial modification, however, may be present on scarps that are lateglacial or interglacial in age.

(4) Surface ruptures must be continuous over a distance of at least 1 km, with consistent slip and a displacement/length ration (D/L) of less than 0.001.

(5) Scarps in superficial material must be shown to be the result of faulting and not due to the effects of differential compaction, collapse due to ice melt, or deposition over pre-existing scarps.

(6) Care must be taken with bedrock scarps controlled by banding, bedding, or schistosity to show that they are not the result of differential erosion, ice plucking, or meltwater erosion.

(7) In areas of moderate to high relief, the possibility of scarps being the result of having been created by deep-seated slumping driven by gravitational instability must be disproved.

With the exception of the fourth criterion, Fenton (1994) applied these criteria in ranking claims for postglacial faulting in eastern North America. The criterion of a rupture length of greater than 1 km was given less weight in discerning postglacial faulting origins in eastern Canada and the northeastern U.S. This criterion was originally proposed for the study of faulting in Scotland and Fennoscandia, where postglacial faults are found almost exclusively in crystalline basement rock. The mechanical behavior of slates and phyllites, within which many smaller postglacial faults are found in eastern North America, may not promote the formation of a large, through-going rupture. Indeed, some sites with multiple small displacements at the surface may be represented by a single, larger fault at some depth. On the other hand, such small displacements may be the manifestation of the pervasive release of shallow rebound stresses.

Muir Wood (1993) put forward a list of "neotectonic diagnostics" to grade claims for neotectonic activity (not merely postglacial fault activity) in formerly glaciated regions. These criteria were:

(a) The surface or material that appears to be offset has to have originally formed as a continuous, unbroken unit. Can the surface be dated? Is it the same age?

(b) Can the apparent evidence of an offset be shown to be related directly to a fault?

(c) Is the ratio of displacement to overall length of the feature less that 1/1,000? For most faults this ratio, a function of the strength of the rock prior to fault rupture, is between 1/10,000 and 1/100,000 (Scholz, 1990).

(d) Is the displacement reasonably consistent along the length of the feature?

(e) Can the movement be shown to be synchronous along its entire length?

Similar criteria have been adopted by the Norwegian Geological Survey (O. Olesen, Norwegian Geological Survey, *written communication,* 1996). From the authors experience in eastern Canada, the criteria of Muir Wood (1993) proved particularly useful for differentiating between glaciotectonic deformation and postglacial faulting.

As stated previously, the main aim of this paper is to develop criteria for recognizing postglacial faulting. If we look at the physical characteristics of postglacial faulting, tectonic faulting and glaciotectonic deformation, we see that no single criteria is unique to any type of faulting:

	Postglacial Faults	**(Reverse) Tectonic Faults**	**Glaciotectonic Deformation**
Length	10 m to 100's km	> 10 km	m to km (< 3 km)
Continuity	Generally Continuous	Continuous to Discontinuous	Discontinuous
No. Scarps	Single	Single to Multiple	Generally Multiple
Sense/Style	Predominantly Reverse	All	Reverse (& Normal)
Plan	Linear, Angular	Linear, Arcuate	Irregular
Scarp Height	mm to 10s m	Up to km	Up to several m
Displacement History	Single Event	Repeated	Continuous
Secondary Deformation	Minor Faulting	Faulting and Folding	Faulting and Folding
Relationship to Ice Cover	Within Area of Former Ice Cover	No Relation	Margins of Former Ice Cover
Timing	Postglacial	No Constraint	Synglacial

Faults can be described in terms of their geologic (stratigraphic), geomorphic, and structural expression. In addition, their spatial association with secondary, or off-fault, deformation, such as liquefaction and landsliding, can provide clues as to whether the deformation is seismogenic. The usefulness of these characteristics in providing criteria with which to determine the origin of faulting will be discussed in the following sections.

4.1 Geologic and Stratigraphic Criteria

By definition postglacial faults occur in recently deglaciated regions. The areas that had significant late Quaternary ice cover are predominantly intraplate/craton, passive margin, and failed rift environments. Although postglacial faults do not appear to be constrained by rock type, they are controlled by pre-existing structure, nearly always following existing faults, shear zones, or fractures. Determining a postglacial faulting genesis for a fault structure has more than just academic interest. Since postglacial faulting appears to occur almost exclusively in the immediate postglacial period, while the transient glacial unloading stresses are sufficient to trigger faulting, it is important to differentiate between 'one-off' postglacial faults and tectonic faults with long recurrence intervals in order to evaluate seismic hazards.

In order to assess whether postglacial faulting is seismogenic or not, we have to assess whether movement has been episodic or continuous. Stratigraphic evidence for slow, continuous deformation will preclude seismogenic surface-rupturing, even if continuous movement is the result of tectonic creep. Episodic movement, could also be the result of non-seismic processes. The following criteria are intended to address the questions of episodicity of displacement and seismogenic versus non-seismic processes.

CRITERION: Evidence for Continuous Deformation v. Episodic Deformation

In order to assess whether scarps in formerly glaciated environments are tectonic or nontectonic, we have to discern whether they are the result of continuous deformation or by discrete and/or episodic events. Structural or stratigraphic evidence for continuous deformation would indicate that the scarp would not have formed coseismically. However, 'event scarps' could also have been produced by nontectonic, hence nonseismic, mechanisms. The following criterion are discussed in an attempt to differentiate between postglacial faults and glaciotectonic or other nontectonic processes.

To date, there have been very few trench excavations across postglacial faults. The most notable exception is the Lansjärv fault in northern Sweden (Lagerbäck, 1988). Each trench exposure indicates that there has been only one faulting event during the postglacial period. Like tectonic faults, the fault movement history is determined by stratigraphic offsets, upward terminations of fault strands, and colluvial wedge stratigraphy. It is clearly observed that each trench exposure contains only one colluvial wedge and that the trench stratigraphy shows uniform offset, regardless of age. Were these scarps to have formed by a continuous process, whether it be tectonic creep or some nontectonic mechanism, we would observe continuous onlap within the trench stratigraphy, with increasing offset with age within the faulted units. Thus, for the few Fennoscandian faults, and one example each from Scotland and eastern Canada, it appears that the development of colluvial wedge stratigraphy indicates that these scarps were produced by discrete, one-off events. Glaciotectonic features, in contrast, are the result of continuous deformation and, although they may result from pulses or surges of glacial movement, they do not produce the stratigraphic and structural relationships that are indicative of sudden coseismic offsets.

The deformation observed in trenches across postglacial faulting is entirely steeply dipping, reverse faulting. Glaciotectonic faulting, on the other hand, is much more variable, showing a wide range of dips from near vertical to subhorizontal, often along the one fault plane. Many examples of glaciotectonism show shallow decollement, often at stratigraphic contacts (e.g. till-bedrock contact). In addition, glaciotectonism often displays both compressional and extensional deformation within the same outcrop (Adams *et al.*, 1993). Compressional glaciotectonic deformation often involves the formation of folds. Folding has not been reported to be associated with reverse postglacial faulting, even where expressed in unconsolidated sediment.

Thus, stratigraphic relationships that show discrete, one-off displacements can be used to distinguish postglacial faulting from glaciotectonic deformation. Folding and combined extensional and compressive deformation on the other hand are more typical of glaciotectonic deformation.

CRITERION: Timing of Faulting
By definition, postglacial glacio-isostatic faulting occurs following the disappearance of ice cover within the region of concern. Glaciotectonic deformation, however, can only occur while ice is present and actively advancing or retreating. Establishing the timing of faulting, therefore, should prove useful in distinguishing between postglacial and glaciotectonic faulting (Lagerbäck, 1992).

4.2 Geomorphic Criteria

Tectonic faulting and glaciotectonic deformation can both produce geomorphic scarps. In areas of high relief, gravitational slope movements can also produces scarps that can be mistaken for tectonic faulting. Glacial action, in particular plucking at the base of the ice-sheet/glacier along pre-existing faults, fractures, or bedding/schistosity planes, can result in bedrock steps that may be misidentified as fault scarps. The following criteria are an attempt to differentiate between these differing scarp-forming mechanisms.

CRITERION: General Geomorphology
If we ignore the small displacements of glaciated pavements by movement along steeply-dipping cleavage planes in slatey horizons in northeastern North America, postglacial faults are generally a kilometer or more in length, with a roughly continuous surface trace. Postglacial faults expressed entirely in bedrock show displacement of glacial geomorphic features such as flutes and striations (see description of Mather, 1843). By definition, postglacial faults scarps should show no evidence of glacial modification such as moulding or plucking of the scarp face. Although a number of postglacial fault scarps may have suffered periglacial degradation, such as freeze-thaw frost heave (e.g., the Lainio fault scarp in northern Sweden), they show no evidence for glacial modification.

By their morphology alone, the large postglacial faults of northern Sweden can be clearly identified as the surface rupture of major faults (Muir Wood, 1993). Postglacial fault ruptures resemble tectonic surface faults, in particular in their along strike continuity and consistency of sense and amount of throw. Many postglacial faults that are expressed in bedrock rather than unconsolidated materials display "elementary textbook" thrust fault geometry, having very dramatic, steeply-dipping planar fault planes, with little or no modification of the fault scarp. Postglacial fault zones appear to relatively simple, often comprising a single fault plane, with no other evidence of surface deformation. Overall the fault trace is often planar, with only minor local geometric complexities. These local geometric complexities often arise from postglacial faulting utilizing pre-existing fault and fracture sets.

In contrast, glaciotectonic deformation is often localized and irregular, representing the position of the icefront during periods of glacial advance. There is no along strike continuity in either orientation, sense of displacement or amount of offset.

In areas of high relief, such as the West Highlands of Scotland and Norway, several large landslide scarps have been misidentified as fault scarps (Fenton, 1991; Muir Wood, 1993). Differentiating between landslide scarps and faults scarps in areas of high relief is a global problem not merely confined to recently deglaciated regions (see McCalpin, this volume, Appendix A). Slope failure scarps often parallel slope contours and show considerable along-strike differences in both amount and sense of throw. Slope failures often create multiple subparallel scarps, whereas postglacial faults are generally simple, single scarps.

Postglacial faults have much simpler morphologies than either glaciotectonic and slope failure scarps. They have consistent throw and sense of throw (reverse) along strike. In addition, their geomorphic expression is continuous along strike. They displace glacial deposits and glacial bedrock geomorphology, and they are not controlled by either slope morphology or the position of the former ice front.

CRITERION: Scarp Aspect Ratio

When we compare the surface traces of postglacial faults with recent continental reverse faults, the larger Fennoscandian postglacial faults have similar scarp height to rupture length ratios to tectonic faults. Shorter faults, however, have much larger scarp heights (displacements) than tectonic faults of comparable length, i.e. "short, fat faults." This may result from incomplete mapping of the surface fault trace, the mechanism of fault rupture differs from 'normal' tectonic faults, possible involving deep crustal rupture in thick, cold cratonic crust, or that the relatively strong crust in these regions requires higher levels of angular strain before the rupture threshold is exceeded. Regardless, for the majority of postglacial faults, it appears that they have higher displacement to length ratios than similar sized tectonic faults. It should also be noted, comparing postglacial surface ruptures with reverse fault ruptures from similar intraplate/craton environments, postglacial faults are up to four times longer than tectonic faults.

4.3 Structural Criteria

Postglacial faults almost all have reverse offsets. In addition, all reported postglacial faults have reactivated pre-existing faults, fractures, or shear zones (e.g., Eliasson *et al.*, 1991). The morphology of postglacial faults is very similar to that of tectonic reverse faults. There are a number of structural characteristics that can be used to differentiate between postglacial faults, nontectonic deformation, and shallow stress-relief features. The latter, which include pop-ups and offset boreholes, have often been misinterpreted as postglacial or tectonic faults (Fenton, 1994).

CRITERION: Dip of Fault Plane and Depth of Deformation

The style of glaciotectonic deformation differs significantly from that of postglacial faulting in that glaciotectonic deformation is a shallow phenomenon (Dredge and Grant, 1987), which although it may involve bedrock deformation, faults tends to shallow with depth, sometimes dying out in a shallow decollement (Adams *et al.*, 1993). Postglacial faulting on the other hand, appears to steepen with depth (Muir Wood, 1993) and, at least for larger faults, involves rupture of the entire brittle crust (Arvidsson, 1996; Muir Wood, 1989b, 1993). Shallowing of the fault planes in the near surface may be a result of increased density of rebound fractures that are oriented subparallel to the ground surface and the relative ease of movement along these fractures in the presence of postglacial crustal fluid overpressuring (Fenton, 1991). Although some larger postglacial faults show sections that have relatively low angle dips (Lagerbäck, 1988; Muir Wood, 1993), overall, the fault planes dip at steep angles. Stress relief pop-ups (e.g., Wallach *et al.*, 1993), when observed in section, die out into shallow decollements (Fenton, 1994; Wilson, 1902).

CRITERION: Orientation With Respect to the Contemporary Tectonic Stress Field
The majority of postglacial faults, being thrust faults, are oriented orthogonal to the direction of maximum horizontal compressive stress. Some smaller postglacial faults may not show any relationship to the tectonic stress field (see Appendix A). Shallow stress relief features such as pop-ups will show the same orientation (Adams, 1989). Glaciotectonic deformation, however, will be oriented parallel to the former ice front, and therefore, is unlikely to show any consistent relationship to the ambient stress field.

CRITERION: Rupture Complexity and Secondary Deformation
Postglacial faulting utilizes pre-existing faults, fractures, and shear zones, many of which could have been inactive for considerable periods of time (Eliasson *et al.*, 1991). Most postglacial faulting involves single fault planes, with relatively simple fault geometries. Larger faults like the Pärvie and Länsjarv, however, do display more complexities, at least on a local scale (Lagerbäck, 1988; Muir Wood, 1993). These complexities are a function of the geometry of pre-existing faults and fractures. Thus, fault bends and changes in orientation tend to be angular in nature. These fault bends result in secondary accommodation features, most commonly in the hangingwall, namely subparallel normal faults above extensional bends, and thrust faults above compressional fault bends. Complexities in the rupture planes, since they are expressed in brittle bedrock are manifest as secondary faults. Movement of a relatively simple fault beneath an area with multiple fault/fracture sets can lead to chaotic looking rupture patterns (Fenton, 1991). Despite this localized heterogeneity, overall, the sense of displacement is consistent along the entire fault length.

Glaciotectonism conversely results in more heterogeneous deformation (e.g. Croot, 1988). The main difference being the lack of consistency in the sense and amount of throw along strike. Even when involving bedrock, glaciotectonic deformation results in broader deformation zones comprising multiple fault planes, with both normal and reverse offsets.

When postglacial faulting involves both bedrock and unconsolidated materials, the zone of deformation remains a discrete fault plane in both materials (e.g., Lagerbäck, 1988). In glaciotectonic deformation, faults often do not propagate up from bedrock into overlying unconsolidated deposits as discrete planes. It is more common for this deformation to be expressed as folding within these deposits (Adams *et al.*, 1993).

4.4 Associated Deformation

Associated deformation features afford criteria that can be used to differentiate between seismogenic and nonseismogenic processes.

CRITERION: Contemporaneous Association with Seismically-Induced Features
Many postglacial faults are confined to bedrock outcrops, therefore it is difficult to determine whether they have formed as a result of discrete movement episodes or are the result of creep. At smaller scales, e.g., the millimeter-scale offsets of glaciated pavements on slate outcrops in eastern North America, the mechanism of formation becomes a point for academic debate, since, by their size alone, these are clearly not potentially seismogenic structures. However, at larger scales, and also if there are numerous small offsets (possibly representing a larger, through-going structure at depth), the size of the scarps and the continuity of these structures suggests that they may be seismogenic. In order to show that these

structures are tectonic (seismogenic) in origin, we need to demonstrate contemporaneous association with off-fault seismogenic deformation. The most common seismically-induced deformation are liquefaction and landsliding. Unfortunately, both landsliding and liquefaction are a common product of climatic and groundwater conditions in the postglacial environment. Landsliding results from gravitational instabilities on oversteepened glaciated slopes and oversaturation of unconsolidated materials as a result of melting permafrost. Liquefaction occurs spontaneously in saturated glacio-fluvial outwash deposits and also occurs as a result of the expulsion of permafrost meltwater. In addition, freeze-thaw can produce involutions in unconsolidated sediments that resemble seismically-induced liquefaction features. It is therefore important to discount non-seismic mechanisms for triggering landsliding and liquefaction (Ringrose, 1987).

In order to show coseismic genesis for liquefaction and landsliding, the criteria of Sims (1975) provides a useful guide, i.e., these deformation features must have a spatial and temporal association with the suspected seismic source. A spatial and temporal association between liquefaction, landsliding and faulting, however, will only show that the faulting is seismogenic, it will not allow us to differentiate between postglacial faulting and tectonic faulting *sensu stricto*. In formerly glaciated regions, however, the uncertainty is usually over whether the surface deformation observed is the result of (seismogenic) faulting or glacial or periglacial processes. Therefore, a spatial and temporal association with landsliding, liquefaction, or any other seismically-induced deformation, may be a useful criterion in identifying tectonic faults *sensu lato* in recently deglaciated regions (Fenton, 1991; Lagerbäck, 1991; Ringrose 1987, 1989b).

4.5 Geophysical Criteria

The geophysical character of a region can be used to confirm that the area of concern is/was capable of having postglacial faulting.

CRITERION: Association with Contemporary Seismicity

A spatial association with contemporary seismicity is a common criteria used to define active faults. Postglacial faults occur in seismically quiet regions, and to date, associations between postglacial faults and seismicity have, at best been equivocal. One recent study, however, indicates that there may be a spatial association between postglacial faulting and contemporary seismicity and the Lansjärv fault in northern Sweden (Arvidsson, 1996). Areas with recognized postglacial faults, however, are almost always in areas where there is insufficient seismograph coverage to accurately locate microseismic activity. Thus, without further data, it appears that association with contemporary seismicity cannot be used as a criterion for recognizing postglacial faulting.

CRITERION: Spatial Association With Areas With High Rates of Glacio-Isostatic Uplift

By definition, postglacial faults occur in regions that have undergone recent deglaciation. It follows, therefore, that postglacial faults will be found in regions that have or are undergoing significant glacio-isostatic rebound. Evidence from uplifted shorelines (e.g., Gray, 1974a, 1974b, 1978), geodetic measurements (Randjärv, 1993; Saari, 1992), and gravity measurements, can all be used to identify areas of glacio-isostatic uplift.

4.6 Summary of Criteria

From the preceding discussions, it is clear that no single criterion enables unambiguous identification of postglacial faulting. Rather, several criteria can be used to distinguish between postglacial faulting and nontectonic deformation. Differentiating between postglacial faulting and tectonic faulting *sensu stricto* proves to be more difficult.

Initially, it must be demonstrated that the observed faulting occurs within the former ice limits of the region in question and that the observed scarp or stratigraphic offset post-dates the disappearance of ice cover. Displacement must be shown to have occurred as a discrete event(s) and be continuous in terms of both amount and sense of throw along the fault length. A spatial and temporal association with liquefaction and/or landsliding may indicate that postglacial faulting is seismogenic.

Required studies to differentiate between postglacial faulting and nontectonic deformation include: detailed geomorphic mapping; exploratory trenching (where possible) and accurate age-dating; detailed structural analysis; and mapping of contemporaneous deformation features.

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Table CF.1 Summary of key characteristics: glacio-isostatic faulting phenomena

Regional context	
Physiographic	Occur in recently deglaciated regions.
Tectonic	Commonly found in intraplate/craton environments and passive margin or failed rift settings undergoing rapid glacio-isostatic uplift. To date, no examples have been reported from active tectonic settings.
Geologic	Most often found in Precambrian shield and early Paleozoic fold belts (intraplate settings). Not confined to any particular rock type.
Geophysical	Regions undergoing glacio-isostatic rebound. Thickened continental crust with low heat flow (cratons) or moderately attenuated crust with low to moderate heat flow (passive margins and failed rifts).
Seismological	Regions of presently low to moderate seismic activity.
Hydrologic	Not restricted to any particular regional hydrologic setting. Subglacial fluid recharge may be important in triggering postglacial faulting.
Local context	
Topographic	Commonly low topographic relief, although examples have been observed in regions of alpine relief.
Geomorphologic	Deformation associated with surface faulting is localized along existing fault or fracture zones. Small-scale offsets may be part of broader zones of deformation.
Stratigraphic	Not restricted to any particular local stratigraphic setting. Small faults in eastern Canada and northeastern U.S., however, are better expressed in fissile slatey rock types with steep to vertical cleavage planes.
Structural	All reported examples of postglacial faults have reactivated pre-existing faults, fractures, and shear zones. Scarps are high-angle, and vary between simple linear traces and complex angular traces. Displacement is predominantly reverse, although strike-slip components of displacement have been noted along lateral ramps and minor accommodation splays. Minor normal faulting has been reported from the hangingwall blocks of larger faults in Fennoscandia.
Geophysical	Associated with regions of high bending strain as a result of glacio-isostatic rebound. Subsurface geometry of these faults are unknown, although sparse seismicity data does suggest that they may become shallower with depth.
Seismological	Spatial and temporal association with widespread liquefaction and low-

angle landsliding indicates that postglacial faulting is associated with large magnitude seismicity. Contemporary seismicity is low.

Hydrologic
Evidence from trench exposures across the Lansjärv fault in northern Sweden indicate expulsion of groundwater during faulting. Models for triggering postglacial faulting indicate that crustal fluid overpressuring may be required to initiate faulting.

Feature characteristics

Spatial characteristics

Morphology
Steeply-dipping, reverse fault scarps, generally in bedrock or shallow glacial and postglacial deposits. Small normal faults sometimes present in the hangingwalls of larger faults. Small strike-slip components observed on small accommodation structures. Scarps range between a few millimeters to 15 m high. Fault lengths range from a few tens of meters, up to 200 km.

Geometry
Variable. Continuous, roughly linear to slightly arcuate reverse fault scarps to highly angular, irregular scarps utilizing pre-existing fracture sets. Aspect ratio (H:V) < 1:11,000, generally less than similar tectonic reverse faults.

Scale
Fault lengths vary from 1 km to 100+ km. Smaller faults (<10 m length in slatey units in eastern North America. Scarp heights (offsets) vary between 1 mm and 15 m.

Sense of deformation
Reverse faulting, with rare secondary normal faulting. Small components of strike-slip faulting have also been reported on small accommodation splays.

Depth
Larger faults extend through the entire seismogenic crust (up to 40 km in Fennoscandia). Smaller faults (in slatey units) may extend no more than a few meters.

Relation to stress field
Most faults are oriented orthogonal to the direction of maximum horizontal compression. Some faults show no relation to the regional stress field.

Spatial associations
Faulting confined within regions of former ice cover. Postglacial faulting also occurs along pre-existing faults, fractures, or shear zones.

Hydrologic
Occasional association with groundwater expulsion.

Temporal characteristics

Rate of deformation
High rates of deformation, followed by lengthy periods of inactivity.

Episodicity	Single event.
Duration of deformation	Short term around the time of deglaciation.
Recurrence deformation	Single event. No conclusive evidence for multiple events.
Temporal associations	Triggered by deglaciation and consequent glacio-isostatic rebound.
Investigative techniques	Geologic and geomorphic mapping; exploratory trenching; age-dating; structural analysis; drilling.
References	Adams (1981, 1989); Bäckblom and Stanfors (1989); Fenton (1991, 1994); Grant (1990); Lagerbäck (1988, 1992); Lundqvist and Lagerbäck (1976); Mather (1843); Mörner (1978, 1981); Muir Wood (1989b, 1993); Olesen (1988); Oliver *et al.*(1970); Ringrose (1987, 1989a, 1989b); Sissons and Cronish (1982a, 1982b); Tanner (1930).

APPENDIX A
THE EFFECTS OF ICE LOADING AND UNLOADING ON CRUSTAL STRESSES

The following discussion is intended to supplement Section 3.3. Further discussion of the generation of ice loading and unloading stresses can be found in Fenton (1991, 1992) and Muir Wood (1993).

A.1 THE EFFECTS OF ICE-CAP LOADING: GENERAL CONSIDERATIONS

Consider a fault of any orientation subject to crustal stresses, such that the principal stresses $\sigma_1 > \sigma_2 > \sigma_3$ are mutually perpendicular. If one stress is assumed to be the vertical overburden pressure and another principal stress is constrained to lie in the plane of the fault the scenario is simplified to a two-dimensional case. These principal stresses can be resolved into in-plane shear stress (τ) and normal stresses (σ_n) acting on the fault plane such that:

$$\sigma_n = \frac{\sigma_1 + \sigma_3}{2} - \frac{\sigma_1 - \sigma_3}{2} . \cos 2\theta \qquad (1)$$

and

$$\tau = \frac{\sigma_1 - \sigma_3}{2} . \sin 2\theta \qquad (2)$$

where θ is the angle between the fault plane and the minimum principal stress.

Since the fault under consideration is of arbitrary orientation equations (1) and (2) can be simplified to:

$$\sigma_n \;\; \alpha \;\; a\sigma_1 + b\sigma_3 \qquad (3)$$

and

$$\tau \;\; \alpha \;\; c(\sigma_1 - \sigma_3) \qquad (4)$$

with a, b and c all constants. An increase in τ will promote failure of the fault, while an increase in σ_n will promote fault stability.

Now consider the simple cases of two crustal stress regimes, one compressional, one tensional, both subject to an increase in vertical stress (S_{VI}) due to ice loading (Figure A.1). It is seen that in crustal compression, an increase in the value of $S_{V'}$, the effective vertical stress, causes a reduction in τ while σ_n is increased, thereby stabilizing the fault. In crustal tension, the additional load causes an increase in the magnitude of σ_1 which in turn causes an increase in τ and also an increase in σ_n, thereby creating an ambiguous situation that may either cause destabilization of the fault, or equally may cause no difference to the deviatoric stress ($S_{D'} = [\sigma_1 - \sigma_3]$), thus maintaining a situation of stability. A similar situation is created in strike-slip environments. It has been shown that the imposition of large ice sheets on continental crust subject to compression suppresses the occurrence of earthquakes (Johnston, 1987, 1989). As well as reducing the deviatoric stress ($S_{D'}$) acting within the crust, the induced load also increases the effective normal stress (σ_n) acting on potentially seismogenic fault planes, moving the faults away from failure. However, this rather simplistic view only holds for a 'dry' crust where there are no effects due to resident crustal fluids and fluids introduced from basal melting of the ice sheet.

A.2 THE EFFECTS OF ICE-CAP LOADING: WET CRUST

Water is known to exist at the base of ice sheets (Zotikov, 1986) and indeed, many of the erosional features attributed to glacial systems are a result of the action of basal fluid. The presence of sub-glacial water requires the temperature of the ice to rise above its pressure melting point. The presence of stagnant hollows beneath a mobile ice-sheet enhances basal ice melt (Muir Wood, 1989b). Such sub-glacial water will penetrate the underlying crust through open fractures, raising the hydrostatic pressure by the full weight of the ice-cap. This in turn, will encourage further basal melting creating a dynamic hydrological regime.

The penetration of sub-glacial fluid into fractured crust will be a time-dependent diffusion process. Analogous studies of reservoir impoundment (Costain *et al.*, 1987) suggest that, in crustal volumes with sufficient fracture permeability, increases in head of water (ice thickness in this case) can be transmitted to depths of 10-20 km. The timing of crustal fluid diffusion beneath a volume of ice is unknown, however, if it follows the behaviour observed from reservoir-induced seismicity, this may occur over a relatively short time period, with crustal fluid pressures equilibrating with respect to the increased head in a matter of a few months to a few years. If fluid flow within the crust is channeled along a number of aligned fractures, this may greatly increase the flow rate (Roeloffs, 1988). Whatever the rate of sub-glacial fluid flow into fractured crust, over the time period of ice-cap residence, crustal fluid pressure will have time to equilibrate at depth with respect to the increased head, and indeed may be subject to subtle changes due to shorter time scale changes as the ice grows and recedes. Whatever the result, basal ice melt and the penetration of such fluid into the underlying fractured crust will have an important control on the generation of the glacial and postglacial stress regime.

The imposition of a significant thickness of ice will having differing effects at different crustal depths due to changes in rheology and depth of penetration of sub-glacial melt water. Near surface, where the crust is essentially brittle, it is expected that there would be little or no horizontal deformation over the period of ice loading. However, there may be vertical movement in the form of differential compressibility of adjacent rock types and the closing of open fractures. The effective vertical stress ($S_{V'}$) acting on the crust would be increased by the full weight of the ice load (S_{VI}) where:

$$S_{VI} = \rho_I g h \qquad (5)$$

with $\rho_I = 0.9$ Mgm^{-3} (the density of ice), $g = 9.81$ ms^{-2} and h = thickness of the ice cover.

In the presence of increased fluid pressure driven by the weight of the ice load, the increase in the effective vertical stress will be counteracted by a corresponding increase in fluid pressure (P_f) where:

$$P_f = \rho_I \qquad (6)$$

The resultant vertical stress acting on the crust due to ice loading will be the product of the vertical lithostatic stress (S_V) and the weight of the ice load (S_{VI}), minus the effects of the increase in pore fluid pressure (P_f) i.e.

$$S_{V'} = S_V + S_{VI} - P_f \qquad (7)$$

Which, from 5 and 6, simplifies to become:

$$S_{V'} = S_V \qquad (8)$$

Thus, there will be no net increase in the effective vertical stress acting on the crust at depths where the crustal fluid pressure has equilibrated with respect to increased hydrostatic head.

The imposed ice load will also cause an increase in the effective horizontal stress ($S_{H'}$) acting on the crust. Dependent on whether lateral constraint or isotropic boundary conditions are assumed, the relative increase in $S_{H'}$ due to the weight of the ice load (S_{HI}) will be:

$$S_{HI} = 0.3 \text{ to } 1.0\ (S_{VI}) \qquad (9)$$

However, since the time scale of ice loading is relatively short (in a geotectonic sense) and little or no horizontal deformation is expected, the lateral constraint conditions, (assuming no horizontal deformation) seem to be more appropriate, where:

$$S_{HI} = (\upsilon / 1\text{-}\upsilon)\ S_{VI} \qquad (10)$$

where $\upsilon = 0.25$, Poisson's Ratio, gives a relative increase in $S_{H'}$ equivalent to $0.33S_{VI}$. The presence of increased hydraulic pressure will reduce this by the value of P_f. Thus the resultant horizontal stress becomes:

$$S_{H'} = S_H + 0.33S_{VI} - P_f \qquad (11)$$

which, from 5 and 6, can be simplified to:

$$S_{H'} = S_H - 0.66S_{VI} \qquad (12)$$

This decrease in the value of effective horizontal stress will in turn cause a reduction in the value of the deviatoric stress such that:

$$S_{D'} = [(S_H - 0.66S_{VI}) - S_V] \qquad (13)$$

i.e. a reduction in $S_{D'}$ equivalent to $0.66S_{VI}$, thereby removing critically stressed crust from failure.

In the absence of sub-glacial fluid penetration into the shallow crust there would be an increase in the values of $S_{H'}$ and $S_{V'}$ equivalent to $0.33S_{VI}$ and S_{VI}, respectively. This would also cause a reduction in the deviatoric stress promoting crustal stability. This latter case would pertain at mid-crustal depths below the reach of sub-glacial fluid recharge.

Towards the base of the brittle crustal regime there would be a similar increase in stress to that found in 'dry' mid-crustal levels. However with longer periods of ice residence there may be flow of underlying ductile material away from the center of ice loading giving a decrease in σ_1 for flow parallel to $S_{H'}$ and a corresponding decrease in σ_2 for flow parallel to the direction of $S_{h'}$ (Muir Wood, 1989b). If fluid flow does penetrate to lower crustal levels as proposed by Costain *et al.* (1987), this may have an effect on the depth of the brittle ductile transition zone (Strehlau, 1990) and consequently, greatly affect the build-up of stress during the period of ice loading.

Johnston (1989) modeled the crustal stresses beneath ice sheets using both elastic and visco-elastic crustal models with differing boundary conditions and pore pressure regimes. From this, he was able to show that in compressive environments containing weak faults optimally orientated for reactivation at or near their failure threshold, the imposition of an ice cap load removed the crust from failure.

A.3 UNLOADING: ICE-MELT REBOUND AND DEGLACIATION TECTONICS

Seismic activity can be triggered by stress changes of a few tens of bars (c. 1.0 MPa). In cases where the deviatoric stress is very near to or at the failure threshold, a few bars may be sufficient to promote instability. This is dramatically shown where activities such as mining (Yerkes, 1983), reservoir impoundment (Costain *et al.*, 1987; Roeloffs, 1988) and fluid extraction (Segal, 1989) are sufficient to cause failure in critically stressed crust. Therefore, it is expected that deglaciation, with the potential to remove a load equivalent to hundreds of bars (c. 10 MPa), should trigger fault movement and seismic activity.

Naturally, the build-up of stress during the glacial period controls the amount of stress release during deglaciation. The rate of deglaciation is usually significantly faster than that for ice growth, therefore the time period for relaxation of the ice load is considerably shorter than that for load accumulation.

The immediate effect of removal of the ice cover will be the release and redistribution of the stresses imparted in the rock mass due to both the weight of the ice load and due to tectonic forces that were built up over the time of ice residence. As with crustal loading, unloading will have differing effects at different crustal levels.

At shallow crustal depths where there was formerly basal ice melting, the declining lithostatic pressure is exceeded by hydrostatic pressure. This fluid overpressuring leads to a situation where $S_{V'}$ is reduced not only due to the response of the crust as the load is removed, but also by the effect of fluid pressure, P_f, such that the effective vertical stress becomes:

$$S_{V'} = S_V - P_f \qquad (14)$$

This fluid overpressuring cannot be sustained in the shallow crust and must be released by either flow to the surface or by movement along fractures. Although deglaciation marks a period of climate amelioration, the immediate post-glacial time would be one of a climate severe enough for the development of permafrost (Ballantyne, 1984; Sutherland, 1984). This could create an impermeable layer, of the order of a few tens of meters thick, that could prevent the release of fluid overpressuring by retarding flow to the surface. The entrapment of excess fluid pressure in near-surface environments could lead to critical build-up of pressure and to sudden release of episodic fluid 'outbursts' causing dilational disruption of the near-surface fractured rock mass (Talbot, 1986; Muir Wood, 1989b). The depth (h) to which fluid pressure (P_f) exceeds lithostatic pressure (P_L) is controlled primarily by the thickness of the former ice cover (h_I), such that:

$$P_f = \rho_I g h_I = P_L = \rho_L g h \qquad (15)$$

i.e.

$$h = \rho_I h_I / \rho_L \qquad (16)$$

If we consider $\rho_L = 2.7$ Mgm^{-3} and $\rho_I = 0.9$ Mgm^{-3} this is simplified to:

$$h = 0.33 h_I \qquad (17)$$

Thus, to depths equivalent to a third of the former ice thickness, fluid escaping along sub-horizontal fractures will be capable of 'lifting' the overlying rock mass (Talbot, 1990).

Below the depth to which fluid overpressuring exists, there will be a decrease in S_V, equivalent to the stress imparted by the ice load. The reduction in S_V, is normally accomplished 'elastically' (Carlsson & Olsson, 1982), while the reduction in S_H, due to ice wastage falls by only 0.25 of the increase due to ice loading due to the slow visco-elastic response of the crust. In the case where the crust was at or near failure, this will increase the deviatoric stress, having the effect of putting the rock mass into the field of failure. This net increase in the value of deviatoric stress will be compounded by the increase in S_H, due to the long-term tectonic strain rate of the area. Such a rise in S_H, due to long-term rock-creep during the residence of the ice cover would not be relieved during the short term of deglaciation.

Towards the base of the brittle crust the removal of the ice load would also cause an increase the deviatoric stress according to the same mechanisms acting at mid-crustal (dry) levels. This would be enhanced by increases in σ_1 and σ_2 due to stress transfer from flow of ductile lower and sub-crustal material into the former center of ice loading.

Crustal uplift due to isostatic rebound has been shown to begin during deglaciation and reach a maximum immediately prior to the final disappearance of the ice cover (Mörner, 1978), possibly reflecting the increasing rate of ice wastage as deglaciation progresses (Muir Wood, 1989b). The stresses created during ice loading and subsequent deglaciation are capable of causing failure in critically stressed regions, but are rarely capable of dictating the mode of failure (Quinlan, 1984; Stein *et al.*, 1989). The high stress levels imparted by glacial loading need not be released immediately upon deglaciation in an elastic manner, but are more likely to be released in 'bursts' of activity in the period following ice wastage, decreasing in frequency and magnitude as time progresses. The presence of high stress levels due to former ice loading are shown by anomalously high stress measurements in Sweden (Carlsson & Olsson, 1982) and in Scotland (Knill, 1972). Visco-elastic behaviour will particularly affect horizontal stress levels due to lateral constraints preventing deformation over the short time of unloading. Vertical stresses will equilibrate more readily to the state of ice removal due to the presence of a free surface (i.e., the ground surface) allowing a more elastic-like response to the removal of the ice. Indeed such rebound is most marked when the time period of load removal is short in comparison to the relaxation time for the material. The rate of relaxation decreases away from newly created free (ground) surface (Nichols, 1980). If rebound is merely an elastic response to the removal of an external load it would seem that recovery should be instantaneous and involve little or no permanent deformation, that is unless relaxation creates large stress concentrations within the crust (Nichols, 1980). The instantaneous and time-dependent aspects of rebound show that geological materials behave in a visco-elastic manner when subject to unloading. Rebound in an ideal visco-elastic material occurs in response to past loading and is also influenced by the stress at the time of unloading as well as the geometry of the body undergoing relaxation. The visco-elastic response of the crust is shown by the development of deformation features such as extensional sheeting fractures sub-parallel to, and increasing in density towards, the ground surface in crystalline rock masses and the long periods of post-glacial uplift exhibited by a succession of raised shorelines in areas such as Scandinavia and Scotland. Unloading, as seen above, causes changes in the local stress field that can bring the rock mass into failure. The state of stress at shallow depths is such that it is unlikely to cause shear failure of the rock mass except along pre-existing fractures (Nichols, 1980). Failure along pre-existing discontinuities will occur when the stress ratio ($k = \sigma_1/\sigma_3$) exceeds a critical value. The development of near-surface fractures as a response to visco-elastic rebound will create an increase in permeability allowing the movement of overpressurised fluids that may in turn cause further disruption to the rock mass.

A.4 FAULT REACTIVATION

As stated previously, fault instability is created by an increase in τ due to an accumulation of strain or by a decrease in σ_n or an increase in P_f causing a reduction in the effective normal stress acting on the fault plane.

The ease of reactivation of pre-existing faults is dependent on their orientation with respect to the prevailing stress field. For cohesionless faults the conditions for reactivation are :

$$\tau = \tau_o + \mu_s(\sigma_n - P_f) \qquad (18)$$

(Sibson, 1985) where μ_s = static coefficient of friction ($0.6 < \mu_s < 0.85$) and to, the cohesion strength of the fault, is zero. The optimal orientation, θr^*, for fault reactivation with respect to that of the prevailing stress field lies close to the original angle of the fault to the stress field when it first formed, i.e. the angle predicted by *Andersonian* fault behaviour. However, if the orientation of the fault with respect to the stress field deviates significantly from this value, the stress ratio (k) needed to promote failure along the fault will increase substantially, to such a degree that it will exceed that required for the formation of a new fault (Sibson, 1990). In this case reactivation will only occur under conditions of elevated pore pressure, P_f, such that $\sigma_{3'} = (\sigma_3 - P_f)$ -> 0. Thus, for unfavorably orientated faults, reactivation will only occur if $\sigma_{3'} < 0$ or $P_f > \sigma_3$. In the absence of elevated fluid pressure, faults become frictionally locked as the angle of reactivation approaches twice that for the optimum orientation for reshear in the prevailing stress system (Sibson, 1985, 1990). Thus, conditions of fluid overpressuring in the post-glacial period are capable of reactivating unfavorably as well as optimally orientated faults.

APPENDIX B
SELECTED EXAMPLES OF POSTGLACIAL FAULTING FROM EASTERN CANADA AND NORTHEASTERN U.S.

Below are a number of examples of postglacial faults reported from eastern Canada and the northeastern United States. See Adams (1981) and Fenton (1994) for a more complete list.

(1) Adams, J. 1989. Postglacial faulting in eastern Canada: nature, origin and seismic hazard implications. Tectonophysics v. 163, p. 323-331.

(2) Hasegawa, H.S. & Adams, J. 1981. Crustal stresses and seismotectonics in eastern Canada. Earth Physics Branch, Energy Mines & Resources Canada Open File Report 81-12, 42 p.

Summarize over 70 postglacial fault localities in eastern Canada and northeastern US, most of which are also compiled in Adams (1981) and Oliver *et al.* (1970). Further examples are given from eastern Ontario/western Quebec in the vicinity of former glacial Lake Barlow-Ojibway. These are in the area around Sudbury (46.42°N, 81.15°W), Haileybury (47.45°N, 79.74°W), Larder Lake (48.12°N, 79.68°W), Duparquet (48.52°N, 79.22°W), Chelmsford (46.57°N, 81.29°W) and North Bay (46.33°N, 79.55°W). Most postglacial faults studied are reverse with offsets commonly being of the order of 10-100 mm. Movement occurred along pre-existing planes of weakness such as bedding, cleavage and joints located at, and orthogonal to, the former Laurentide ice margin.

(3) Anderson, W.A, Kelley, J.T., Thompson, W.B., Borns, H.W.jr., Sanger, D., Smith, D.C., Tyler, D.A., Anderson, R.S., Bridges, A.E., Crossen, K.J., Ladd, J.W., Andersen, B.G. & Lee, F.T. 1984. Crustal warping in coastal Maine. Geology v. 12, p. 677-680.

(4) Reilinger, R. 1987. Reanalysis of crustal warping in coastal Maine. Geology v. 15, p. 958-961.

(5) Gehrels, W.R. & Belknap, D.F. 1993. Neotectonic history of eastern Maine evaluated from historic sea-level data and 14C dates on salt-marsh peats. Geology v. 21, p. 615-618.

Anderson et al. (1984) proposed that the coastal region of Maine was subsiding at 9 mm/yr, a rate much greater than anywhere else along the eastern seaboard of the US. In addition they claimed that there had been an increase in the rate of downwarping since 1940 associated with an increase in seismic activity. Reilinger (1987) and Gehrels & Belknap (1993) highlight systematic errors in the work of Anderson et al. (1984) showing that the rate of subsidence in coastal Maine is 1-2 mm/yr, the same as elsewhere along the eastern seaboard. Gehrels & Belknap (1993) state that anomalous

delta elevations, c. 12 m lower than the regional elevation, around Machias Bay (44.75°N, 67.00°W) in NE Maine are the result of either diachronous formation or glacio-isostatically driven tectonic effects prior to 5 kyr BP, of which there is no trace today.

(6) Block, J.W., Clement, R.C., Lew, L.R., & de Boer, J. 1979. Recent thrust faulting in southeastern Connecticut. Geology v. 7, p. 79-82.

Postglacial faults occuring in steeply dipping Pennsylvanian sandstones at Attleboro (41.93°N, 71.29°W), Massachusetts, have previously been described by Woodworth (1907) and by Oliver et al. (1970). New observations indicate that the sandstones strike 050°and dip 71°NW, and that slickensides on the fault surfaces trend 337°and dip 42°NW. The northwest sides of the fault blocks are upthrust 12 to 18 mm along 157°, parallel to borehole offsets near Salem (41.46°N, 72.29°W).

(7) Bollinger, G.A., Law, R.D., Pope, M.C., Wirgart, R.H. & Whitmarsh, R.S. 1992. Geologically recent near-surface faulting in the Valley and Ridge Province: new exposures of extensional faults in alluvial deposits, Giles County, SW Virginia. Geological Society of America Abstracts Programs v. 23, p. A152.

Two extensional faults, striking NE, dipping 64-80° to the NW cut recent (Tertiary or Quaternary) alluvial terrace deposits on the north side of the New River Valley (37.20°N, 80.75°W) in southwest Virginia. Fault zones are marked by 15-20 cm wide gouge zone. Faults are 0.6 m apart, with dip-slip offsets of 1.0 and 2.8 m.

(8) Broster, B.E. & Burke, K.B.S. 1990. Glacigenic postglacial faulting at St. John, New Brunswick. Atlantic Geology, v. 26, p. 125-138.

The authors reinterpret glacial striations offset by high angle faults in the area of St. John (45.20°N, 66.10°W), New Brunswick as being the result of glacitectonic loading and unloading (ice-push) acting on topographic highs. These offsets were originally attributed to postglacial faulting by Matthew (1894a, 1894b). Their argument does not account for similar small offsets that also occur on flat glaciated surfaces.

(9) Chagnon, J.Y. & Locat, J. 1992. Offset river terrace in the Charlevoix seismic zone. Atomic Energy Control Board Canada Research Report INFO-0413 (MAGNEC Report 91-02), 57 p.

Levelling of a tilted marine/estuarine terrace on the eastern side of the Rivière du Gouffre, near Baie-St-Paul (47.45°N, 71.50°W), Quebec, revealed a possible 20 m right-lateral displacement. The inferred fault would strike NE-SW. No indication of vertical movement.

(10) Chalmers, R. 1897. Report on the surface geology and auriferous deposits of south-

eastern Quebec. Geological Survey of Canada, Annual Report X, part J, 160 p.

Postglacial faults occur at 10 localities in a northeast-trending area of "Cambro-Silurian" slates from Sherbrooke to St. Georges. At seven sites the fault throws are a few tens of mm. All are reverse, upthrown to the south. However, at three localities the offsets are much greater (of several metres), and have produced some remarkable structures. At one locality a layer of slate, 1.5 m thick and several hundred metres long, has been forced 2 m higher than the surrounding rock. At St. Evariste (46.00°N, 71.00°W) a second layer, 1.3 m thick and 180 m long, is thrust 1.7 m above the surrounding rock. The third outcrop shows a 2 m band displaced 1.3 m.

(11) Coates, D.R. 1981. Environmental Geology. J.S.Wiley & Sons International, New York, 735 p.

(12) Szymanski, J.S. & Laird, H.S. 1981. The genesis of fault-controlled buckling in thinly layered sedimentary rock. In: Proceedings of the International Symposium on the Mechanical Behavior of Structured Media, Ottawa, May 18-21 1981, p. 383-393.

A fault showing movement within the last 3.5 kyr was found during excavations at Nine Mile Point, New York (43.50°N, 76.30°W). Fault shows reverse offset of Oswego sandstone and overlying tills and lacustrine sediments. Fault trends roughly E-W, dipping to NNE. Szymanski & Laird (1981) attribute the observed offset to a stress-release buckle or pop-up occurring on a pre-existing fault.

(13) Cowan, W.R. 1980. Hudson Bay field meeting. Geoscience Canada v. 7, p. 36-37.

Makes brief mention of evidence of "postglacial rock fracturing, primarily due to isostatic effects". There are no details of location, orientations or offsets (see Mörner 1979 for details). There is no reference to postglacial faulting in the fieldtrip guidebook (Hillaire, M.C. & Vincent, J.S. 1980. Holocene stratigraphy and sea level changes in southeastern Hudson Bay, Canada. Université du Montréal, Collection paléo-Québec, no. 11, 165 pp.).

(14) Decker, C.E. 1915. Recent crustal movements in the eastern part of the Great Lakes region. Illinois Academy of Science Transactions v. 3, p. 97-100.

Along the south shore of Lake Erie, from Cleveland, Ohio, to the New York State border, Paleozoic shales are deformed by faults and folds that in some places are postglacial in age. Minor folds, parallel to stream valleys, are up to 30 m long and displace only a few metres of strata; they seem to have formed as the valleys were eroded. Major folds, transverse to the valleys, are a few metres to 150 m wide and may disturb as much as 20 to 25 m of strata. Some folds also deform overlying glacial deposits; one spectacular fold upwarping a terrace surface by 1.6 m that grades downwards within 11 m into a thrust fault dipping about 45°. No orientation data are given, and the origin of the major folds is not discussed.

(15) Dredge, L.A. 1993. Glaciotectonic structures in Eastern and Arctic Canada. Geological Survey of Canada Open File 2660, 55 p.

Although primarily a literature review of glacigenic, near surface nontectonic deformation, this report contains several incidences of postglacial glacio-isostatic faulting in eastern and northern Canada, most of which are referenced in this bibliography. A number of postglacial faults are also cited from unpublished work:

L. Dredge reports small high angle postglacial faults on striated slate surfaces on numerous outcrops in the area of Edmunston (47.38°N, 68.37°W), New Brunswick. Individual throws are up to 20 mm, and lateral displacements are 0-2 mm. See also Rampton et al. (1984).

R. Stea reports small folds and offset striations in central Nova Scotia in the area of Antigonish (45.58°N, 61.59°W) and Pictou (45.33°N, 62.57°W) counties. Orientation, offset or sense of throw are not given.

A. Seaman reports faults from Fundy Park (45.61°N, 65.04°W), New Brunswick, striking E-W in buried outcrop offsetting glacial striae by 50 mm. The direction or sense of throw are not stated.

(16) Dredge, L.A. in press. Surficial geology of northern Melville Peninsula, Northwest Territories. Geological Survey of Canada Memoir.

Small, parallel bedrock faults post-dating glaciation are found along the northern coast of Melville Peninsula (69.69°N, 83.24°W). These faults strike approximately N-S and show individual offsets of 1 to 2 m. Emergence curves for the area suggest that this faulting occurred during the period 6.5-6.8 kyr BP.

(17) Dyke, A.S., Morris, T.F. & Green, D.E.C. 1991. Postglacial tectonic and sea level history of the central Canadian Arctic. Geological Survey of Canada Bull. 397, 56 p.

(18) Dyke, A.S., Morris, T.F., Green, D.E.C. & England, J. 1992. Quaternary geology of Prince of Wales Island, Arctic Canada. Geological Survey of Canada Memoir 433, 142 p + 3 Maps.

Reports large movements along a lineament running north-south between Prince of Wales Island and Somerset Island/Boothia Peninsula (72.41°N, 96.02°W) in Arctic Canada. Suggested cumulative offsets from shoreline (9.3 ka) isobases are of the order of 60 to 100 m, with downthrow to the west. Isobases seem to suggest that this structure was active during only a short period (9.3-8 ka) following deglaciation. In addition, isobases show that Prince of Wales Island has behaved in an anomalous manner as a single block (400 km x

400 km) during uplift, shorelines only exhibiting minor tilting prior to 8 ka (in contrast to the steep gradients to the east of the Boothia structure) and remaining essentially horizontal during subsequent uplift. Onshore surface expression of faulting is restricted to a number of small lineaments trending parallel (roughly N-S) to Peel Sound on the east coast of Prince of Wales Island (71.67°N, 96.84°W). These are normally v-shaped trenches cutting across flights of raised beaches and marked by lines of active-looking mud boils where they cut across tills. When they cross bedrock the fractures disrupt glaciated surfaces, sometimes forming small (c. 1 m) east-facing scarps, antithetic to the offset suggested by the shoreline isobases. Lineaments north of Flexure Bay cut beaches of mid to late Holocene age, while those on Dixon Island descent to within a few metres of the present day shoreline.

(19) Fakundiny, R.H., Pferd, J.W., & Pomeroy, P.W. 1978a. Clarendon-Linden Fault system of western New York: Longest(?) and oldest (?) active fault in eastern United States. Geological Society of America, Abstracts with Programs v. 10, p. 42.

(20) Fakundiny, R.H., Myers, J.T., Pomeroy, P.W., Pferd, J.W. & Nowak, T.A., jr. 1978b. Structural instability features in the vicinity of the Clarendon-Linden fault system, western New York and Lake Ontario. In: Thompson, J.C. (ed.), Symposium on Advances in Analysis of Geotechnical Instabilities: Solid Mechanics Study 13, University of Waterloo, p. 121-178.

Although no significant surface fault has been related to faulting at depth, earthquake activity is associated with the 160 km long Clarendon-Linden Fault system (43°N, 78°W), New York State. Pop-ups in bedrock are evidence of high horizontal compressive stresses, but disturbed zones in glacial deposits may have resulted from ice-push or ice block melting. There is no unequivocal evidence for postglacial differential movement anywhere along the length of the fault. Also see Jacobi & Fountain (1991).

(21) Fyles, J.G. 1990. Beaufort Formation (Late Tertiary) as seen from Prince Patrick Island, Arctic Canada. Arctic v. 43, p. 393-408.

(22) Hodgson, D.A., Taylor, R.B. & Fyles, J.G. 1993. Sea level changes during the Holocene and latest Pleistocene on Brock and Prince Patrick Islands, Canadian Arctic Archipelago. Géographie Physique et Quaternaire, in press.

Fyles (1990) in reporting on the sedimentology of the Pliocene Beaufort Formation on Prince Patrick Island (77.01°N, 118.92°W), Arctic Canada shows two faults on an isopach map. One striking NE has no demonstrable offset while the other, striking NNE, shows downthrow to the west in places in excess of 500 m; by implication this faulting is of Pliocene age or younger. Faults cutting the Beaufort Formation are marked by considerable topographic expression, throws of 10-15 m, suggesting possible Quaternary movement (J. Fyles, Geological Survey of Canada, Oral Communication, 1993). The faults parallel the strike of the bedrock and are post-dated by alluvial terraces. Origin could be either glaciotectonic or seismotectonic.

(23) Gates, O. 1983. Brittle fractures in the Eastport 2-degree sheet, Maine. In: Barosh, P.J. (ed.), New England seismotectonic study activities in Maine during fiscal year 1982. US Nuclear Regulatory Commission contract NRC 04-76-291.

A benchmark in the town of Lubec (44.90°N, 67.02°W), on the eastern part of the Maine coast has dropped 169 mm, between 1935 and 1978, relative to a bench mark 5 km away to the south across the ENE-trending Lubec fault. There is no surface expression of active faulting. See entry for Anderson et al. (1984) concerning level changes.

(24) Gehrels, W.R. & Belknap, D.F. 1993. Neotectonic history of eastern Maine evaluated from historic sea-level data and 14C dates on salt-marsh peats. Geology v. 21, p. 615-618.

Anomalous marine delta elevations around Machias Bay (44.75°N, 67.00°), eastern Maine are the result of either diachronous delta formation or glacio-isostatic movements prior to 5 kyr BP. See entry for Anderson et al. (1984).

(25) Gerber, R.G. & Rand, J.R. 1978. Late Pleistocene deformations of bedrock and till, Sears Island, Searsport, Maine. Geological Society of America, Abstracts with Programs v. 10, p. 43-44.

At Searsport (44.41°N, 68.95°W), Maine, trenches excavated through glacial deposits into the weathered phyllites of an ancient fault zone revealed one southeast-dipping reverse fault with a 25 mm throw on the (pre-Laurentide) till-bedrock contact and a 150 mm intrusion of weathered phyllite into the overlying till. Other exposures show minor crumpling of the till. The till-bedrock structures are ascribed to compression of incompetent, weathered rock by ice loading.

(26) Goldring, W. 1935. Geology of the Berne quadrangle; with a chapter on glacial geology by John H. Cook. New York State Museum Bulletin 303, 218 p.

Postglacial faults are found just outside John Boyd Thatcher State Park (42.60°N, 74.00°W), approx. 1.2 km NW of Indian Ladder, west of Albany, New York State. The faults, striking 040°to 045°, offset glacially-striated surfaces. There are at least four faults with vertical displacements of 250, 140, 140 and 100 mm. A nearby fault in the bed of the Onesquethaw River, c. 3 km east of Clarksville (42.54°N, 73.89°W), has a throw of c. 300 mm. Further smaller faults are mentioned but no details are given. Also see Ruedemann (1930).

(27) Goldthwait, J.W. 1924. Physiography of Nova Scotia. Geological Survey of Canada Memoir 140, 179 p.

Postglacial faulting of striated slate outcrops is described from three locations in Nova Scotia as "ledges faulted since the ice age". At Caledonia Corners (45.27°N, 62.41°), there are two outcrops; at one three scarps strike northeast with the northwest side is upthrown in each case by 130, 180 and 200 mm. At the Prince of Wales Tower, Point Pleasant Park (44.65°N, 63.59°W), Halifax, and on Fairy rocks at Lake Kejimkujik (44.41°N, 65.26°W), other slate outcrops are upthrown to the north by step faults of 25 to 50 mm throws (see also Grant 1987b, p15).

(28) Grant, D.R. 1975. Surficial geology of northern Cape Breton Island. In: Report of activities, Part A, Geological Survey of Canada Paper 75-1A, p. 407-408.

A wave-cut platform extends south for several kilometres from Cape North at an elevation of 6±2 m. At the Aspy Fault zone, over a few hundred metres, it reappears on the south side at twice its elevation on the north. The platform may be last interglacial (c. 100 ka), and the height discontinuity could mean relatively recent movement on the flank of the Aspy evaporite-carbonate basin. See also Grant (1990).

(29) Grant, D.R. 1980. Quaternary stratigraphy of southwestern Nova Scotia: Glacial events and sea-level changes. Geological Association of Canada, Guidebook to Excursion 9.

A glacially-smoothed outcrop of slate at Cape Cove (43.04°N, 66.17°W), near Salmon River, Nova Scotia, shows regular up-to-the-south displacements along vertical 035° striking cleavage planes. More than 20 faults occur across a width of 3 m, and the throws average 50 mm, giving a total movement of more than 1 m. The outcrop is overlain by a lower grey till, the faulting having occurred prior to deposition of this till. The movements might represent postglacial crustal rebound which, rather than being a smooth warping, is concentrated along incompetent zones such as slate belts. Also see Grant (1987a) and Dredge & Grant (1987).

(30) Grant, D.R. 1987. Glacial advances and sea-level changes, southwestern Nova Scotia, Canada. in Geological Society of America Centennial Field Guide - Northeastern Section, p. 427-432.

Mentions postglacial faulting in vertically cleaved slates in the area of Cape Cove (43.04°N, 66.17°W), southwestern Nova Scotia. Striations, related to early Wisconsin glaciation, are upthrown to the south in a number of steps (slatey cleavage) by at least 1 m over a distance of 3 m. See also Grant (1980a) and Dredge & Grant (1987).

(31) Grant, D.R. 1990. Late Quaternary movement of the Aspy Fault, Nova Scotia. Canadian Journal of Earth Sciences v. 27, p. 984-987.

Describes a spectacular 15 m offset of the Sangamonian (125 ka) intertidal rock platform across the NE-trending Aspy fault on Cape Breton Island (47.00°N, 60.42°W). Offset is attributed to isostatic response to rapid ice retreat or erosional unloading, or it may be the result of the release of stress stored during the residence of the Wisconsinan glacier cover. To date this is the most impressive late Quaternary offset in eastern Canada. Map also shows minor faulting at Cape Ray, Newfoundland (47.63°N, 59.41°W). See also Grant (1975 and 1987b, p36).

(32) Grant, D.R. 1992. Quaternary geology of St. Anthony - Blanc-Sabalon area, Newfoundland and Quebec. Geological Survey of Canada Memoir 427, 60 p + Map 1610A.

Mention of a number of possible postglacial offsets in northern Newfoundland and eastern Quebec. At the southern end of Ten Mile Lake (51°N, 56.75°W), Newfoundland, a glaciated rock pavement has been upthrown by 50 mm to the north on each of two E-W joints. A larger north-facing E-W scarp, east of Baie des Belles Amours, Quebec (51.49°N, 57.50°W), vertically offsets a roche moutonée by over a metre. Other "sharp edged lineaments" were photo-mapped to the north of Baie des Belles Amours (51.70°N, 57.35°W) and northeast of Red Bay (51.90°N, 55.90°W), Labrador, most of which show upthrow to the south. A field visit in July 1993 showed most of these lineaments to be eroded basic dykes (Fenton, in prep.)

(33) Grant, D.R. 1993. A preliminary note on possible Neogene faults in Newfoundland, Canada. Bulletin of INQUA Neotectonics Commission v. 16, p. 41-42.

Around La Hune Bay (47.60°N, 56.70°W), Newfoundland, a concentration of several hundred north-facing bedrock scarps is coincident with an anomalous 500 m high dome in what is otherwise a flat, low lying, gently tilted peneplain landscape. The scarps are up to 50 m (average 5 m) in height and dips range between vertical and 70° to the north and south. A component of horizontal movement is suggested by apparent shoreline displacements along the eastern shoreline of La Hune Bay. Cumulative offset on the faults roughly accounts for the anomalous height of the dome. The age of the faulting is not clear and the age of the domed and offset peneplain surface is uncertain; it could be either Tertiary or an exhumed Carboniferous unconformity. However, the facts that the scarps remain as topographic features in an area that has been subject to repeated Quaternary glaciation, offsetting glacial landforms such as roche moutonées and also show varying degrees of glacial rounding, suggest that they have been created during Quaternary time. It remains for fieldwork to discover if any of the faults are truly postglacial.

(34) Heywood, W.W. 1968. Southampton Island, District of Keewatin. Geological

Survey of Canada Paper 69-1, Part A, p. 171.

In a short description of fieldwork activities on Southampton Island, Hudson Bay: "The Paleozoic-Precambrian contact between South Bay (64.00°N, 83.30°W) and the Duke of York Bay (65.30°N, 85.00°W) appears to be a fault scarp although in part it may be a fault-line scarp. Movement on some of the faults has occurred after post-glacial uplift."

(35) Hitchcock, C.H. 1905. The geology of Littleton, New Hampshire. University Press, Cambridge, Massachusetts, 28 p.

On the summit of Kilburn's Crag (44.30°N, 71.77°W), near Littleton, New Hampshire, a striated slate surface is offset 25 mm by postglacial faults striking roughly east-west, downthrow to the south.

(36) Hobbs, W.H. 1907. Earthquakes. Appleton & Company, New York, 336 pp.

Known occurrences of postglacial faulting are reviewed. These are confined to distributed, step-like movements of less than 125 mm in steeply dipping rocks. It is suggested that the displacements have caused earthquakes. A hinge line of deformation, separating near-horizontal from tilted lake shorelines, extends from Ashtabula (41.85°N, 80.75°W), Ohio, to Manistee (44.25°N, 86.25°W), Michigan, and might be related to local instability of the Michigan Peninsula (Hobbs 1911).

(37) Hobbs, W.H. 1911. The late glacial and post glacial uplift of the Michigan Basin. Michigan Geological and Biological Survey, Publication 5, Geological Series 3, p. 45.

At Green's Quarry (44.84°N, 87.25°W), Sturgeon Bay, Wisconsin, a polished and striated surface of the Niagara Limestone has been offset by movements of 3 to 40 mm on joints striking 345° and 083°. No note is made of the sense of movement. Local quarrying effects are discounted, and it is suggested that the offsets may be associated with a nearby hinge line revealed by a study of raised beaches (Hobbs 1907).

(38) Hume, G.S. 1925. Palaeozoic outlier of Lake Timiskaming, Ontario and Quebec. Canada Department of Mines, Geological Survey Memoir 145, 129 p.

Comments on Miller (1908) adding that the postglacial faults in the area of Cobalt strike NE-SW. The straight course of the Blanche river (47.45°N, 79.45°W), from Englehart to Lake Timiskaming, through (lacustrine) stratified clays is suspected as also being the product of rebound fracturing following deglaciation.

(39) Isachsen, Y.W. & Wold, R.J. 1977. Geodetic, geological, and geophysical evidence for Holocene vertical movements in the Adirondack region, New York. Geological Society of America, Abstracts with Programs v. 9, p. 279.

At several Adirondack sites, faulting has been found along vertical joints cutting glacially smoothed bedrock (?gneiss) surfaces. In the grounds of the Adirondack Museum at Blue Mountain Lake (43.82°N, 74.45°W), the west side of a gently-curved vertical joint plane, striking 348°, has been downthrown 10 mm. As the joint plane has been widened by weathering, this offset may have occurred as long ago as 10 kyr B.P.

(40) Isachsen, Y.W., Geraghty, E.P. & Wright S.F. 1978. Investigation of Holocene deformation in the Adirondack Mountains Dome. Geological Society of America, Abstracts with Programs v. 10, p. 49.

It is suggested that uplift of the Adirondacks is rapid. Of 270 joints measured at 55 glacially polished outcrops, 14 show offsets (dip-slip) of 0.3 to 3.0 mm, and occurring much as 35 m from blasted faces. The cause of the displacement cannot yet be assigned clearly to either tectonic causes or to ice wedging along possible hidden sheeting joints.

(41) Johnston, W.A. 1916. The Trent Valley outlet of Lake Algonquin and the deformation of the Algonquin water-plane in Lake Simcoe district, Ontario. Geological Survey of Canada, Museum Bulletin 23, 27 p.

Describes a number of irregularities in the isobases of the shorelines of glacial Lake Algonquin in the area of Lake Simcoe (44.50°N, 79.00°W) and glacial Lake Balsam. These include non-uniform spacing of the isobases, closer spacing in the north than in the south and closer spacing in the area of Beaverton on the eastern shore of Lake Simcoe. In the area of Kirkfield and Balsam Lake the isobases show a marked change in the direction of tilt and character of the uplift in this area. This latter area of irregular uplift corresponds to a fault that is "reverse in character and is evidently due to a slight buckling or thrust." Johnston claims that this is a hinge zone separating the movements that have affected the northern isobases from the undisturbed shorelines to the south. The photograph of this fault (Plate III) suggests that it is a pop-up and not a true postglacial fault rupture (see Liberty 1969; Finamore 1985). However the deformation of the isobases, locally depressed by 5 m, does suggest some form of tectonic movement since the formation of the shorelines.

(42) Karrow, P.F. 1988. The Lake Algonquin shoreline, Kincardine - Port Elgin, Ontario. Canadian Journal of Earth Sciences v. 25, p. 157-162.

Anomalously low levels (by 10-15 m) in the elevation of the shorelines of glacial Lake Algonquin are reported from the area between Kincardine and Port Elgin (44.25°N, 81.50°W), Ontario. Although this is attributed to differential erosion of surficial materials as the lake water levels fell the possibility of bedrock disturbance is not completely ruled out. An offset of "several metres of the Lake Iroquois glacial lake shoreline in Prince Edward County, Ontario" is also reported as a personal communication from W.A. Gorman (Queens University, Kingston, Ontario).

(43) Koons, D. 1989. Postglacial bedrock faulting in Maine. In: Anderson, W.A. & Borns, H.W.jr. (eds.), Neotectonics of Maine. Maine Geological Survey, Department of Conservation, p. 149-155.

Over fifty postglacial bedrock displacements (5-15 mm) are tabulated from Maine, most from the along the border with Quebec (47°N, 69°W). Faults are vertical with dip-slip offsets, but no information is given on the sense of movement. Faults parallel pre-existing structures, most striking NE-SW with subordinate NW-SE and E-W sets.

(44) Lawson, A.C. 1911. On some post-glacial faults near Banning, Ontario. Seismological Society of America Bulletin v. 1, p. 159-166.

At Banning (exact location obscure, but about 49°N, 92°W), faults offset a glaciated outcrop of dark gray, slates. The faults are bedding-parallel, striking 260° and dip to the north at 65°. Offsets reverse, upthrow to the north. There are no slickensides on the fault planes and no horizontal component of movement. Across 20 m there are 24 faults with an average offset of 22 mm up to a maximum of 95 mm. The faults can be traced for 5 to 20 m along strike, but two are seen to die out within the extent of the outcrop. One transverse fault, striking 352° and dipping east at 85°, is upthrown 25 mm to the east. Faulting has occurred sometime following the ice retreat; however, the exact timing cannot be determined.

(45) Lee, S.M. 1965. Inussuaq - Pointe Normond Area, New Quebec. Quebec Department of Natural Resources, Geological Report 119, 134 p.

Most of the faults in the area show no evidence of scarps, and it is likely that faulting was largely pre-Pleistocene, or even pre-peneplanation in age. However, south of the Kongut River near its mouth (58.40°N, 78.00°W), there are some faults in granite that are assumed to be much younger as they form vertical scarps ranging from 3 to 6 m in height. The fault planes are marked by slickensides with quartz and calcite, and their recent (?postglacial) origin may be inferred.

(46) Liberty, B.A. 1969. Palaeozoic geology of the Lake Simcoe area, Ontario. Geological Survey of Canada, Memoir 355, 201 p.

Some minor faults, 11 km northeast of Kirkfield (44.46°N, 78.87°W), Ontario, are marked by topographic relief of about 1.5 m across straight or slightly curved linears that extend for 160 to 310 m. In trenches cut across the linears, slabs on the flanks appear to dip in opposite directions. The displacement varies from a few tens of mm to 2 m or more, and one fault and an associated fold can be traced for more than 0.8 km. All the faults are normal, of small displacement and are considered to have resulted from vertical stress release following retreat

of the ice sheet as the linears have not been destroyed by glacial erosion. The faults may be superficial, as are the structural rolls and pop-ups that have occurred in quarries.

(47) Long, E.T. 1922. Minor faulting in the Cayuga Lake region. American Journal of Science v. 203, p. 229-248.

At Salmon Creek, Ludlowville (42.56°N, 76.5°W), New York State, a horizontal fault has some relation to a postglacial terrace. Although the relationship implied is not clear, the fault may displace the terrace and so be postglacial in age.

(48) Loomis, F.B. 1921. Postglacial faulting about Mount Toby, Massachusetts. Geological Society of America Bulletin v. 32, p. 75-80.

At Mount Toby (42.46°N, 72.53°W), Massachusetts, a postglacial lake shoreline is downthrown 24.4 m across a fault. The fault strikes 350° for 0.4 km, then turns abruptly and strikes 045° for 1.6 km, gradually decreasing in displacement. The first segment of the fault displaces the shoreline (regional elevation 103.6 m) down to the southwest is two steps, one of 15 m the other of 9.4 m, so that the shoreline is at an elevation of 79.2 m for a distance of 3-5 km along the southwestern side of Mount Toby. Two streams flowing over the second segment of the fault have hardly eroded a notch in the two scarps, suggesting that the scarps are relatively young features. Over 20 other "fault escarpments" ranging from 1.5 to 15 m are present on Mount Toby; these escarpments may be the result of frost heave of jointed blocks during deglaciation.

(49) Mather, W.W. 1843. Geology of New York, Part I, comprising the geology of the first geological district, 653 p.

Since this paper preceded acceptance of Agassiz's continental glaciation hypothesis, the origin of the observed grooves and scratches was then problematic. Nevertheless, it represents the earliest recognition of postglacial faulting, and so a section is quoted in full.

At Copake (42.07°N, 73.44°W), New York State, "masses of slate had been shifted a few inches in a vertical direction by a slight fault, so that the grooves and scratches on the lower part of the mass were continued quite up to that part that had been elevated; and on the upper mass, the same grooves that had been once continuous, were prolonged in their former direction, with the same breadth and depth. This shift of position, or slight fault, must have been subsequent to the period when the scratches were made". Similar faults also occur at Clinton (34 km to the southeast), where vertical strata were offset by 5 or 6 dislocations ranging from 10 to 75 mm, and at Hyde Park (41.79°N, 73.84°W).

(50) Matthew, G.F. 1894a. Post-glacial faults at St. John, N.B. American Journal of Science v. 148, p. 501-503.

Postglacial faults are described from five locations within the city of Saint John (45.10°N, 61.00°W), New Brunswick. At each locality small step faults in steeply dipping slates displace glacially striated surfaces. There are two sets of faults; a primary set striking northeast-southwest and a secondary "diagonal" set striking north-south and east-west. With one exception, the faults are upthrown to the southeast with throws ranging from 5 to 130 mm, the average offset being 25 mm. The faults are reverse and dip from 70° to 90° SE. Displacement along individual faults may change along strike and the simple movement on the NE-SW faults is complicated by displacements on the diagonal set. At almost every suitable outcrop of glaciated bedrock in the slates of the "St. John Group", some displacement (often slight) can be seen. See also Broster & Burke (1990).

(51) Matthew, G.F. 1894b. Movements of the earth's crust at St. John, N.B., in Post-Glacial times. Bulletin of the Natural History Society of New Brunswick v. 3 (12), p. 34-42.

The postglacial faults within Saint John (45.10°N, 61.00°W), New Brunswick, described in Matthew's earlier (1894a) paper are discussed in more detail. The displacements of individual faults at two localities are given. At City Road, 9 faults with a total displacement of 250 mm occur over a 3.5 m wide outcrop. At Rock Street, 62 faults with a total displacement of 1.7 m occur over 48 m. Postglacial faults (additional to those in the 1894a paper) also occur at the east and west ends of Peel Street and at Pond Street, and are similar to those described above. Two objects, a boulder and a lance head, both broken, displaced and re-cemented, that Matthew attributed to earth movements might well have been subject to the effects of frost action. Faulting is considered to be the result of tectonic thrusting from the SE. See also Broster & Burke (1990).

(52) Miller, W.G. 1908. The cobalt-nickel arsenides and silver deposits of Temiskaming (3rd ed.). Ontario Bureau of Mines Report XVI, part II, 212 p.

In the vicinity of Cobalt and Lake Temiskaming (47.40°N, 79.60°W), "it can be proved that slight faulting, at least, has taken place in post-glacial times". No further details are given, and the sentence does not appear in the 2nd edition (v. XIV, part II, 1905). Hume (1925) commented positively on the above, but found no firm proof of such recent movements in his area.

(53) Mörner, N-A. 1979. Activities of the INQUA Neotectonics Commission. Bulletin of INQUA Neotectonics Commission v. 2, p. 7-8.

Reports on an excursion to Hudson Bay in 1979 (Cowan 1980) where "numerous examples of postglacial fracturing (and some dm vertical offset) of the bedrock were examined (at Poste-de-la-Baleine, 55.30°N, 77.30°W), for example, perfectly striated and grooved bedrock surfaces were fractured, often with one side or block moved vertical with respect to the main surface." In one area of the LG-II dam project, these "rebound features" were said "suddenly to increase in frequency below the 100 m depth (c.f. the report of Mörner)." Unfortunately the Mörner report was never written (N-A. Mörner, University of Stockholm, *personal communication*, 1993).

(54) Newland, D.H. 1933. Earthquakes in New York State. New York State Museum Circular 14, 18 p.

At Altamont (42.66°N, 74.05°W), near Albany, New York State, an ice-smoothed sandstone floor is fractured and displaced c. 250 mm along a fault plane. The offset is shown in a photograph. No orientation data are given. Possibly the same locality near John Boyd Thatcher State Park described by Goldring (1935).

(55) Oliver, J., Johnson, T., & Dorman, J. 1970. Postglacial faulting and seismicity in New York and Quebec. Canadian Journal of Earth Sciences v. 7, p. 579-590.

Eleven new occurrences of postglacial faulting and a review of previous literature are given. Along the Hudson River newly-reported faulting occurs at Bulls Head Road (41.94°N, 73.66°W), on the Taconic State Parkway (42.28°N, 73.53°W), and at Fair Haven, Vermont (43.58°N, 73.26°W). In the area of postglacial faulting described by Chalmers (1897) in Quebec, four new locations were found. In western Ontario postglacial faults are found at Beardmore (49.50°N, 88.00°W), Flanders (49°N, 92°W), Longlac (50°N, 87°W) and Shebandowen (49°N, 90°W). At each locality the faults occur in slates and the fault throws are a few tens of mm or less.

A zone of recent faulting begins near Hyde Park east of the Hudson River, passes east of Lake Champlain, and then trends northeast parallel to the St. Lawrence River as far as St. Georges. The faults parallel the trend of the zone, and, nearly always, the east or south side is upthrown.

(56) Pavlides, L., Bobyarchick, A.R., Newell, W.L. & Pavich, M.J. 1983. Late Cenozoic faulting along the Mountain Run Fault zone, central Virginia piedemont. Geological Society of America Abstracts with Program (Southeastern Section) v. 15, p. 55.

Near Everona (38°N, 78°W), Virginia, a possible late Tertiary age debris flow and underlying saprolite is offset by c. 1.5 m on northward-dipping, S and SW directed low angle thrusts. Minimum age of this faulting episode has not been determined. Where the fault cuts gravels, the clasts have been imbricated. Possibly related to the Mountain Run fault zone.

(57) Ruedemann, R. 1930. Geology of the Capital District. New York State Museum Bulletin 285, 218 p.

Describes a number of postglacial faults near Albany, New York State. One fault with a throw of 40 cm in the

Oriskany Sandstone forms a scarp in a stream bed in Onesquethaw Creek, c. 3 km east of Clarksville (42.54°N, 73.89°W). This fault is considered "so recent that it has not even been channelled by the stream." Neighbouring faults are described by Goldring (1935).

(58) Shilts, W.W., Rappol, M. & Blais, A. 1992. Evidence of late and postglacial seismic activity in the Témiscouata-Madawaska Valley, Quebec - New Brunswick, Canada. Canadian Journal of Earth Sciences v. 29, p. 1,043-1,069.

Report an offset glaciated pavement exposed in a sand pit outside Saint-Jacques, New Brunswick (47.45°N, 69.45°E). A striated slate outcrop is offset vertically by 70 mm. The fault strikes 150-330°, downthrow to the NE. The fault is open but is not infilled with sediment. Several smaller extensional fractures are associated with the main fracture; these are normal to the main fracture and show little lateral or vertical displacement. Martin (1988) reports offset glacial striae NW of St. Jacques, from unpublished work by M. Rappol. Here a 045° striking fracture in slate shows vertical offset of 70-80 mm (no direction of movement given).

(59) Stanley, G.M. 1936. Abrupt decline of the Whittlesey beach at Birmingham, Michigan. Papers Michigan Academy of Science Arts & Letters v. 21, p. 445-452.

In the area of Birmingham (Detroit), Michigan (42.30°N, 83.15°W) the Whittlesey shoreline (of glacial Lake Whittlesey) shows an abrupt decline in elevation of c. 3 m. To the SSW of Birmingham the shoreline is horizontal with an elevation of c. 226 m. The shoreline suddenly drops 3 m to an elevation of 223 m. The shoreline then increases in altitude to the NE at 0.19 m/km. Faulting coincident with "one of the hinges of (postglacial) uplift" is proposed as the cause of the shoreline offset.

(60) Thompson, W. 1979a. Postglacial faulting along the Norumbega Fault Zone. Geological Society of America, Abstracts with Programs v. 11, p. 56.

(61) Thompson, W.B. 1979b. Postglacial faulting along the Norumbega Fault Zone. Maine Geologist (Geological Society of Maine Newsletter) v. 5, p. 6.

(62) Thompson, W.B. 1981. Postglacial faulting in the vicinity of the Norumbega fault zone, eastern Maine. Maine Geological Survey Open File No. 81-48, 22 p. (also USGS Open File Report 81-1039).

In the area between Bangor and Calais, Maine, (45°N, 68°W) the NE-striking Norumbega Fault Zone six exposures show postglacial displacements on northeast-striking bedding plane faults. Four of these are within the Norumbega Fault Zone, but are not directly on any of the principle fault traces, and the other two are on, or close to, subsidiary faults 8 to 10 km to the southeast. The postglacial faults have throws of 1.5 to 30 mm, with little or no horizontal displacement. The scarps face in various directions, and rarely extend

more than a few metres along strike. One fault extends from the bedrock into the overlying till for 100 mm or more. Late-glacial frost heaving may account for some of the faulting, and regional tectonics for the rest.

(63) Westerman, D.S. 1983. Structural analysis of the Guilford, Dover-Foxcroft, and Boyd Lake 15-minute quadrangles, south-central Maine. Maine Geological Survey Open File Report 83-7, 22 p.

South of Boyd Lake (45.15°N, 68.88°W), Maine, a postglacial fault striking 055° vertically offsets striae along strike for c. 20 m. Displacement is c. 7 mm, downthrow to the southeast. Outcrops along Route 15 also show offset of striae, downthrow to the SE along 045° striking cleavage planes.

(64) Wones, D.R., & Thompson, W. 1979. The Norumbega fault zone: a major regional structure in central eastern Maine. Geological Society of America, Abstracts with Programs v. 11, p. 60.

Across the northeast-striking Norumbega fault zone (about 45°N, 68°W), glacially-derived landforms appear undeformed. However, vertical displacements of a few centimetres have been observed in glaciated pavements; lateral displacements have not been observed.

(65) Woodworth, J.B. 1905. Ancient water levels of the Champlain and Hudson Valleys. New York State Museum Bulletin 84, 265 p.

Postglacial faults occur at Defreestville (42.61°N, 73.65°W), New York State, and slightly to the south on the banks of the Hudson River at Greenbank and "is indicative of a measurable change of local levels in the terrace of this part of the valley".

(66) Woodworth, J.W. 1907. Postglacial faults of eastern New York. New York State Museum Bulletin 107, 28 p.

East of the Hudson River, New York State, postglacial faults occur in four localities:

At South Troy (42.75°N, 73.70°W) reverse faults in slates strike 009° and dip 40°E. The faults show systematic downthrow to the west of 25 to 130 mm, with a total displacement of 305 mm across a 9 m wide exposure. One fault runs normal to the others and is upthrown on the north side. Two faults converge southward and die out; the others extend to the limit of the outcrop.

In Rensselaer (42.40°N, 73.70°W) there are postglacial faults with a total throw up to the east of 130 mm. At Defreestville seven postglacial faults with a total throw up to the east of 330 mm displace a 3.5 m wide outcrop of slates that strike 026° and dip steeply.

At Copake (42.08°N, 73.44°W) 13 postglacial faults with total throw up to the east of 200 mm displace a 3.5 m ice-smoothed slate outcrop. In a second exposure, 32 faults are

upthrow a total of 180 mm to the east in a 3.5 m wide outcrop. Full details of the individual displacements are tabulated. Two faults overlap and die out in opposite directions, the throw on one continuing the throw from the other.

Near Pumpkin Hollow (42.06°N, 73.74°W) some 430 mm of displacement up to the east occur on postglacial faults in a 60 m exposure. Postglacial faults also occur at Attleboro (41.93°N, 71.29°W), Massachusetts, where a striated exposure of vertical Palaeozoic sandstone striking 052° is offset a total of 76 mm over a 30 m wide exposure. A 300 mm throw may also be present.

At two locations the evidence for postglacial faulting is less clear. Near Port Kent (44.52°N, 73.40°W), New York State, on the west shore of Lake Champlain, a postglacial lake bench appears to be upthrown to the east across a north-south trench that may mark a bedrock fracture. On Mt. St. John (45.40°N, 73.35°W), Quebec, there is extensive fracturing and some dislocation of the bedrock that may represent recent faulting.

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FIELD-BASED IDENTIFICATION OF SALT-RELATED STRUCTURES AND THEIR DIFFERENTIATION FROM TECTONIC STRUCTURES

Peter W. Huntoon[1]

EXECUTIVE SUMMARY

There are three broad classes of salt related structures: (1) salt flowage structures including salt anticlines, salt domes, salt diapirs, and growth faults; (2) salt-dissolution and collapse features ranging in size from small basins downward to breccia pipes and sinkholes, and (3) gravity glide structures having the form of large, low-angle landslides. Salt structures tend to be very dynamic. If salt remains, it is reasonably certain that deformation is on-going. Consequently opportunities exit for sensing it.

A considerable literature has developed around salt-tectonic structures, their differentiation from deep-seated orogenesis, and their differentiation from each other based on causative process. Huntoon differentiates three diverse classes of salt structures based on causative process: flowage, dissolutional collapse and glide.

Salt flowage structures include salt anticlines, salt domes, salt diapirs and growth faults which are characterized by uplift and extension within the rocks over the accumulating salt cores (Cater). Stress and strain mechanics in salt domes and diapirs are treated in Kupfer and Muehlberger and Clabaugh.

The dissolution and collapse structures associated with salt bodies include collapse anticlines, subsidence basins, sinkholes and breccia pipes. The common characteristic of these structures is land subsidence associated with dissolution of the underlying salt body. Dissolution collapse of salt anticlines is the focus of Doelling and Sugiura and Kitcho. The development of subsidence basins and superposition of breccia pipes on them are examined in Huntoon and Richter.

Gravity glide structures result from sliding of the rocks above the salt, but they are not particularly common. Huntoon summarizes a well-exposed glide structure, identifies the causative stresses, and classifies the resultant strain features.

The differentiation of faults and folds having a salt tectonic origin from those related to other causes is a problem of proving (1) that the structural element in question resulted from deformation within the salt - either flowage, dissolution or gliding - and (2) that the deformation does not affect rocks older than the salt body.

Extremely important is that fact that faults associated with salt structures have the identical structural and geomorphic appearance as faults of numerous other origins. Their form and the stress regimes that caused them are non-unique. Consequently the issue of discriminating between faults associated with salt structures and those of other origins is an exercise in properly placing them into their larger geologic context.

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Salt structures should be anticipated if the presence of salt is revealed in a region through a survey of the stratigraphic literature, verification from drilling records, and/or outcrop observations.

Geologic and geomorphologic mapping that focuses on Quaternary deformation, Quaternary sedimentation, and youthful landforms will reveal evidence for surface forms and instabilities that may be related to salt tectonism.

Regional structural mapping is a valuable method for discriminating between salt tectonic and deep-seated tectonic structures. The objective of regional mapping is to correlate the trends and styles of deformation between areas underlain by salt and adjacent areas in order to identify structures that are exclusively of salt tectonic origin.

Construction of geologic cross sections and isopach maps, coupled with geophysical profiling, provides a stratigraphic context for locating and differentiating between salt-related and deep-seated faults. The most reliable, non-intrusive means for discriminating between salt-related and deep-seated structures is seismic profiling. Drilling can be employed to assess the presence of suspected buried faults.

Gravity glide structures are characterized by fissures and fault scarps at the head of the slide, and creep features, thrust faults and related folds at the base of the slide. The three dimensional form of the detached mass, including slip planes can be interpreted from field mapping and seismic profiling.

Micro seismic surveys can be useful in discriminating between active salt tectonic and deep-seated faults because the focal points for the salt tectonic faulting is restricted to the salt or the overburden rocks. The failures in the overburden rocks above deforming salt bodies generally have low seismicity because the faulting is extensional. Wong, Humphrey and Silva (1987) employed microseismic techniques to differentiate between salt tectonism and deep-seated tectonism in a salt province.

The long geologic deformation histories in typical salt structures is critical to their identification. Cater and Seni and Jackson summarize evidence for growth in salt anticlines, domes and diapirs. Coleman relates salt growth and dissolution within a diapir to sedimentation. Worrall and Snelson document salt growth faulting within the salt structures in the Gulf coast region.

1 INTRODUCTION

The presence of salt in a geologic section leads to structural instability because the salt behaves as a viscous fluid that is continuously deforming. The resulting salt tectonic structures are primarily produced as a result of differential stresses caused by loading. Many salt structures are unloading structures, owing their origin to differences in lithologic loading caused by valley or canyon incision.

There are three broad classes of salt structures based on the processes involved in their development. (1) Flowage of the salt produces salt anticlines, salt domes, salt diapirs and growth faults. (2) Dissolution of the salt produces collapse and subsidence structures such as collapsed salt anticlines, structural basins and breccia pipes. (3) Gravity glide of rocks above the salt results in structures similar to large, low-gradient landslides in which the underlying salt serves as viscous décollement surface.

Most salt structures have been classified as salt growth features, which implies that the features develop over sustained periods of geologic time, either continuously or episodically. Consequently the manifestations of strain in the rocks overlying the salt reveals a record of long term activity. The oldest parts of such structures exhibit far more strain than the younger parts. One example is thinning of strata over the crests of salt anticlines, where every sedimentary layer above the deforming salt thins as it is traced over the rising crest of the fold. The deformed section can represent continuous activity spanning a long interval of geologic time, being in excess of 300 million years in the cases of deforming Pennsylvanian salt bodies. The oldest unit directly above the deforming salt is steeply folded and highly fractured, whereas the youngest unit is virtually undeformed and largely unfractured. Another example is the Gulf Coast salt growth faults, in which the accumulating sedimentary units on the downdropped side are thicker relative to their counterparts on the upthrown side (Worrall and Snelson, 1989).

One primary issue involved with characterizing salt tectonic structures is acknowledging that the complexity of deformation within the overburden increases with depth because the older rocks have experienced a longer history of deformation. The likelihood of concealed faults is high. A second important consideration is that salt-related deformation is restricted to the salt and rocks above it.

2 SALT TECTONIC STRUCTURES

Three broad classes of deformation structures occur in geologic provinces underlain by thick salt deposits: (1) salt flowage, (2) salt dissolution and (3) overburden glide features (Huntoon, 1988). All of these are classified as gravity tectonic structures because they owe their origin to differential strain caused by loading.

2.1 Salt Flowage Structures

2.1.1 Characteristics and Geometries

Salt flowage structures include in decreasing order of scale growth faults, salt-cored anticlines, salt domes and salt diapirs. These structures are characterized by accumulation of salt in their cores - usually as a result of lateral flowage - and uplift of the overlying rocks. Thrust faults and recumbent folds predominate in the flowing salt. Salt flowage structures are complimented by local or regional stratigraphic thinning equal to the volume of the displaced salt in their cores. For example, salt anticlines are paralleled by synclines from which the salt has flowed.

The salt anticlines of the Paradox Basin, Utah and Colorado, are among the best documented in the United States (Cater). Individual anticlines are up to 80 kilometers in length and are spaced from 10 to 15 kilometers apart. The structural relief between the synclines and anticlines can range up to a kilometer or two. In contrast, salt domes and diapirs are smaller, most being less than two kilometers in diameter (Figure PH.1). Salt domes and diapirs generally rise from the cores of salt anticlines in provinces that contain salt anticlines. The salt growth faults of the Gulf Coast region have lengths comparable to salt anticlines.

One common attribute of salt provinces is a sustained record of deformation that can span tens to hundreds of millions of years. This contrasts to the geologically shorter duration of mountain orogenesis that usually spans less than tens of millions of years. The longevity of deformation is useful in discriminating between salt structures and deep-seated tectonism. For example, a stratigraphic record that illustrates continuous sedimentary thinning spanning more than 100 million years over a growing salt anticline is not typical or expected in an environment where only deep-seated tectonism has operated.

Successive sedimentary layers bury the developing salt structures in areas in which the land surface has or is undergoing subsidence. The result is a pattern of stratigraphic thickening over areas undergoing salt depletion and thinning over areas of salt

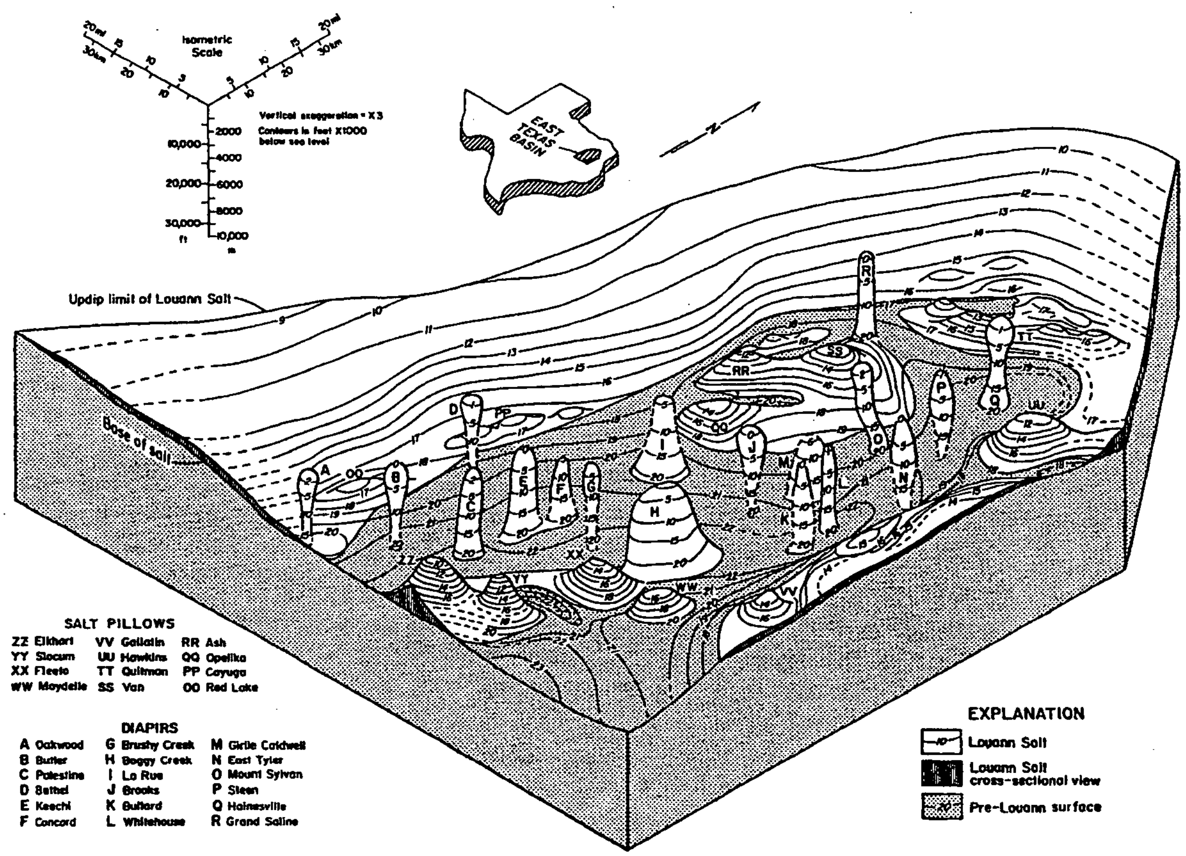


Figure PH.1 Block diagram of salt domes and diapirs in the East Texas Basin showing the three-dimensional configuration of structural contours on top of the Louann Salt or, where the salt is absent, on top of pre-Louann basement. From Seni and Jackson (1984, figure 6A). Permission to use this copyrighted material is granted by the Bureau of Economic Geology, University of Texas.

accumulation. Episodic high rates of salt flow are often revealed in the sedimentary record as angular unconformities and pinchouts over the positive areas. In extreme cases, piercements occur so that the layered rocks abut the rising caprock or even salt cores in the structures.

Sedimentation patterns can also reveal active salt tectonism in uplifted areas. Positive features, such as growing salt anticlines and domes, stand in relief whereas adjacent areas are commonly infilled. The presence of depocenters reveals the presence of subsidence or collapse features over areas of active salt dissolution. However, deposition also results from blocked drainages or isolated topographic lows caused by adjacent rising positive elements. For example, a stream that crosses a rising salt anticline will be characterized by a diminished gradient and deposition upstream of the axis, whereas downstream the channel will be oversteepened and scoured.

Diapirs are roughly circular or oval structures with sharp contacts between the wall rock and the intruded material (Figure PH.2). Diapirs are surrounded by ring fractures in the wall rock. The intrusive rocks are older rocks, caprock materials or the salt itself. The intruded strata is commonly

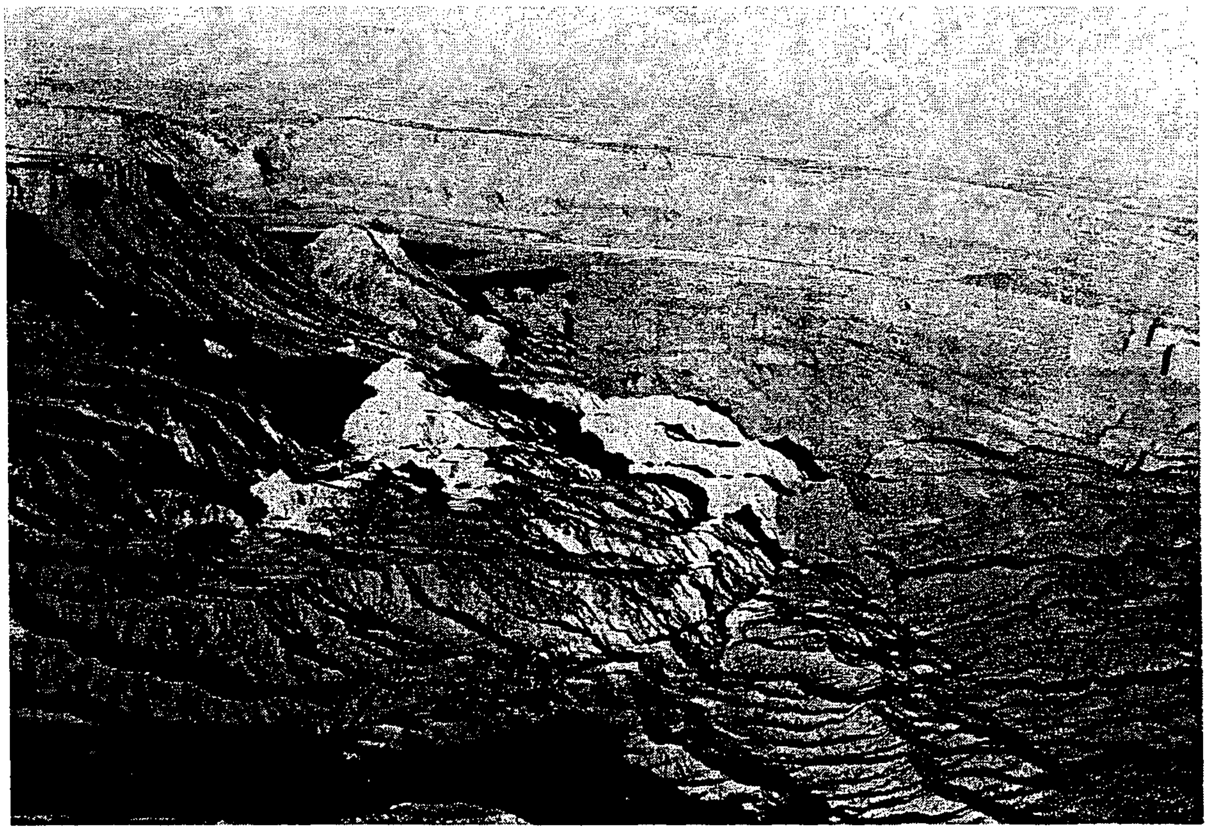


Figure PH.2 Photograph showing the Onion Creek salt diapir (the white mass in the lower center) comprised of the Pennsylvanian Paradox salt, Cache Valley Salt anticline, Utah. The upturned wall rock adjacent to the right of the diapir is the Permian Cutler Group. The elongate diapir measures 4 by 2.5 km. View is toward the north.

altered through reduction, as are some ring fractures. If the evaporites are present, such as the salt itself or gypsum residues, the bedding in this material is highly contorted and ductily thinned. The bedding appears extruded (Figure PH.3). A visual analog is sheets of paper laid flat against the bottom of a board containing a hole. The paper is then forcefully pushed upward through the hole with a finger. The bedding in the intruded mass is crenulated when viewed from above and there are numerous structural unconformities that sup-parallel the near-vertical bedding. Drag folding is common in the wall rocks surrounding the intruded material. The faults present, including ring faults and faults internal to the intruded material, are predominantly high-angle reverse faults with numerous conjugate shears. If rates of salt dissolution exceed upward intrusion rates, the rocks within the diapir exhibit collapse characteristics.

2.1.2 Processes of Formation

The stress regimes in salt flowage structures are variable and thus dependent on position within the structures (Muehlberger and Clabaugh, 1968). In regions distant from active compressional orogens, the near-surface stresses in the overburden along the crests of salt anticlines and in the immediate vicinity of salt domes are usually dominated by vertical maximum principal stresses.

Consequently the faulting associated with these structures at these locations are high-angle, reverse faults with dip-slip (Kupfer, 1968). In contrast, maximum principal stresses tend to parallel the flow lines in the salt bodies at depth and thus are parallel to the bedding enclosing the salt body.

Salt anticlines, domes and diapirs commonly develop over extended periods of time (Seni and Jackson, 1984). Growth rates can range from fairly uniform to episodic. If the structures occur in settings undergoing active deposition, the various units will thin over the structures during periods of growth (Figure PH.4). Unconformities can develop between units when growth rates are particularly high. The older, deeper strata, which have longer histories of deformation, exhibit more complex faulting and folding than the younger strata (Figure PH.5). Consequently, numerous

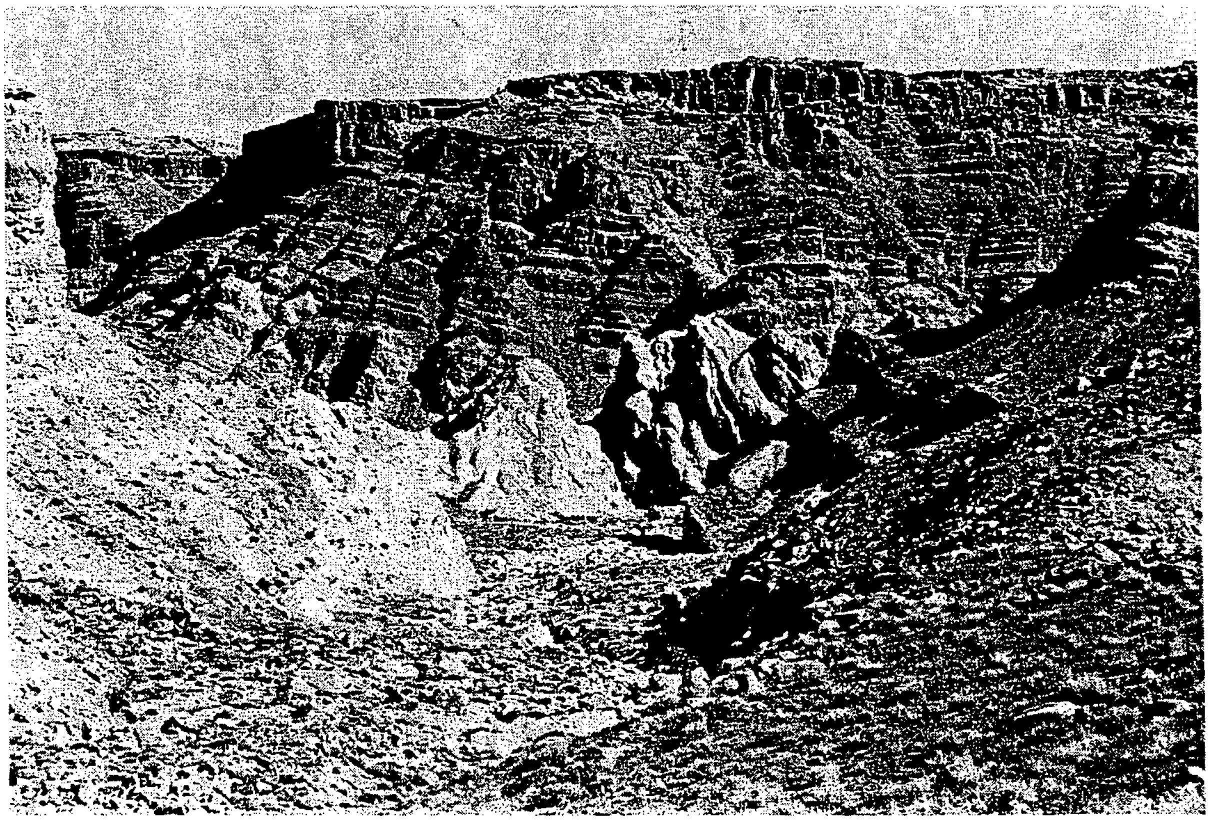


Figure PH.3 Photograph showing the Crum dome, a 0.5 km diameter salt diapir, which has pierced the floor of Cataract Canyon, Canyonlands National Park, Utah. The white rocks in the center are intruded Pennsylvanian Paradox strata. The salt has dissolved so that each cubic foot of rock in the piercement represents six cubic feet of intruded Paradox lithologies. The bedding in the remaining gypsum and insoluble clastics thins toward the center, contain numerous internal unconformities, and appears crenulated when viewed from above. View is toward the west.

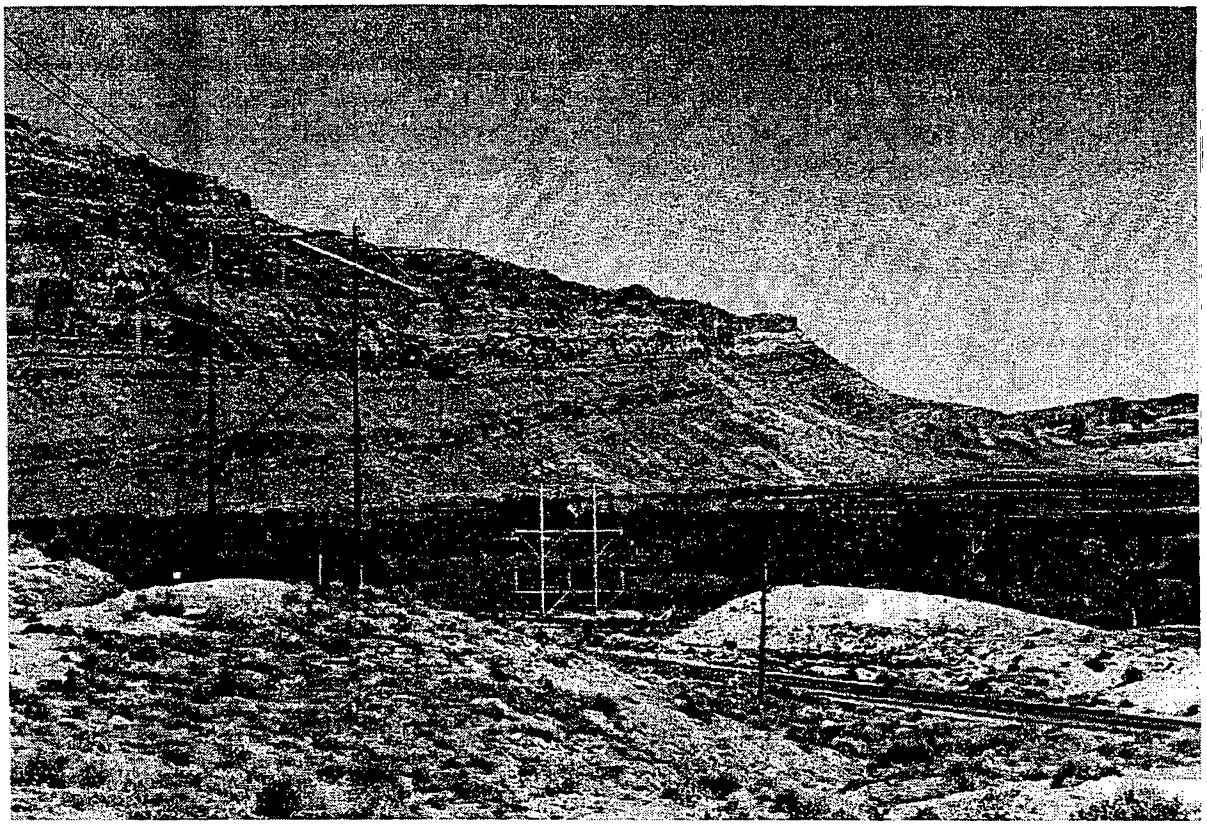


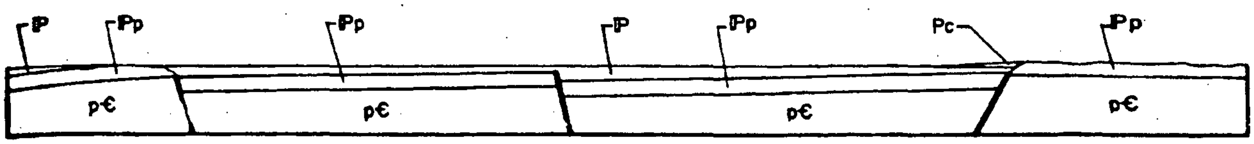
Figure PH.4 Photograph showing the Spanish Valley salt anticline, Moab, Utah. Every sedimentary unit thins from left to right toward the crest of the anticline. Strata visible here is Pennsylvanian through Jurassic in age. The Pennsylvanian Paradox salt, which has flowed into the core of the anticline, crops out as rounded gypsum caprock hills in the foreground. View is toward the north.

blind faults are possible in these structures, and their density increases with depth.

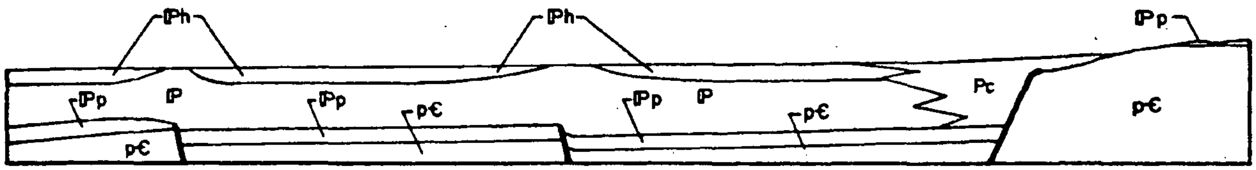
2.2 Salt Dissolution Features

2.2.1 Characteristics and Geometries

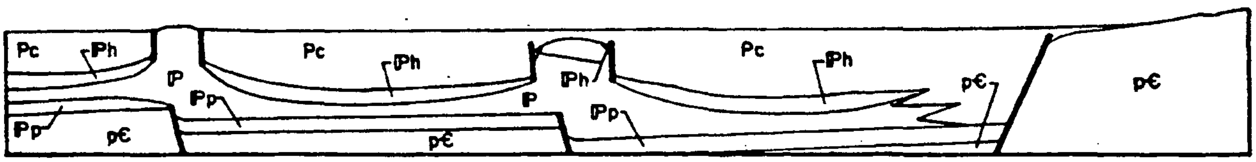
Salt dissolution structures include in increasing size: small-scale sinkholes measuring a few meters across, vertically extensive breccia pipes, dissolution basins measuring a few kilometers across (Figure PH.6), and large-scale valley collapses measuring tens of kilometers in length and a few kilometers in width (Sugiura and Kitcho, 1981). In extreme cases, entire salt bodies have been removed under many tens of square kilometers. Dissolution of the salt body is the causative mechanism, where salt removal occurs primarily as a result of ground water circulation. In breccia pipes, the pipe cores are comprised of brecciated roof rocks that have fallen or subsided into the upward stoping structure (Figure PH.7). The wall rocks adjacent to the collapse usually dip inward. In the case of collapsed salt anticlines,



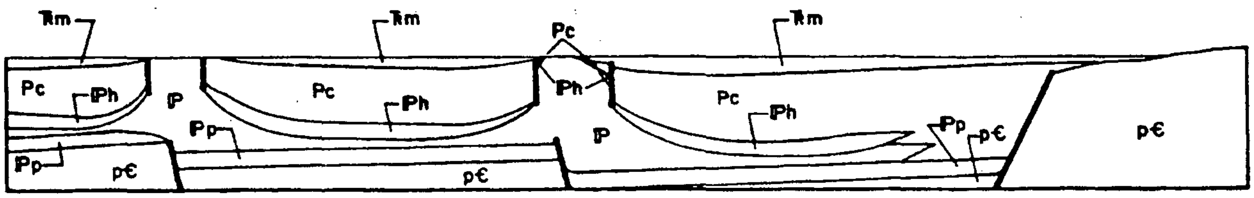
A. Precambrian and Paleozoic rocks during deposition of the paradox formation (middle Pennsylvanian) after early downfaulting and subsidence of deep part of Paradox Basin and uplift of Uncompahgre uplift.



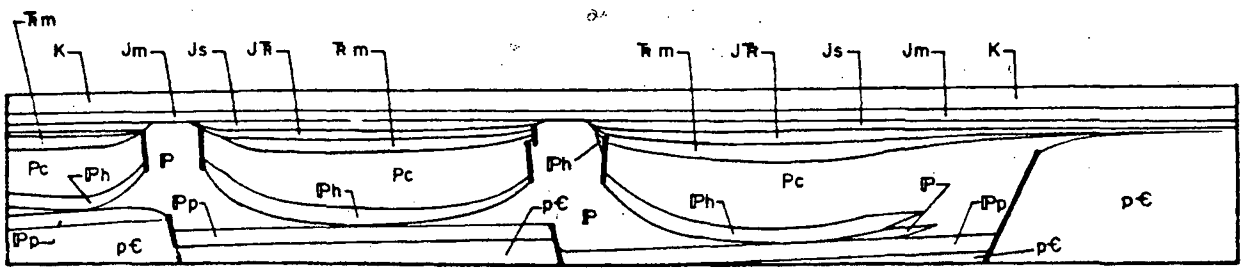
B. Precambrian and Paleozoic rocks at end of Honaker Trail deposition (late Pennsylvanian). Downfaulting and subsidence of basin and uplift of highland continuing.



C. Precambrian and Paleozoic rocks near end of Cutler Deposition. Development of salt cores well advanced.



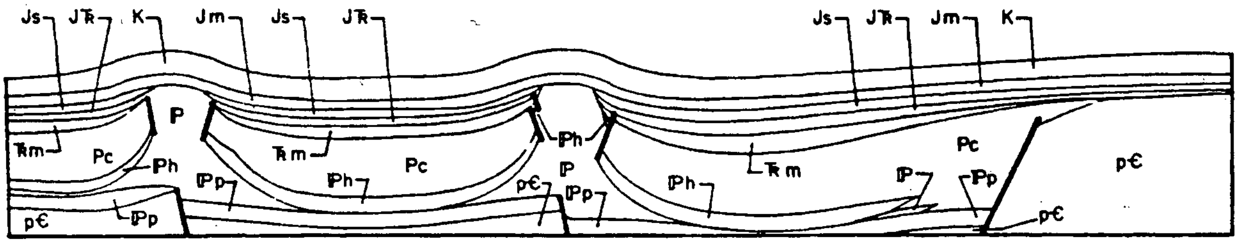
D. Precambrian, Paleozoic and Mesozoic rocks at end of Meonkopi Deposition (Triassic).



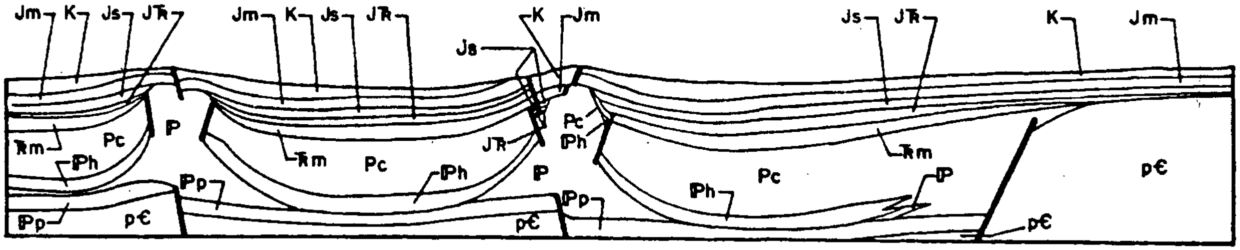
E. Precambrian, Paleozoic and Mesozoic rocks at end of Cretaceous.

Figure PH.5 Structural evolution of the Gypsum Valley-Paradox Valley salt anticline, a typical salt anticline in the Paradox Basin of Utah and Colorado. Figure from Woodward-Clyde Consultants (1983, figure 6-8) as modified from Cater (1970, figure 13). Permission to use this copyrighted material is granted by the Battelle Memorial Institute.

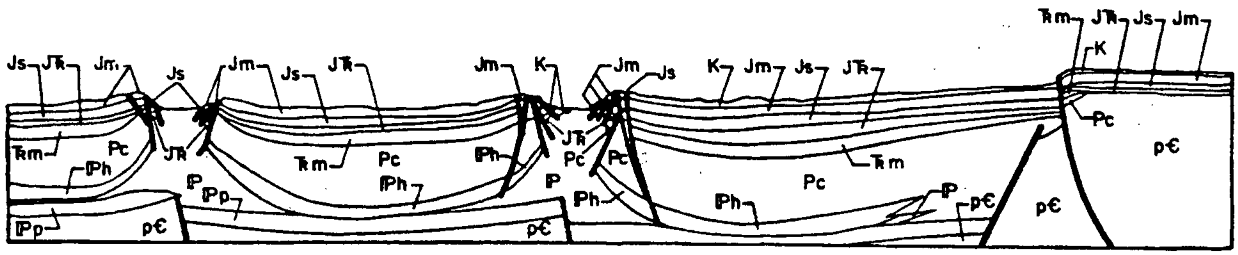
Figure PH.5 (Continued)



F. Precambrian, Paleozoic, and Mesozoic rocks at end of early Tertiary folding.



G. Precambrian, Paleozoic, and Mesozoic rocks during first stages of crestal collapse of salt anticlines (early Tertiary).



H. Precambrian, Paleozoic, and Mesozoic rocks since Pleistocene time. Renewed faulting on southwest boarder of Uncompahgre Plateau during late Tertiary and Pleistocene.

EXPLANATION

K	CRETACEOUS ROCKS	Pc	CUTLER FORMATION
Jm	MORRISON FORMATION	IPh	HONAKER TRAIL FORMATION
Js	SAN RAFAEL GROUP	IP	PARADOX FORMATION
JTR	GLEN CANYON GROUP	IPp	PINKERTON TRAIL FM AND PRE-PENNSYLVANIAN PALEOZOIC ROCKS
TRm	MOENKOPI FORMATION	p€	PRECAMBRIAN ROCKS

parallel rows of grabens occur within and immediately adjacent to the axis of the structure, and major normal faults tend to trend parallel to the axis (Figure PH.8) (Doelling, 1983).

Importantly, these faults are restricted to the rocks above the salt. Chevron folds of all scales that trend parallel to the trends of the collapsed anticlines are also common. Conjugate shears pervade the collapsed rock to belie the vertically-oriented causative maximum principal stresses. The downward displaced rocks are usually highly altered owing to circulation of brines through them.

2.2.2 Processes of Formation

Collapse structures ranging in size from sinkholes to collapsed salt anticlines exhibit downward displaced overburden rocks, significantly increased fracture densities within the displaced material, and a preponderance of inward-dipping normal faults. Maximum principal stresses usually are vertical in regions experiencing salt dissolution. Consequently, displacements associated with salt dissolution structures are downward. Faulting characteristically occurs along high-angle normal faults with simple dip-slip.

Breccia pipes occur within the overburden rocks within the larger collapse structures. The pipes nucleate from space created in the dissolving salt under the structure or extensional space created in the collapsing overburden (Huntoon and Richter, 1979). The pipes stope upward, maintaining remarkably constant diameters. The infallen or subsided rocks in the pipes are all displaced downward. The wall rocks are usually folded downward toward the pipe and contain ring fractures that surround the pipe. Bleaching of the breccia cores is common attesting to upward circulation of brines through the pipes. The cemented breccias in many pipes stand in relief above their surroundings in eroded terranes.

2.3 Gravity Glide Structures

2.3.1 Characteristics and Geometries

Gravity glide structures can develop above a salt layer where the overburden plate moves toward topographic lows. The salt serves as a viscous glide surface. These structures can be considered to be a type of large, low-relief landslide, with many shared strain characteristics.

The best documented salt-floored gravity glide structure in the United States is the Needles fault zone in Canyonlands, Utah (Huntoon, 1982). This slide feature involves a surface area of more than 200 km^2, and involves a detached plate that is approximately a kilometer thick. Undoubtedly this is not the largest salt glide structure in the United States.

2.3.2 Processes of Formation

The structure of glide sheets is very similar to that of low-angle landslides (Figure PH.9). A décollement surface underlies the entire glide sheet, and the salt surface upon which the plate slides serves as a viscous shear surface. The rocks comprising the plate usually are internally deformed by arcuate grabens, concave toward the direction of motion, which progressively develop upslope as space is created at the toe of the detachment (Figure PH.10). The normal faults merge downward into the décollement. The moving plate can push up an anticline in the rocks at the leading edge of the slide (Figure PH.11). If the mass is moving toward a valley and the thickness of the rocks between the valley floor and top of the salt are minimal, the décollement can propagate upward, allowing the plate to overrun both the valley and anticline. The motion occurs on one or more thrust planes that dip back toward the advancing mass.

Stress regimes in gravity glide structures are analogous to those found in landslides.

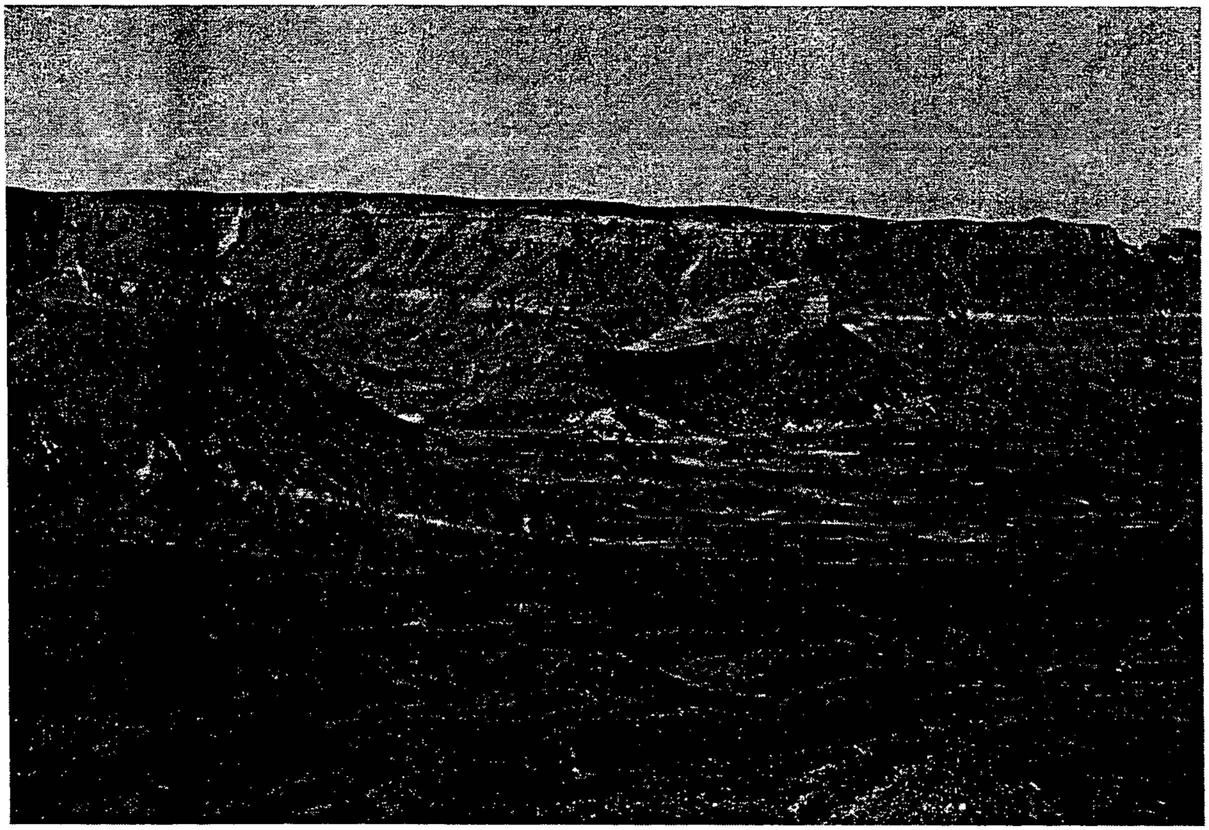


Figure PH.6 Photograph showing Lockhart Basin, Canyonlands area, Utah, a 50 km diameter dissolution collapse structure that has subsided into space created by dissolution of the Pennsylvanian Paradox salts 1,000 m below the center of the basin. The rocks on the top of the hill in the center are the same rocks that form the rim of the basin. The vertical subsidence at the center of the basin is 800 m. View is toward the south.

Deformation at the leading edge of the plate is compressional in style. In contrast, extension predominates at the trailing edge and within the moving plate. A detachment fault bounds the trailing edge of the plate, and the plate itself is commonly deformed internally by high-angle normal faults, producing a horst-graben complex with faults that strike perpendicular to the direction of sliding.

3 FIELD CRITERIA FOR IDENTIFYING AND DIFFERENTIATING

The faults associated with salt structures are indistinguishable in structural and geomorphic appearance from faults having other origins. Their geometries and the stress regimes that caused them are non-unique. The problem with discriminating between faults associated with salt structures and those of other origins involves placing them into their larger geologic context.

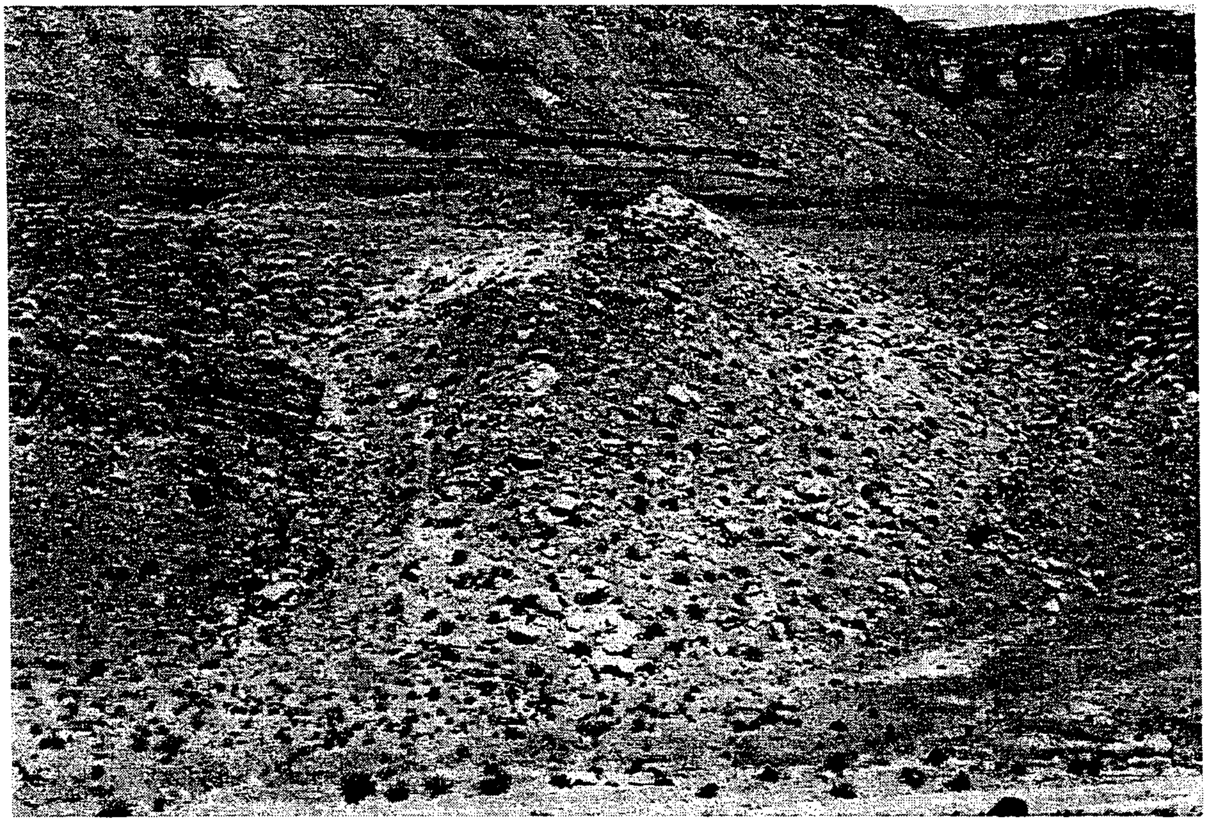


Figure PH.7 Outcrop of a breccia pipe in Lockhart Basin, Canyonlands area, Utah. The hill in the center is comprised of breccia which has collapsed downward within the pipe. The small draws facing the viewer on either side of the hill mark the wallrock-breccia contact. Notice that the wall rocks are folded inward toward the breccia core. The breccia core is 60 m across.

3.1 Stratigraphic Criteria

The presence of salt and related evaporites is unusually ascertained from the stratigraphic literature for a region of interest. Verification can be made from petroleum drilling records and downhole geophysical logs. Available seismic and gravity surveys will aid in the delineation of the salt bodies present. Field reconnaissance often will reveal unambiguous indicators, including evaporite exposures, caprock outcrops, diapirs, salt domes and salt anticlines. Salt dissolution indicators are equally useful indicators, including various types of collapse features such as sinkholes, breccia pipes, subsidence basins and collapsed salt anticlines.

Careful geologic mapping that emphasizes the planimetric distribution and subdivision of Quaternary deposits, delineation of caprock exposures, and trends and closures along fold axes will lead to identification of both the positive and negative structural elements (Coleman, 1983). Once identified, further geologic or geophysical investigations can focus on locating anticipated hidden faults and folds at depth.

Followup studies that are designed to discern thickening-thinning relationships and unconformities can be particularly valuable in assessing the stability of the overburden rocks in evaporite provinces. Cross sections based on measured sections from outcrops, drillhole logs and

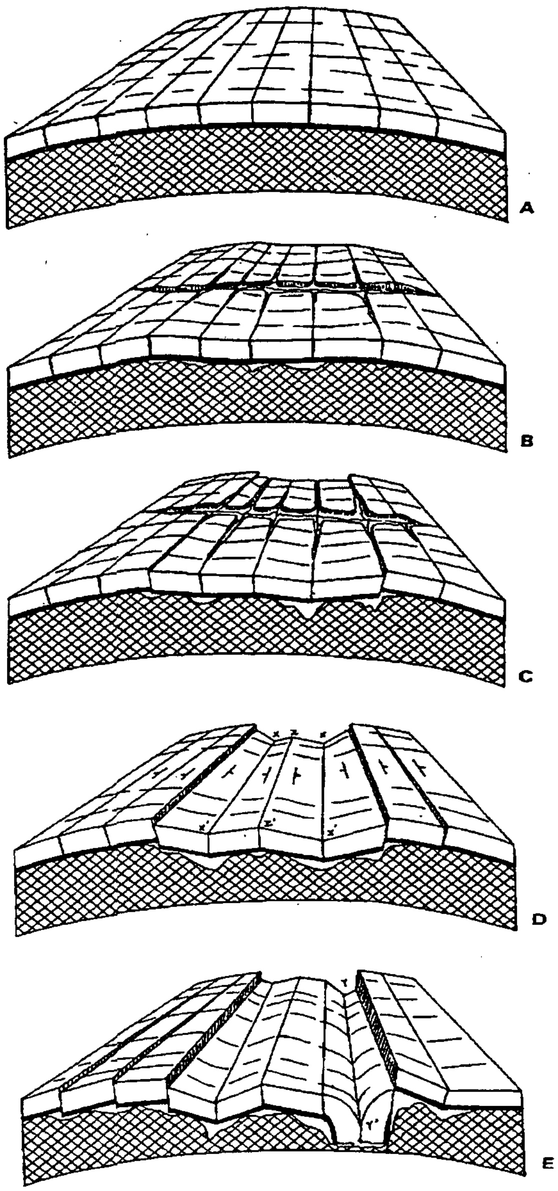


Figure PH.8 Progressive stages as a salt anticline collapses through dissolution of the slat in its core. Normal faults, graben, and chevron folds are primarily oriented parallel to the trend of the fold. Figure from Doelling (1983, figure 9). Permission to use this copyrighted material is granted by the Grand Junction Geological Society.

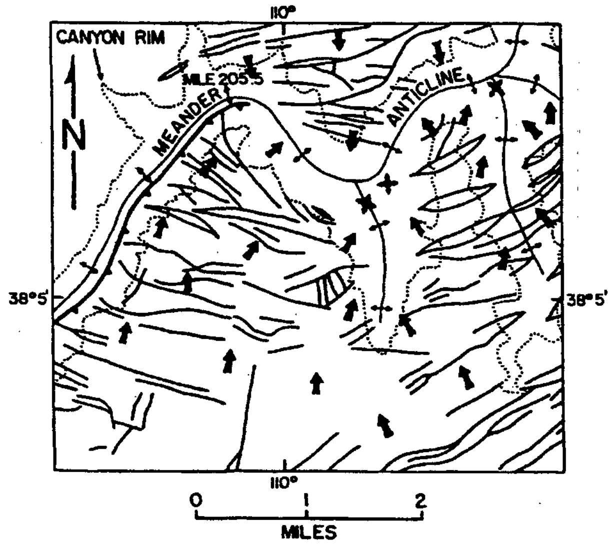


Figure PH.9 Structural map of the leading edge of a gravity glide plate, Canyonlands, Utah, showing the traces of faults in the Needles fault zone, the traces of anticlines being pushed up in front of the moving rocks, thrust faults where the leading edge of the plate is overriding stationary rocks, and the locations of three small diapirs which have pierced the valley floors. The arrows show the direction of transport within the plate. Figure from Geological Society of America Bulletin, P.W. Huntoon. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1982 Geological Society of America.

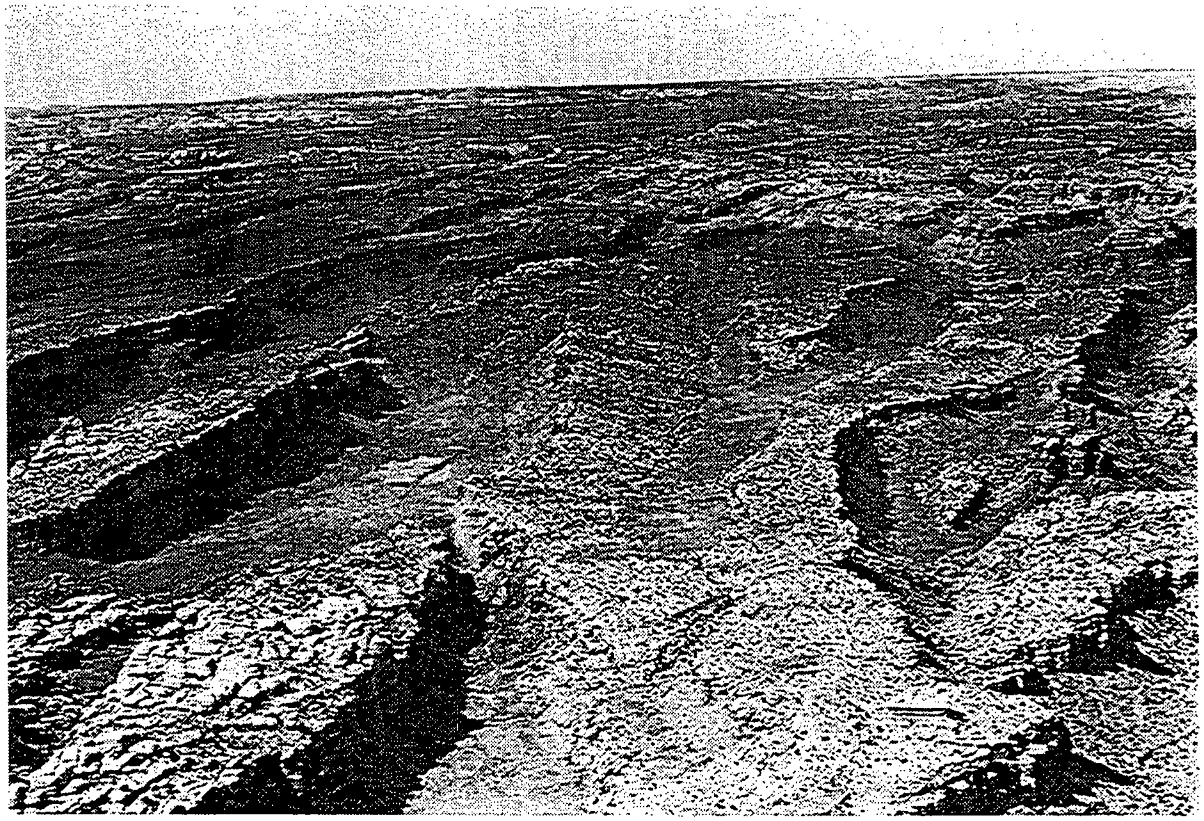


Figure PH.10 Horsts and grabens in the Needles fault zone, Canyonlands, Utah, an extensional fault zone formed by the disaggregation of a 700 m thick glide plate that is sliding toward the right above the Pennsylvanian Paradox salt, the upper surface of which serves as a décollement surface. Drainage is internal toward the graben left of center where the water is lost to an extensional figure on the valley floor. View is toward the south.

downhole geophysical logs supplemented by seismic profiles are required. Isopachous mapping of key units is particularly useful because it reveals where the underlying salt units have been thinned or thickened . A prime indicator of salt movement at depth is a record of long-term, non-uniform continuous or episodic uplift and/or subsidence, which can be read from the cross sections and isopach maps. Hidden faults in the stratigraphic section can be revealed directly in seismic profiles and from missing or repeated sections in well records. Abrupt lateral variations in the thickness of salt units can reveal the locations of potential faults, which can be further delineated using appropriate geophysical surveys.

Particular attention should be paid to the Quaternary deposits because the patterns of deposition of such deposits can be useful in identifying areas that are undergoing active deformation, both upward and downward. Consequently, all evidence for recent deposition should be inventoried. Cross sections and isopachous maps should be thoroughly analyzed for thickening patterns that can reveal the presence of sedimentation in depressions caused by subsidence associated with salt flowage or dissolution, or collapses arising from salt dissolution. In addition to measured sections and drillhole records, this type of work can be augmented by shallow seismic profiling, test drilling and even trenching to further characterize the

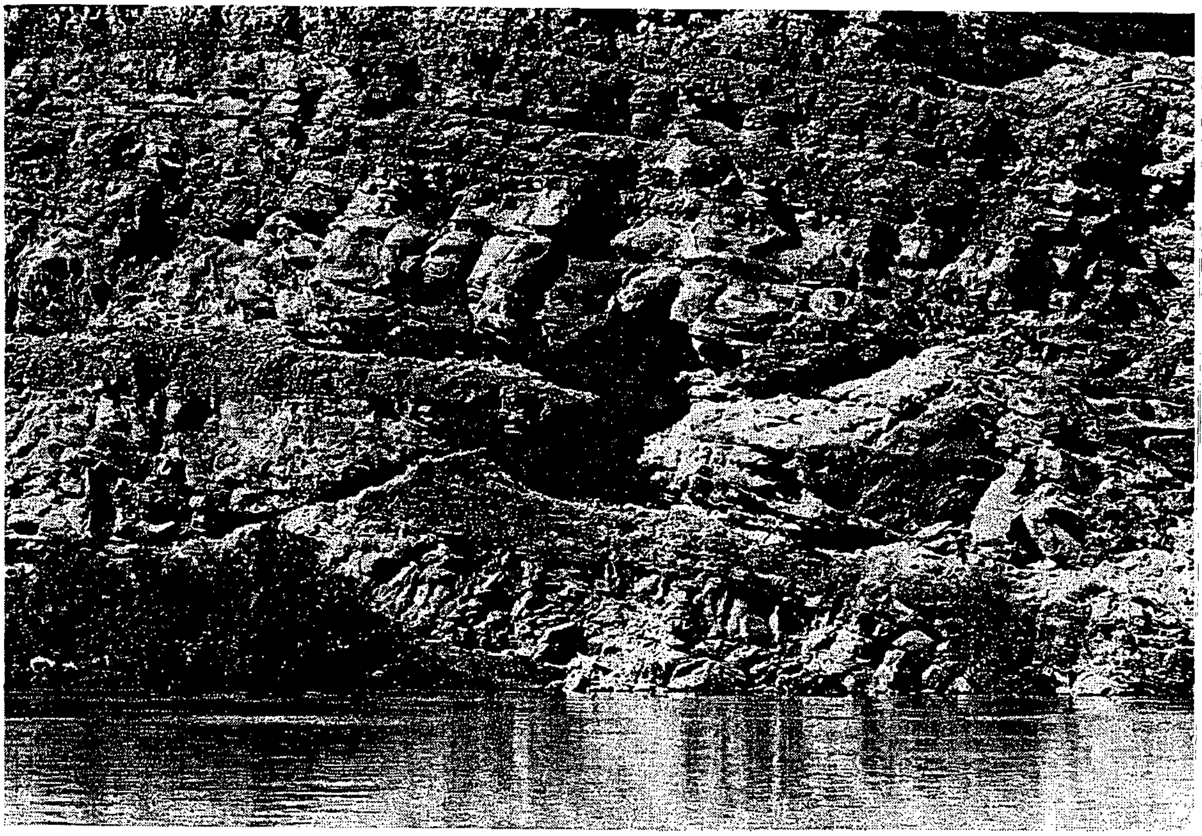


Figure PH.11 Thrust fault in an anticline adjacent to the Needles fault zone being pushed up by the leading edge of a glide plate, Canyonlands, Utah. Low angle conjugate fractures here reveal that the maximum principal stresses are horizontal, and oriented parallel to the direction of plate motion, which is toward left. View is toward the northeast.

thickening, thinning and unconformable relationships present.

3.2 Geomorphologic Criteria

The process of salt accumulation in the cores of salt structures competes with salt dissolution. The surface expression of the structures thus depends on which process dominates. For example, rising salt diapirs produce positive topography in the aridwest. Caprocks comprised of dissolution residues such as gypsum, limestone, shale and other clastics protrude above the land surface. In contrast, the same features are expressed in humid environments as circular or oval lakes which are collapse or subsidence features under which the salt cores of the structures are dissolving.

Particularly active rising anticlines, domes and diapirs will exhibit youthful, high-gradient, steep-walled drainages that appear to be out of character with other drainages in the surroundings. In extreme cases, the hill slopes will be at the angle of repose and barren of climax vegetation because spall rates exceed growth rates. An example of oversteepened slopes where the vegetation is continually sloughed off is the Onion Creek diapir along the Cache Valley salt anticline near Moab, Utah.

More subtle indicators of salt instability in the guise of variations in stream gradients can aid in pinpointing the locations of potential buried structural elements. Critical are abrupt variations in stream gradients read from stream profiles that reveal the presence of active fold axes associated with uplift or subsidence. Likewise, contrasts between reaches exhibiting sedimentation or scouring of valley deposits clearly reveal changes in stream gradients. Appropriate followup geologic and geophysical investigations can focus on documenting the presence of thickened salt bodies under the rising surface elements in order to demonstrate that they are in fact caused by salt flowage.

Salt dissolution and subsidence is commonly revealed by pseudo-karst topographic features. These include closed topographic basins of all sizes up to several kilometers across that are filling with recent, locally derived sediments. Semi-circular subsidence and collapse depressions ranging from several tens to hundreds of meters across are quite common above dissolving salt with a good example being the pock-marked surface of the Llano Estacado in the Texas Panhandle. Extensive sinkhole fields develop over some dissolving bedded salt with an excellent example being the extensive pseudo-karst sinkhole field near Snowflake, Arizona.

The presence of gravity glide structures is revealed by the same geomorphological indicators that are observed in association with large, low-gradient landslides. One key indicator is the presence of zones of ground failures and fault scarps that delineate the trailing end of the moving plate. The leading edge is demarcated by various types of lateral creep indicators such as oversteepened slopes on the leading edge of the mobile plate, ductile shear zones at the base of the plate, and tilted trees in wooded areas. The moving plate itself can be internally deformed by grabens and normal faults that trend roughly perpendicular to the direction of motion, along which the intervening blocks are commonly rotated. In the case of rotated blocks, most are back rotated on listric faults that dip in the direction of plate motion. Particular active glide plates will exhibit open fissures and sinkholes developed along buried extended fissures. Clastic material fills older extended fissures. Disrupted and/or reorganized surface drainage channels can be expected and careful analysis of datable deposits, benches, and other channel elements will reveal the recency and duration of the deformation. Sedimentation in grabens occurs where through-flowing drainage has been disrupted by faulting, or where graben subsidence exceed channel incision rates.

Salt tectonism and dissolution can lead to serious foundation stability problems including collapses, surface heaving, extensional fissuring, minor fault displacements and landsliding. Because salt tectonism is an ongoing process, these types of hazards usually can be identified through careful geomophologic mapping to identify preexisting surface manifestations of pervious activity. Typical indicators are the presence of sinkholes, fissures, arcuate fault scarps, slump structures, and chaotic tilting of adjacent bedrock blocks. Trenching can be useful in identifying these types of features in areas with ground cover and thick soil.

3.3 Structural Criteria

The primary means for discriminating between salt tectonic and other classes of structures is proving that the deformation is restricted to the salt and rocks overlying the salt bodies.

Regional structural mapping provides a valuable reconnaissance tool that can help discriminate between salt tectonic and deep-seated tectonic structures. The mapping should extend well beyond the known boundaries of the salt deposits, and all the tectonic structures, including faults and folds, must be plotted. Next, an overlay is made showing the areal extent of the salt deposits. Structures lying outside the boundaries of the salt deposits can be assumed to represent deep-seated, or at least non-salt, structures. Styles of deformation and trends are then analyzed to determine if there are patterns that can be correlated exclusively to salt tectonism, and others that imply the presence of non-salt structural features that should be segregated for separate analysis.

The most reliable non-intrusive means for determining the vertical character of tectonic structures is seismic profiling. Here the objective is to demonstrate that the profiled structure is restricted to the rocks above the salt. However, one ambiguity that commonly arises when using seismic techniques to discriminate between salt and deep-seated tectonic structures is the difficulty in sensing reflectors below salt bodies. A second difficulty is the fact that some salt flowage occurs in response to topographic irregularities on the surface upon which the salt was deposited. The pre-salt topography often resulted from tectonism, so it is then necessary to discriminate between the pre- and post-salt tectonic elements, and to determine if the older elements below the salt are still active.

In cases where information is particularly crucial, drilling can be employed to test for the presence of suspected buried faults. Potential drilling targets can be selected by means of traditional reconnaissance techniques, such as seismic profiling, facies or thickness changes evident from existing drillhole arrays, abrupt thickness changes evident on isopachous maps, high gradient zones on gravity maps, etc. Drilling targets can also be selected based on careful geologic mapping where surface features, such as drape folds, hint at the possible presence of buried faults. Drillholes can be used to reveal missing or repeated strata, occurrences that can be quantified through the use of appropriate downhole geophysical logging techniques.

The rocks on the crests of salt anticlines and domes are generally extended over the inflated salt cores. Open joint sets are commonplace, and grabens or normal faults occur along the crests of these folds with trends parallel to their axes. Dip slip predominates on the faults. Fault densities generally increase with depth. Shallow blind faults in this environment are sometimes revealed by drape folds which overlie them. The more open fractures in these zones - both faults and joints - are sometimes bleached as a consequence of ground water circulation through them.

3.4 Seismological Criteria

Salt tectonism is generally an ongoing geological process. Consequently opportunities exist to use micro seismic recording networks to discriminate salt structures from other types of structures having similar forms (Wong, Humphrey and Silva, 1987). The failures in the overburden rocks associated with active salt structures are usually caused by extension, consequently they generally have low seismicity. More important, hypocentral location data will reveal a pattern of earthquake foci that are predominantly restricted to the rocks down through and including the salt. Deep-seated crustal tectonism is indicated if the earthquakes occur at depths below the salt.

3.5 Geophysical Criteria

Gravity surveys have proven effective in delineating the presence of salt bodies owing to the low density of salt. Gravity surveys are appropriate as low-cost reconnaissance techniques for locating buried salt bodies such as diapirs, and for ascertaining the gross form and aerial extent of known salt structures.

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CRITERIA FOR DETERMINING THE SEISMIC SIGNIFICANCE OF SACKUNGEN AND OTHER SCARPLIKE LANDFORMS IN MOUNTAINOUS REGIONS

By

James P. McCalpin[1]

1 EXECUTIVE SUMMARY

The term "sackungen" (from the German verb "to sag") describes a family of landforms in mountainous areas that include crestal troughs, antislope scarps, and closed depressions (Figure JM.1). Although many authors have concluded that sackungen result from slow mass rock creep (e.g., Chigira, 1992), the more linear sackungen also resemble tectonic fault scarps. In addition, sackung-like landforms have formed or have been rejuvenated during historic earthquakes (Dramis and Sorriso-Valvo, 1983; Wallace, 1984; Cotton et al., 1990; Ponti and Wells, 1991; Nolan and Weber, 1992). Sackungen may therefore have formed by: (1) displacement on tectonic faults, (2) gravity failures caused by earthquake shaking, or (3) gravity failures unrelated to tectonics. Criteria for distinguishing among these three possible origins are geomorphic, structural, and stratigraphic. Geomorphic criteria are based on qualitative observations of scarp length, continuity, plan shape, and relation to topography. The ratio of scarp height:length may be a useful quantitative criterion. Fault scarps of tectonic origin typically exhibit small ($<10^4$) height:length ratios, whereas gravity scarps are commonly much shorter for a given height. In addition, gravity-related sackungen are typically short, discontinuous, arcuate, and occur in swarms of multiple parallel scarps, whereas fault scarps are longer, continuous, linear, and singular.

Structural criteria address the morphology of the shear zone beneath sackungen landforms and contemporary movement. Subsurface shear zones created by gravity creep are consistently asymmetrical, with a sharp upper contact and a transitional lower contact (Figure JM.2). Tectonic shear zones are commonly more symmetrical, especially if they formed at deep crustal levels and are now exposed after considerable erosion. Some very linear antislope scarps, originally interpreted as coseismic fault scarps, possess geodetically documented aseismic slip rates of up to 10 mm/yr (Bovis and Evans, 1995); such contemporary movement strongly suggests a gravity origin. Gravity shear zones also have predictable locations and orientations with respect to the present mountain ridge topography, relations that would be fortuitous for tectonic faults. For example, the toppling test of Goodman and Bray (1976) can indicate whether preexisting discontinuities ought to be failing under gravity stresses.

Finally, the stratigraphy and deformation of fine-grained Holocene sediments in sackung-related troughs and depressions can indicate whether the formation of these landforms was slow and gradual, or episodic. If formation was slow and gradual, then the features cannot be coseismic. Landforms created episodically could be of either tectonic or gravity origin, because even landslides

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may undergo episodic movement in response to climatic forcing, episodic basal erosion, or episodic loading.

In summary, determining a gravity- versus tectonic origin for sackung-like landforms (and their underlying shear zones) is best assessed by the application of multiple geomorphic, structural, and stratigraphic criteria. Use of these criteria requires geodetic, geomorphologic and microstratigraphic investigations (such as from trenching) similar to those utilized for the paleoseismic study of tectonic faults.

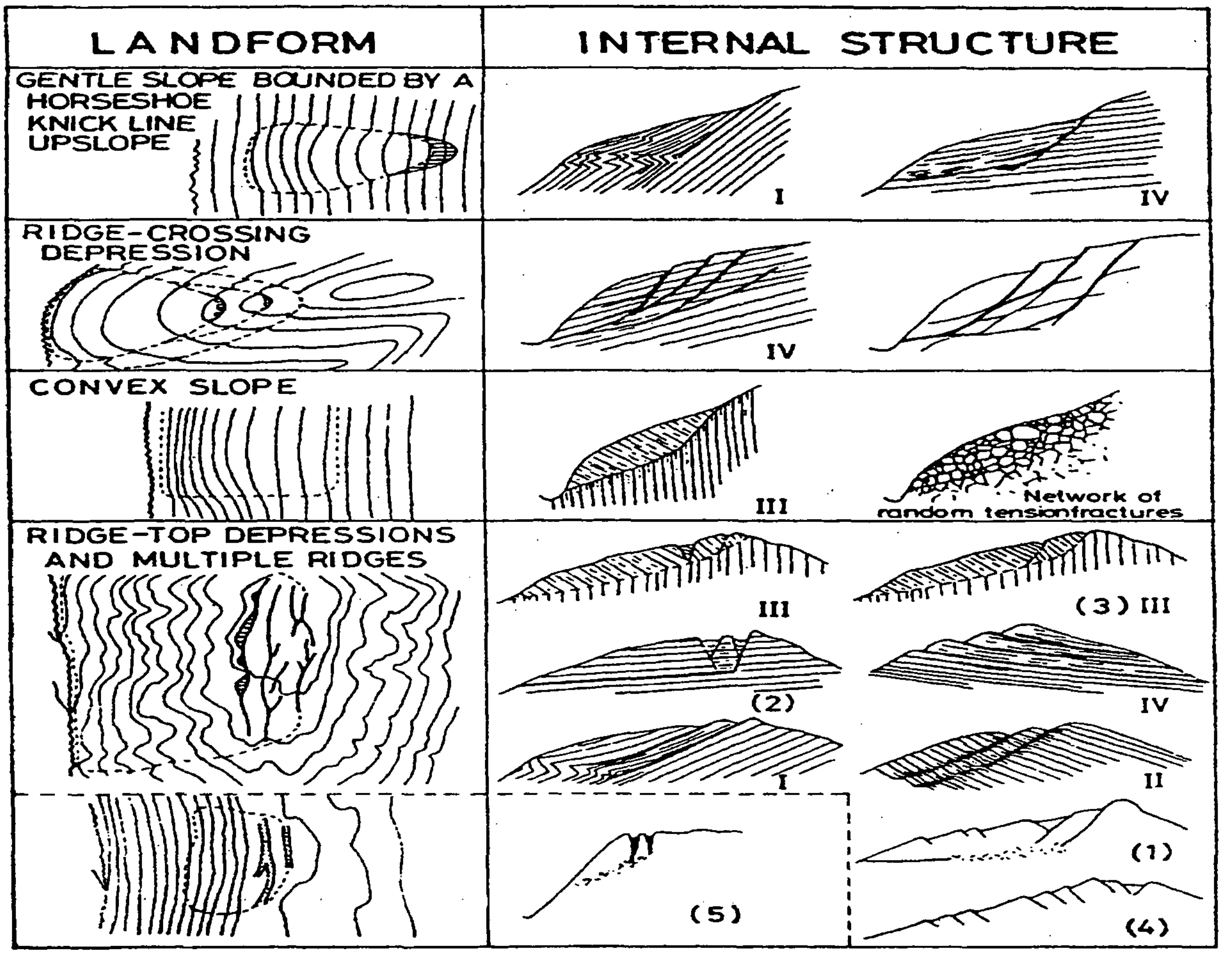

Figure JM.1 Typical landforms and their subsurface structures formed by mass rock creep. Roman numerals show the structure types defined by Chigira (1992). Reprinted from Engineering Geology, v. 32, M. Chigira, "Long-Term Gravitational Deformation of Rocks by Mass Rock Creep, p. 179, © 1992, with permission from Elsevier Science. Interpretive cross-sections at lower right (Arabic numerals) are derived from: (1) Mahr and Nemcok, 1977; (2) Ando et al., 1970; (3) Tabor, 1971; (4) Radbruch-Hall, 1978 and Shimuzu et al., 1980; (5) Jahn, 1964.

2 INTRODUCTION

Sackungen are antislope scarps, troughs (grabens), and closed depressions on the crests and flanks of mountain ridges that are the result of both tectonic and nontectonic processes. The term "sackung" (plural sackungen) is a descriptive landform term derived from the German word for sagging, and carries with it a connotation that the landform was created by slow creep. However, in almost all cases such an origin is merely inferred from morphology rather than demonstrated by geodetic or geologic studies. Thus, the term "sackungen" has been applied to a diverse assemblage of landforms, some of which may have been created suddenly rather than slowly. Throughout this paper I will use "sackungen" only as a descriptive term to identify landforms such as antislope scarps, ridge-crest troughs, and closed depressions, acknowledging that they may have formed either slowly or suddenly.

In general, geologists working in regions of moderate to high seismicity (e.g. Italy; Carpathian and Tatra Mountains (eastern Europe); Caucasus Mountains (southern Russia); New Zealand; western California) ascribe such scarps to either direct surface rupture or to seismic shaking (e.g., photographs in Khromovskikh, 1989; see also Zischinsky, 1966; Beck, 1968; Radbruch-Hall et al., 1976; Mahr, 1977). For example, Salvi and Nardi (1995) state "... strong ground shaking associated with earthquakes is one of the main triggering factors for the growth of 'sackung-like' features". This contention is partly based on the appearance of off-fault, antislope scarps in hilly terrain after moderate-to-large-magnitude earthquakes (e.g., Dramis and Sorriso-Valvo, 1983; Morton and Sadler, 1989; Morton et al., 1989; Cotton et al., 1990; Ponti and Wells, 1991; Nolan and Weber, 1992; Blumetti, 1995), and partly on a spatial association of sackungen with active fault traces. An example of the former from the USA is the Stillwater (Nevada) scarp, produced during earthquakes of 1915 and 1954 (Wallace, 1984). Similar prehistoric scarps occur in the Lost River Range fault zone, Idaho, upslope from the scarps of the 1983 Borah Peak (M 7.3) rupture (Crone et al., 1987, their Fig. 5). Geologists in less seismically active areas (Rocky Mountains, USA, and Coast Ranges, Canada) generally attribute similar scarps to nonseismic processes such as gravity creep and stress relaxation (crustal unloading) following deglaciation.

Antislope scarps and grabens have been the subject of two multi-investigator studies in the USA, and both studies resulted in equivocal conclusions concerning origin. In the Cascade Range of Washington, McCleary et al. (1978) concluded that some (but not all) of the antislope scarps and troughs probably represented tectonic reactivations of older faults, especially scarps found in close proximity to pre-Quaternary fault zones. Similar conclusions were adopted by other workers on this same project (Slemmons et al. [1977, 1978]; Woodward-Clyde Consultants [1978]; Dohrenwend et al. [1978]; Fugro Northwest [1979]; and Anderson et al. [1980]), that the more linear scarps near old faults were probably seismogenic, but that most of the shorter, more sinuous scarps at the lips of steep slopes were of gravitational origin.

A second series of studies in California also resulted in controversy. The extensive ground cracking caused by the 1989 Loma Prieta, California earthquake was coincident with antislope scarps, benches, and ridge top grabens at Summit Ridge, Santa Cruz County. Trenches excavated by Cotton et al. (1990) and Nolan and Weber (1992) across the 1989 ground cracks clearly showed that most of them had experienced prehistoric dip-slip displacement, which in turn had created the sackung-like landforms at the ridge crest. However, workers disagreed about the mechanism that created the ground fissures in

1989, and by analogy, the underlying prehistoric displacements. "The debate over the origin of the 1989 ground ruptures centers around whether the ridge-top fissures formed primarily as a result of tectonic deformation produced by the earthquake or whether they mostly reflect gravity-driven processes triggered by strong ground motions" (Ponti and Wells, 1991). Cotton et al. (1990) argue that fissures and faults resulted from normal bending-moment slip on bedding plane faults, caused by coseismic folding adjacent to the San Andreas fault. Because they considered the landforms to be the result of repeated secondary faulting caused by Loma Prieta-type earthquakes, they suggested that trenching studies in grabens could reconstruct a proxy paleoseismic history of the San Andreas fault. In contrast, Ponti and Wells (1991) concluded that the cumulative extensional and vertical displacements on cracks in 1989 were 35 times larger than could be explained by a bending-moment fault model. They concluded that approximately 90% of the ground displacement was due to downslope movement (landsliding) caused by prolonged seismic shaking.

Similar controversies have arisen in Canada, where occasional long, linear antislope scarps occur in areas with more abundant short, arcuate bedrock scarps. For example, Eisbacher (1983) identified the prominent linear antislope scarp on Mt. Currie (British Columbia) as a young fault scarp. In contrast, Evans (1987) and Bovis and Evans (1995) documented that: (1) the joints beneath the scarp were predicted to fail by the toppling test of Goodman and Bray (1976), and (2) between 1987 and 1991 up to 80 mm of horizontal displacement and 40 mm of vertical displacement occurred across the scarp, in the absence of any earthquakes. They thus concluded the scarp, despite its linearity, was a gravitational failure. The Hell Creek scarp (British Columbia) has been interpreted as a Holocene tectonic reactivation of an older fault (Psutka, 1995), based mainly on evidence for episodic displacement (observed in trenches) with a lateral component (inferred from weak geomorphic evidence). In contrast, Clague and Evans (1994) consider the scarps to reflect nonseismic gravity failure, similar to many other scarps in the area.

The purpose of this paper is to: (1) summarize the identifying characteristics of sackungen, and (2) to provide criteria that enable the differentiation between sackungen-like landforms that are tectonic versus nontectonic in origin. This summary paper is based on a compilation of literature on sackungen and other scarplike landforms throughout the world, and on a synthesis of field reconnaissance and detailed field studies in western North America with the worldwide data base.

3 DESCRIPTION OF SACKUNGEN AND THE SACKUNG PROCESS

3.1 Terminology

Zischinsky (1966, 1969) first proposed the term "sackung" for the surface manifestations of deep-seated rock creep in foliated bedrock of the Alps. In this paper the term "sackung" refers to the process of deep-seated sagging, whereas "sackungen" is used as a generic term to describe any landforms such as antislope scarps or ridge-crest depressions, regardless of origin. Other workers have referred to the same slow rock deformation process as mass rock creep (MRC; Radbruch-Hall, 1978), depth creep (Ter-Stepanian, 1966), deep-seated creep (Nemcok, 1972), deep-seated continuous creep (Hutchinson, 1988), bedrock flow (Varnes, 1978), or gravitational spreading (Radbruch-Hall et al., 1977; Varnes et al., 1989). Unfortunately, these processes have usually been inferred to explain landforms whose exact mode and rate of formation is unknown, and which could be (in part) created by rapid displacements accompanying surface fault rupture or earthquake shaking.

3.2 Characteristics and Geometry

Sackungen are landforms, so inferences concerning their origin have traditionally been based purely on geomorphic evidence. This evidence emphasizes that sackungen must have a deep-seated rock-failure origin because they occur in topographic positions (high on ridge flanks, at ridge crests) where an erosional origin is extremely unlikely. The most common geomorphic features can be divided into four categories: downhill-facing scarps, double-crested ridges, uphill-facing scarps, and erosional notches on ridge axes (Figure JM.1). Double-crested ridges (the "Doppelgrat" of Zischinsky, 1969) are a classic sackung landform. The axial depression typical of Doppelgrat is difficult to explain by an erosional process, because streams are unlikely to flow down the crest of a ridge and the closed depressions could not have been excavated by running water. Uphill-facing scarps and sidehill benches (degraded scarps?) are probably the most common sackung landform (e.g., Varnes et al., 1989). Along strike, sackungen may grade from antislope scarps to benches, and grabens may grade into irregular closed depressions; along-strike changes in height and morphology are common.

An inventory of published sackung scarp dimensions (McCleary et al., 1978, Figure JM.2) yields these typical ranges: scarp length, 15-300 m; scarp height, 1-9 m; slope height, 400-1200 m; slope gradient, 25°-50°. However, Salvi and Nardi (1995) interpret a trough 100 m deep, 700 m wide, 9 km long in the Apennines (Italy) as an earthquake-induced sackung. Almost all of these landforms are found at or near the crests of slopes, in the zone of tensional failure. In contrast, very few distinctive landforms have been observed on the lower slopes, except where authors have postulated that lower slopes have been "oversteepened" or bulged outward by compressive forces at the "toe" of a creeping rock mass.

Two opposing hypotheses have been proposed for the subsurface geometry of sackungen in massive competent rocks. One, held by Zischinsky (1969) and other European and American workers, proposes that "a well-defined slide plane near the headscarp passes downward into a broader zone of rock creep. Consequently the lower portion of this type of failure simply bulges out into the valley" (Morton and Sadler, 1989). The slide plane may dip either into or out of the slope. Such slow, deep-seated failure results in "half-a-landslide" morphology (Morton and Sadler, 1989), with well-developed tensional features near the head, but often with no recognizable evidence of medial landslide features or compressional morphology downslope from the scarp. Radbruch-Hall (1978) claims that rock creep can extend to depths of several hundred meters. However, there are few locations where the depth or shape of the failure plane can be measured with certainty, making the "half-a-landslide" hypothesis difficult to directly test.

The second hypothesis is that sackungen are shallow surface manifestations of toppling and flexural slip along discontinuities that dip steeply into a mountain mass, but which do not penetrate to any great depth (Jahn, 1964, his Fig. 9; Beck, 1968). Bovis (1982) termed this process "flexural toppling" and cited model studies (Barton, 1971) and studies in quarries (Goodman and Bray, 1976) as support for this non-penetrative mode of extensional deformation. During flexural toppling outward rotation of blocks and dilation of sackung cracks lead to attenuation of movement with time, which Bovis (1982) compared to strain-hardening in granular materials. Given the steep dip inferred in this model, the sackungen could possibly

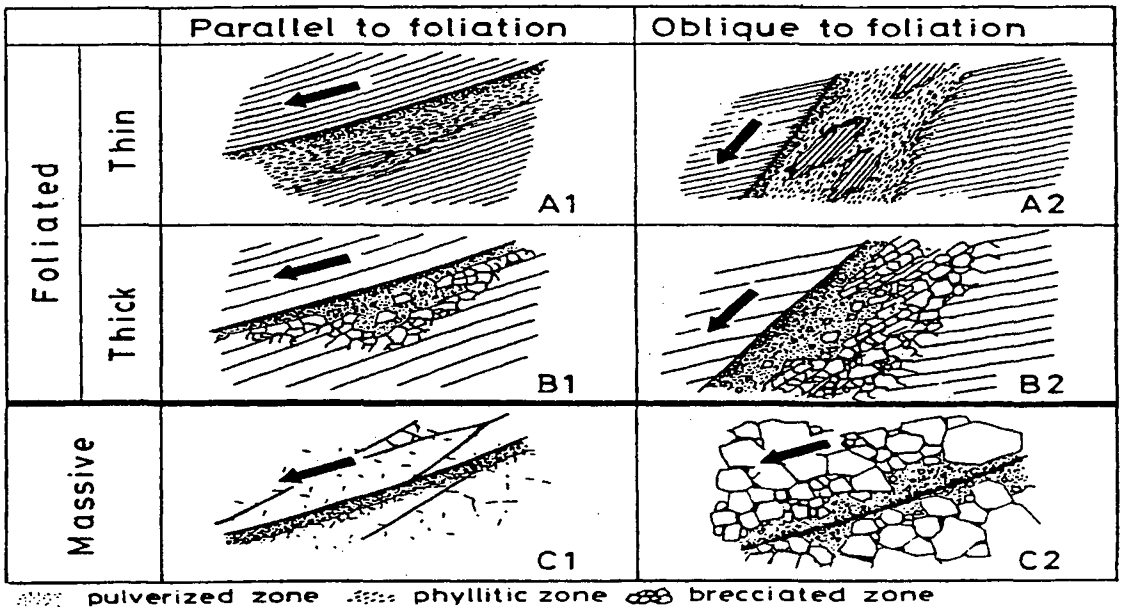

Figure JM.2 Schematic sketches showing mesoscopic features of fault (shear) zones formed by mass rock creep. Dotted areas show pulverized zones, dashed areas show phyllitic zones. The fault in C1 is generated from shear fractures, whereas the fault in C2 is generated from random tension fractures. The fault in C2 has its sliding surface in the middle of the shear zone, whereas the other five fault types have sliding surfaces at their upper edges. Reprinted from Engineering Geology, v. 32, M. Chigira, "Long-Term Gravitational Deformation of Rocks by Mass Rock Creep, p. 174, © 1992, with permission from Elsevier Science.

connect to seismogenic faults at depth, although the vertical separation of the ground surface along sackung scarps would not necessarily bear any relation to fault movement at depth.

The most detailed study of subsurface deformation features (folds, faults) associated with mass rock creep is that of Chigira (1992). Chigira was mainly concerned with characterizing the micro- and meso-scale characteristics of the causative shear zones that underlay sackungen at depths of 10s to 100s of meters. Chigira (1992) describes the discrete "fault" zones that underlie areas of mass rock creep as "a pulverized zone with fault gouge and a phyllitic or brecciated zone (Figure JM.2). In a densely foliated rock (Figure JM.2A1, 2A2), the phyllitic zone is formed by microscopic slip along foliations. In a sparsely foliated rock (Figure JM.2B1, 2B2), the brecciated zone is formed by random crushing. In massive rocks (Figure JM.2C1, 2C2), a brecciated zone is formed through networks of tension fractures, but not if it is formed through connection of shear fractures. The pulverized zone, which usually forms only in the upper part of the shear zone, is formed through grinding by a downsliding block on the fault." He assumed that the features he described were of gravitational rather than tectonic origin, and did not address the issue of possible shaking-induced gravitational movement. The details of Chigira's study are presented in Sec. 4.2.

3.3 Process

Varnes et al. (1989) distinguish three types of sackung: (1) spreading of rigid rocks overlying soft rocks (Radbruch-Hall, 1978; Radbruch-Hall et al., 1976), (2) sagging and bending of foliated phyllites, schists, and gneisses ("true Sackung" of Zischinsky, 1969), and (3) differential displacements in hard but fractured crystalline igneous rocks. Sackungen have been observed in almost all rock types, including phyllite and schist (Jahn, 1964; Zischinsky, 1969; Nemcok, 1972; McCleary et al., 1978; Ertec Northwest, 1981; Morton and Sadler, 1989; Clague and Evans, 1994), slate (Goodman and Bray, 1976), high-grade gneisses and intrusive rocks (Radbruch-Hall et al., 1976, 1977; West, 1978; Varnes et al., 1989, 1990; Bovis and Evans, 1995), volcanic rocks (Tabor, 1971; Bovis, 1982; Beget, 1985), and massive sedimentary rocks (Beck, 1968; Radbruch-Hall, 1978).

The stress field that produces sackung may have five possible origins: (1) ice wedging (the original explanation for European sackung; Zischinsky, 1969), (2) gravity forces that produce slow deformation to the point of instability of a rock mass, (3) stored forces resulting from prior loading conditions (e.g., glaciation) that produce sporadic deformation as strain is recovered, (4) seismic shaking that induces lateral spreading and differential settlement of rock masses, and (5) tectonic displacement connected to deep-seated seismogenic faults (Ertec Northwest, 1981).

The ice wedging theory has largely been abandoned because it has been observed that sackung spreading extends to great depths and occurs in temperate climates. Recent modeling of stresses in long symmetric ridges (Savage et al., 1985; Savage and Swolfs, 1986; Savage and Varnes, 1987; Pan and Amadei, 1994; Pan et al., 1994) provides theoretical support for a gravitational origin, especially where weak foliated rocks are present. Proponents of a stress relaxation origin point out that sackungen are common in areas of high relief, especially where valley walls were "oversteepened" by Pleistocene valley glaciers. Augustinus (1995) suggests that most glacially "oversteepened" slopes are actually in strength equilibrium (Selby, 1993) and, while subject to slow strain, will not fail catastrophically. Many authors (Beck, 1968; Radbruch-Hall, 1978; Bovis, 1982) suggest a causal relationship between the retreat of a valley glacier that once buttressed a steep slope, and subsequent sagging and bulging of the slope. Tabor (1971) noted that sackungen are widespread only where ridges rise more than 1000 m above glaciated valleys. Supporters of an earthquake-shaking origin, working in active seismic areas, have noted sackungen on lower-relief slopes, and have ascribed the spreading to either general earthquake shaking and settling (Beck, 1968; Solonenko, 1977; Clague, 1979, 1980; Ponti and Wells, 1991; Nolan and Weber, 1992), concentration of rock shattering on ridge crests by topographic amplification (Morton et al., 1989), or to surface fault rupture (e.g. Cotton, 1945; Cotton et al., 1990; Johnson and Fleming, 1993). The sackungen landforms cited as evidence for the various theories described above all look remarkably similar, perhaps because they reflect the geometry of deep-seated failure regardless of how it was initiated.

4 CRITERIA AND METHODS TO DIFFERENTIATE TECTONIC VERSUS NONTECTONIC DEFORMATION

Possible methods for differentiating gravity-created sackungen and related structures from tectonic landforms and faults fall into three general categories: geomorphologic, structural, and stratigraphic. Psutka (1995) compiled the first known list of geologic/geomorphic criteria to differentiate tectonic from nontectonic scarps in steep terrain. He compared the Hell Creek fault (which displays an antislope scarp suspected to be a Quaternary fault scarp) to the more abundant gravity-related scarps (true sackungen) in southwestern British Columbia (Figure JM.1). His observations are qualitative, but do provide a starting point for defining criteria that would separate tectonic and nontectonic sackungen. In the following sections I group criteria into geologic and stratigraphic criteria (Psutka's parameters 6 and 7), geomorphologic criteria (Psutka's parameters 1-5), and structural criteria (including geodetic measurements).

Table JM.1 Parameters Used by Psutka (1995) to Differentiate Tectonic From Nontectonic Sackungen

Scarp Parameter	Hell Creek Fault	Gravity Sackungen
1. Length	kilometer scale	10s to 100s of meters
2. Continuity	continuous	discontinuous
3. Number of scarps	single scarp	multiple clustered scarps
4. Plan shape[1]	linear	arcuate
5. Relation to topography	crosses a ridge crest	restricted to one side of, or adjacent to, ridge crest
6. Displacement history	episodic	lateral side scarps
7. Style of deformation	compressional structures	extensional structures

[1] This criterion was also used by the Technical Advisory Group (1991) to differentiate "structurally controlled ground cracks" from "non-structurally controlled landslide cracks" resulting from the 1989 Loma Prieta, California earthquake.

4.1 Geologic and Stratigraphic Criteria

Stratigraphic criteria hold considerable promise of differentiating tectonic versus nontectonic scarps. In this section I consider both the stratigraphy and small-scale deformation features in unconsolidated sediments deposited in sackungen sediment traps. To date, little work has been performed on trenching sackungen and examining the deformed strata in sackungen troughs. The writer knows of only three trench investigations across sackungen: (1) trenching of scarps along the Straight Creek Fault, Washington (McCleary et al., 1978), (2) trenching of a scarp at Aspen Highlands Ski Area, Colorado (McCalpin and Irvine, 1995), and (3) multiple trenches across the Hells Creek fault and subsidiary scarps, British Columbia (Psutka, 1995).

Most of these trenches displayed a moderately steep (45-90°) shear zone across which the vertical, scarp-producing movement had occurred. The trench by McCleary et al. (1978, their Fig. 21) shows a subvertical fault plane beneath the lower scarp face. (Natural exposures of sackung trough fill in the North Cascades, Washington, also show subvertical fault planes [Beget, 1985]). The Aspen trench (McCalpin and Irvine, 1995) revealed both a steeply-dipping shear zone and abundant evidence of fracture dilation, perhaps the result of toppling. The Hells Creek trenches displayed a wide variety of deformation styles, with fault dips ranging from 45°-85°, and fault geometry ranging from planar gouge-covered zones to irregular fault planes in brittle granitic rocks.

Most trenches expose fine-grained sediments that have accumulated in the sackung trough, and these ductile sediments are now tilted sharply upwards near the projection of the fault plane. There is generally little coarse-grained sediment in the trough that may have been deposited from the scarp face (i.e., scarp-derived colluvium), such as is often found along normal fault scarps. The lack of a recognizable colluvial wedge stratigraphy in most sackung trenches suggests that the mode of scarp formation is slow, continuous slip on the underlying fault plane, rather than discrete episodes of large displacement followed by quiescence. However, at Hells Creek the trench that displayed the steepest fault plane also showed a colluvial wedge stratigraphy and buried soils, indicative of episodic movement.

In order to assess whether sackungen are tectonic or nontectonic, we first have to assess whether movement on the sackung shear plane has been continuous or episodic. A stratigraphic record of slow continuous deformation precludes the occurrence of coseismic surface-rupture at a site, although continuous movement could result from tectonic creep. In contrast, episodic movement could be due to normal landslide processes, to gravity failure caused by seismic shaking, or to direct surface faulting. In the following criteria I address both the episodicity of displacement and the tectonic setting of the sackungen.

4.1.1 CRITERION: Evidence of Continuous Deformation of Sediments Suggests a Nonseismic Origin

Given the general absence of scarp-derived sediments in trenches excavated across sackungen, the deformation history can be reconstructed based on the dips of folded sediments and soils. Four idealized geometries can be envisioned, which are combinations of continuous versus episodic deformation, and continuous versus episodic deposition in the trough (Figure JM.3).

If deposition in the trough is continuous, then no buried soils or other hiatuses will be present in the trough fill. Continuous displacement on the sackung shear will then result in progressively greater folding of strata with increasing depth (Figure JM.3a). If displacement has been episodic, then packages of strata will be folded as units, with the oldest packages the most folded (Figure JM.3b). Individual strata within each package will be folded the same amount, and each package will be separated from the next by an angular unconformity.

If deposition in the trough is discontinuous, buried soils probably will be present (e.g., McCalpin and Irvine, 1995). These soils reflect time periods during which little or no additional deposition occurred in the trough. If displacement has been continuous, but sediments have been deposited in pulses separated by long periods of nondeposition, then each stratal package between soils will be deformed (folded) more or less uniformly (Figure JM.3c), with angular unconformities bounding the packages (as in Figure JM.3b). However, in this scenario the angular unconformities will always coincide with the tops of the soil horizons. If displacement has been episodic, and not perfectly in-phase with episodic deposition, then unconformity-bounded packages of strata will also occur, but soils may be found in any stratigraphic position within a package (Figure JM.3d). In other words, if hiatuses in deposition are not contemporaneous with hiatuses in displacement (as they are where most fault-zone sediments are scarp-derived), buried soils will not necessarily occur at angular unconformities.

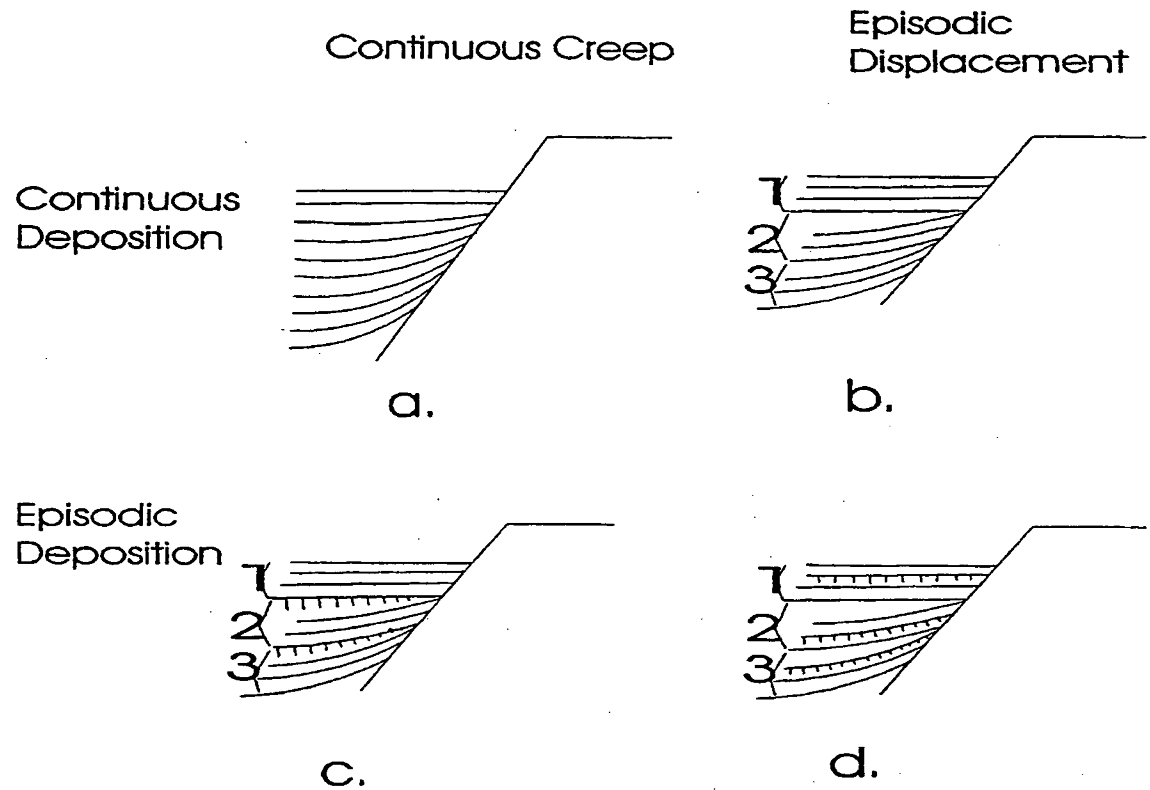

Figure JM.3 Hypothetical cross-sections through a sediment-filled trough (at left) adjacent to a sackung scarp (at upper right). In the trough, thin lines indicate bedding, short vertical lines indicate soils. (a) Continuous creep and continuous deposition yield increasing folding (drag) with depth; no soils are present. (b) Episodic displacement and continuous deposition yield three packages of strata bounded by angular unconformities; no soils are present. The angular unconformities are "event horizons" in the terminology used by paleoseismologists. (c) Continuous creep and episodic deposition yield three unconformity-bounded packages of strata. Each package is topped by a soil, the upper parts of which have been eroded nearest the sackung fault plane. The upper contact of each soil is an event horizon. (d) Episodic displacement and episodic deposition yield discrete unconformity-bounded packages of strata, but soils may be found at any position within the stratigraphic sequence. In this scenario, the angular unconformities are event horizons, but do not coincide with buried soils.

4.1.2 CRITERION: If Deformation Events on a Sackung are Contemporaneous With Other Area Paleoseismic Features, They May Be Coseismic

The timing of episodic sackung movements can provide suggestive evidence for tectonic versus nontectonic origin. This criterion assumes that there are other dated landforms or deformation features in the area that are clearly known to be coseismic. If sackungen can be demonstrated to have formed in discrete displacement episodes, and these episodes are contemporaneous with other independently-recognized and dated paleoseismic features, then a common origin is suggested. Stronger evidence would be an absence of sackung deformation episodes that were not contemporaneous with other paleoseismic evidence. For example, a wide range of sackungen ages might be expected if they were produced by frequent climatic triggers or progressive stress relaxation, rather than rare large earthquakes. A recognizable age progression among the sackungen would thus either suggest a nontectonic cause, or require very frequent earthquakes. For example, Bovis (1982, his Fig. 11) constructed a series of diagrams showing retrogressive (upslope-younging) sackung movement in British Columbia. McCleary et al. (1978) and Ertec Northwest (1981) described a similar upslope-younging of sackungen in the North Cascade Range, whereas Beget (1985) documents scarps that cross-cut each other on the same ridge crest. Consistent upslope-younging is unlikely to result from rare instances of coseismic displacement.

4.1.3 CRITERION: If Sackungen Overlie a Steeply-Dipping Crustal Fault Zone, Which Has A Net Displacement Much Larger Than Scarp Height, The Possibility Exists That The Fault is Active

Merely proving contemporaneity between episodes of sackung movement and paleoearthquakes (previous criterion) does not prove that sackungen are primary tectonic surface ruptures; the sackungen could be secondary gravitational features formed in response to coseismic shaking ("seismogravitational" landforms as defined by Solonenko, 1977). However, the dip and amount of displacement on the sackung failure plane (in relation to modern scarp height) provide further criteria concerning the seismic potential of the failure plane. First, the failure planes of classical sackungen (Zischinsky, 1969) and creeping rock masses (Chigira, 1992) tend to flatten with depth, and are often inferred to daylight on the lower slopes of mountain ridges. Seismogenic faults, in contrast, cannot daylight on lower mountain slopes, but must maintain steep enough dips in the subsurface to reach seismogenic depths (i.e, several km). Second, a seismogenic fault with a long history of coseismic displacements will display a cumulative displacement greater than the height of today's sackung scarps. While it is possible that sackungen may overlie a "new" fault zone that has only experienced the single movement that created the surface scarps, it is more likely that a seismogenic fault zone would have a history of displacement resulting in considerable tectonic offset of strata.

A complication to this criterion would occur if gravitational spreading (either continuous or episodic) occurred along an old, inactive fault or shear zone (such as at Hell Creek). In this case, the underlying fault would possess a large net displacement relative to present scarp height, and the scarp atop the fault might be very linear if the underlying old fault zone was linear. If in addition, trenches across the scarp suggested an episodic history of movement, the temptation would be strong to conclude (as did Psutka, 1995) that the sackungen represented Holocene surface rupture. However, the same surface deformation could be created by episodic gravitational failures (due to climatic events or earthquake shaking) that occupied the upper (steeply-dipping) part of an old fault zone. The key criterion in this case is whether the Holocene slip occupied only the upper part of an old fault zone (as would occur in flexural toppling or mass rock creep) or the entire fault plane down to seismogenic depths. Unfortunately, this distinction would be very difficult to make.

4.2 Geomorphologic Criteria

Sackung spreading and faulting can produce similar scarp-like landforms. However, several suggestive criteria can be proposed to differentiate the two groups.

4.2.1 CRITERION: General Geomorphology of Antislope Scarps

Psutka (1995) compared four geomorphic parameters (Figure JM.1, parameters 1-4) between the Hell Creek fault and the more abundant gravity-related scarps in southwestern British Columbia. These parameters suggest that gravity-related sackungen are short, discontinuous, arcuate, and occur in swarms (Figure JM.4); tectonic scarps tend to be longer, continuous, singular, and straight (Figure JM.5). However, gravity failures that reoccupied an old plane of crustal weakness would resemble tectonic scarps. Parameter 5 indicates that a single gravity-related sackungen will be restricted to a single part of a ridge, and would not indiscriminately cross the ridge crest. These observations are qualitative, but do provide a starting point for defining criteria that might separate tectonic and nontectonic sackungen. A semi-quantitative treatment of parameter 1 is given below; given sufficient data, all of Psutka's geomorphic parameters could be subjected to a similar analysis.

4.2.2 CRITERION: Scarp Height:Length Ratio

On tectonic faults, high (>5 m) fault scarps commonly occur on faults with long (several km to 10s of km) traces. On most sackungen, however, high scarps are typically very short, often only 10s or 100s of meters (e.g., Figure JM.4). In this regard sackungen resemble landslide headscarps. Many workers (e.g. Bonilla et al., 1984; Wells and Coppersmith, 1994) have published empirical relationships between coseismic displacement and rupture length (Figure JM.6). On Figure JM.6 I have plotted the dimensions of the Hell Creek scarp (average height 2.5 m, length 5.8 km), the Mount Currie scarp (average height 5-10 m, length 1.6 km), and a box representing sackungen such as shown in Figure JM.4, with average heights of 5-10 m but lengths of < 1 km. These three points plot far above and to the left of the 95% confidence interval bounding tectonic scarps. The typical dimensions of sackungen cited in Sec. 3.2 (height 1-9 m, length 15-300 m) define height:length ratios that plot off Figure JM.6 to the upper left. These disparities suggest that the scarp height:length ratio might serve as a criterion for tectonic versus nontectonic scarps.

However, the three sackungen scarps plotted on Figure JM.6 may have been created by multiple displacement events, in which case a comparison with single-event tectonic scarps is misleading. Psutka (1995) interpreted the Hell Creek scarp to be the result of three displacement events. If we assume that each event resulted in roughly equal displacement (0.83 m), then the single-event displacement on the Hell Creek scarp plots close to the upper limit of tectonic scarps. There are no data on the number of displacement events at Mount Currie, but in order for that 5-10 m-high, 1.6 km-long scarp to plot close to tectonic scarps, the Mount Currie scarp would have to have been created by 16-32 displacements of roughly 0.3 m each. Although such an origin is possible, it would require very frequent surface faulting in an area that is not highly seismic. In addition, a repeated history of small-displacement events would be discernible in trench stratigraphy (see Sec. 4.1).

4.3 Structural Criteria

Structural criteria involve the character of the shear zone that underlies sackungen. Chigira (1992) proposes several criteria to distinguish deformation resulting from mass rock creep from tectonic deformation, based on detailed work in Japan in igneous, metamorphic, and sedimentary rocks.

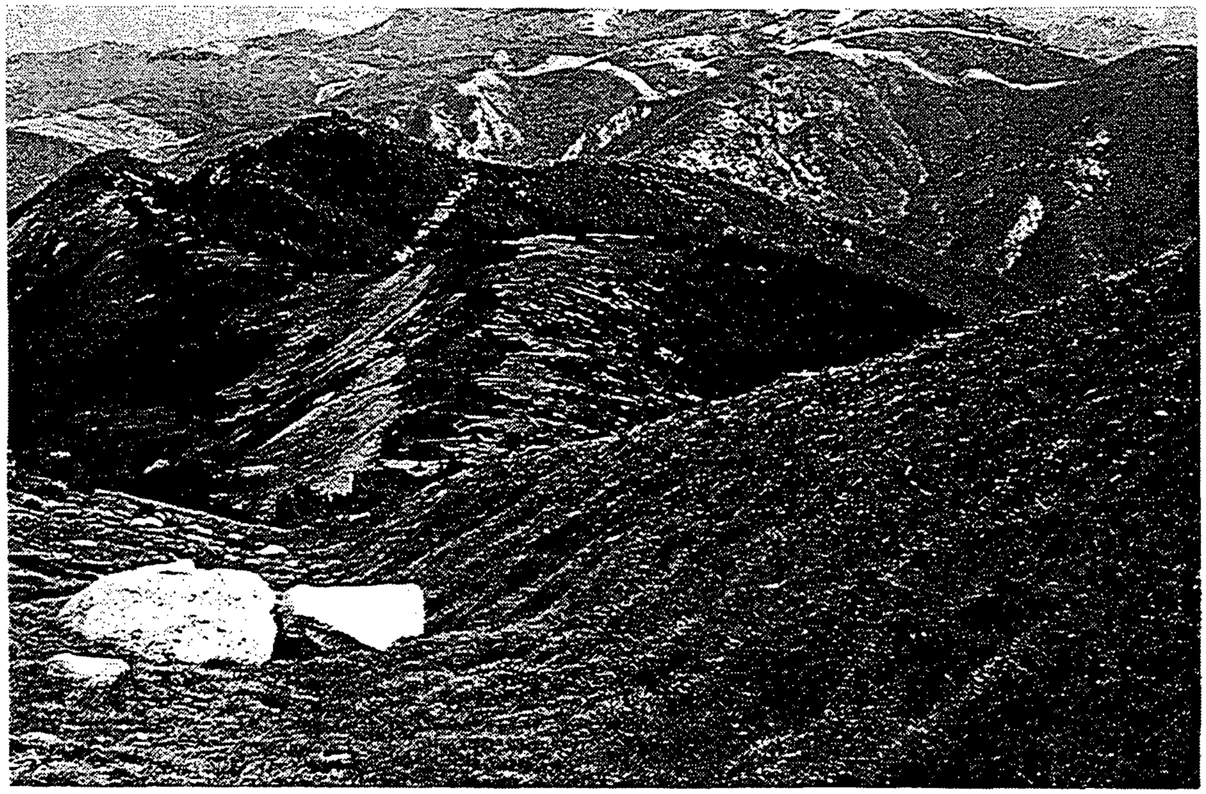

Figure JM.4A

Figure JM.4 The Hell Creek scarp, southwestern British Columbia, Canada. (A) The scarp in right foreground is 3.5 m high. In the middleground the scarp defines a talus-covered slope facing to the left, which probably represents a landslide headscarp that has occupied the Hell Creek fault zone. In the middle distance the scarp appears as a thin light band of unforested terrain surrounded by forest, and trends toward a saddle in the ridge. (B) Close-up of the Hell Creek fault (unforested strip in middle distance of part A), with Psutka's (1995) Trench 1 in foreground; note person with video camera for scale. Scarp in this reach is 2.5 m high.

Figure JM.4, continued

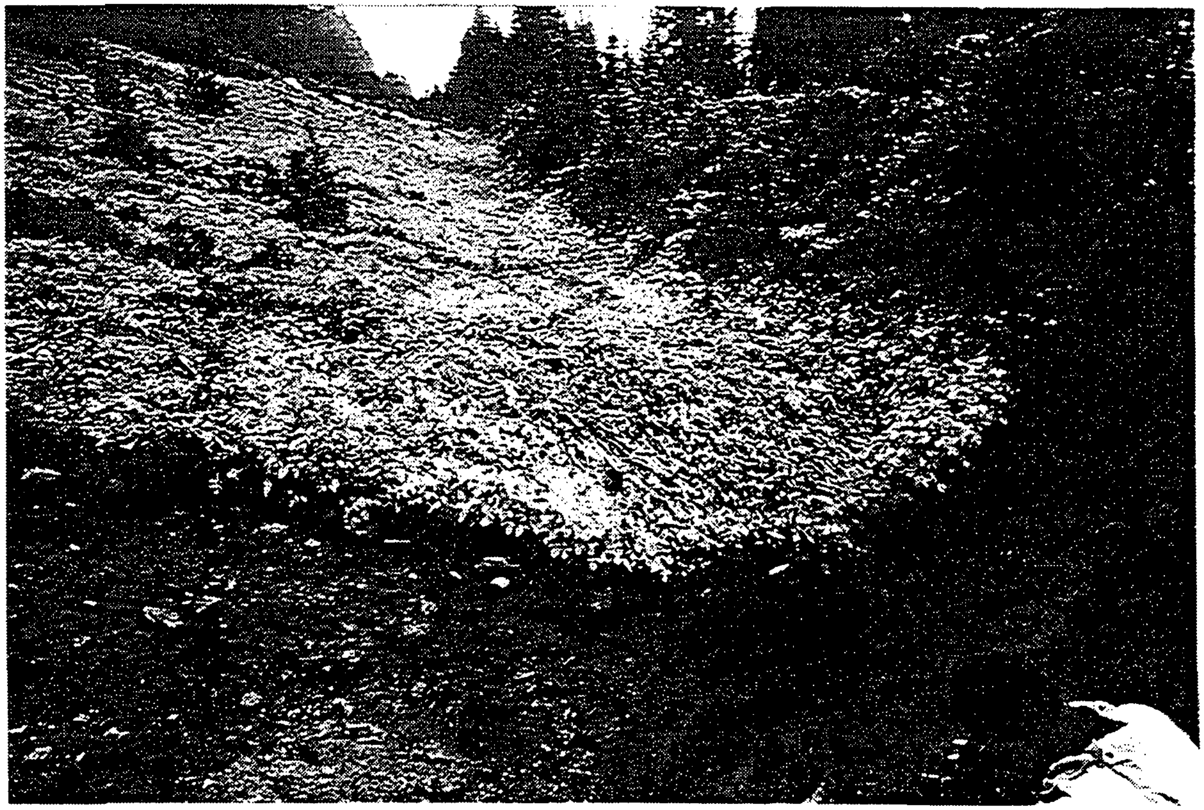

Figure JM.4B

4.3.1 CRITERION: Symmetry of Fault-Rock Structure in the Shear Zone

Chigira (1992, p. 174-175) describes the discrete shear zone that underlies areas of mass rock creep as "a pulverized zone with fault gouge and a phyllitic or brecciated zone (Figure JM.2). Although the above description could well apply equally to tectonic faults, the asymmetry of MRC-related fault-rock stratigraphy and orientation of mass rock creep-related faults may serve to distinguish them from tectonic faults. Chigira (1992) states that "A shear zone of an MRC fault, in general, has a sharp boundary surface with the downsliding block above it, but has an ambiguous transitional zone with the brecciated or phyllitic zone to the stationary rock mass below; in other words, the shear zones grades downward into a non-fractured or weakly fractured rock". This asymmetry is related to the stress field associated with the modern topography, and should not necessarily be mimicked by tectonic faults formed at crustal depths under various stress fields. Therefore, symmetrical (in cross-section) fault zones, or asymmetrical fault zones with the sharp contact on the lower boundary with respect to the modern ground surface, are more likely to be tectonic. Asymmetrical fault zones with the sharp contact at the upper boundary could be either: (1) caused by mass rock creep, or (2) a fortuitously-oriented tectonic fault.

4.3.2 CRITERION: Relation between Slope Morphology and Subsurface Structures

One of the strongest criteria for distinguishing gravity versus tectonic faults is the spatial relation between landforms (both small antislope scarps and troughs and the larger elements of mountain slopes) and the subsurface deformation zones. For tectonic faults, especially inactive ones, the fault trace may occupy almost any conceivable slope position, since the crustal stress field that determined the location and orientation of the fault may long predate the present topography. This is especially true for relatively narrow fault zones, where the zone of sheared rock is insufficiently wide to form a major landform element such as an erosional (strike) valley. In contrast, the

Figure JM.5A

Figure JM.5 Scarps of suspected gravity origin in the Bridge River area, southwestern British Columbia. (A) Arcuate scarp (in shadow) at the crest of "Santa Claus Mountain" (elevation ca. 2150 m ft) at the north edge of the Cayoosh Range, directly south of Shalalth on the south side of Lake Seton (visible at upper left). The scarp parallels the steep glacially-scoured slope down to Lake Seton (out of sight to left), which descends 1524 m in 3000 m, or an average gradient of 27°. (B) Arcuate graben at the crest of Nosebag Mountain, between Lake Carpenter and Lake Seton. Note the arcuate plan shape of the scarp (covered by talus), the erratic changes of height and morphology along strike, and the great height (up to 10 m) in relation to scarp length (about 200 m).

Figure JM.5, continued

Figure JM.5B

subsurface folds and faults created by mass rock creep must have a fixed spatial relationship to the topography that generated them. For example, Chigira (1992) describes how fold and fault zones associated with mass rock creep typically daylight in two locations relative to the local topography: (1) at (or near) the crest of a ridge, forming tensional features such as troughs and antislope scarps, and (2) near the toe of the mountain mass, often on oversteepened slopes. Between these two locations, deformation is localized as folding (in foliated rocks) or as a dilated zone of fractured rock (in massive rocks) (Figure JM.2). The oversteepened slopes are often the sites of secondary landslides, caused by the locally weakened nature of the rock mass and the excessive slope. This necessary spatial correspondence among mountain mass shape, fold/fault geometry, and sackungen-like landforms is consistently observed for mass rock creep-created features, but would be a coincidental occurrence for a tectonic fault.

4.4 Summary of Criteria

No single criterion enables unambiguous distinction between tectonic and nontectonic sackungen. At the reconnaissance level, geomorphic criteria may allow identification of large numbers of short, high, discontinuous, arcuate scarps near slope breaks as gravity failures. For longer, lower, continuous, linear scarps, on-site investigations are necessary. Such studies should include geodetic measurements for several years. Required geologic studies include reconstructing the style of near-surface

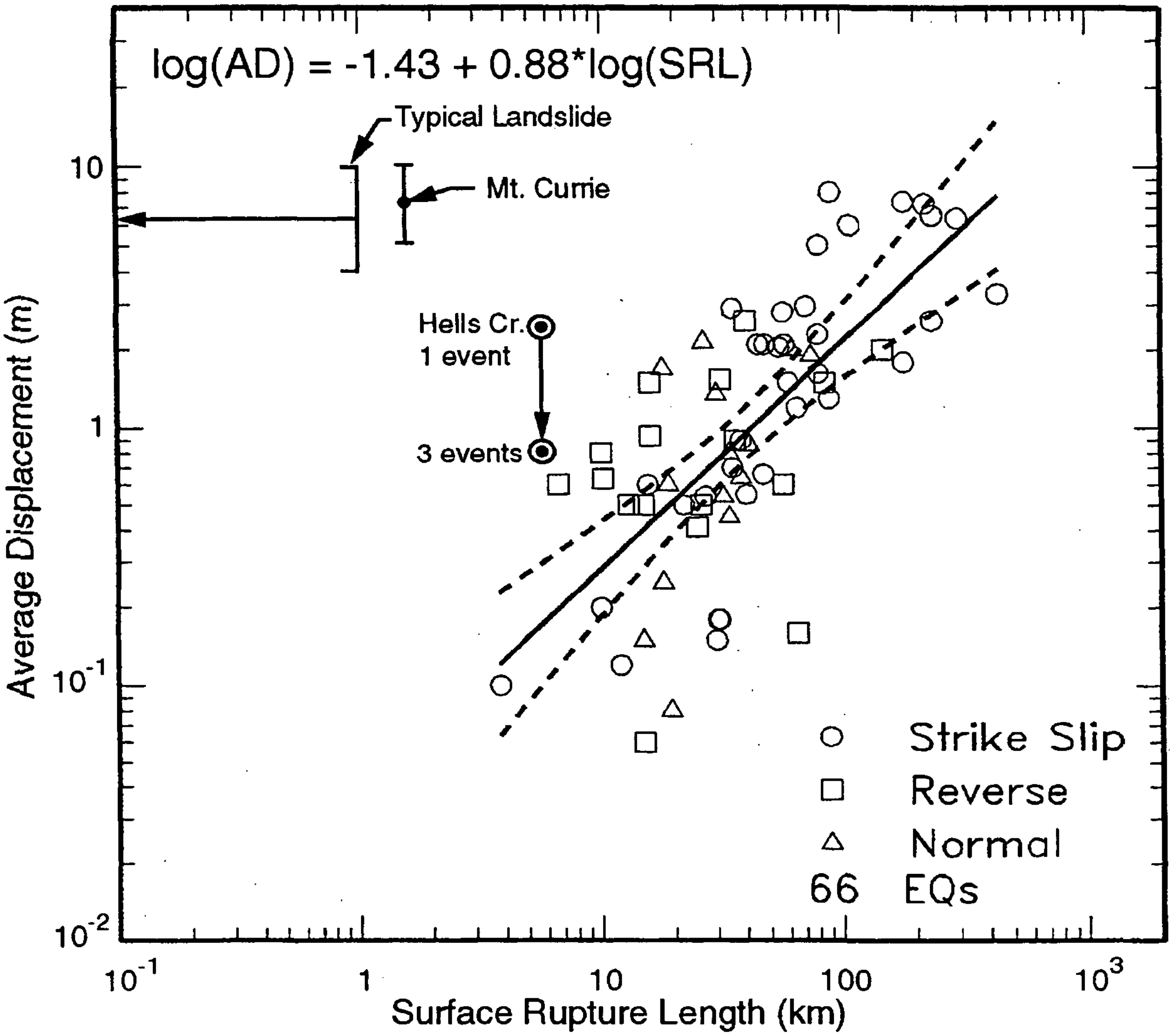

Figure JM.6 Plot of average displacement (m) as a function of surface rupture length (km) for historic fault surface ruptures. Solid line indicates ordinary least-squares regression line, dashed lines show 90% confidence intervals. Similar data for the Hell Creek scarp, Mount Currie scarp, and "typical landslides" are plotted at upper left. Surface rupture data from Wells and Coppersmith, 1994, Fig. 13. Permission to use this copyrighted material is granted by the Bulletin of the Seismological Society.

deformation and the history of continuous or episodic movement (via trenching), and determining the deeper subsurface geometry of the underlying shear zone. Evidence for continuous deformation rules out coseismic movement. If failure planes daylight lower on the slope, the sackungen are probably gravitational, but may be seismically triggered. Only when the sackungen overlie a fault zone with large displacement and steep dip, and deformation has been episodic and contemporaneous with other paleoseismic features, can sackungen be ascribed to primary tectonic faulting.

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ACTIVE TECTONICS IN A PASSIVE MARGIN SETTING

By

Frank J. Pazzaglia[1]

1 INTRODUCTION

1.1 Goals of this Expert Summary

Landscapes are the earth's expression of constructive, endogenic forces, and destructive geomorphic forces. All landforms integrate the effects of geology, climate, tectonics, and time in creating the earth's landscapes. Tectonic geomorphology is the study of how to read the tectonic signal in landforms, and by doing so, understand the underlying tectonic processes. Our understanding of active tectonic processes in active plate tectonic settings, such as a convergent margins, has been guided by tectonic geomorphic analyses. In this setting, tectonism is manifest clearly and dramatically as fault scarps, disrupted drainages, and deformed, young Quaternary deposits. However, in relative stable tectonic settings such as continental interiors or passive continental margins, we have only begun to understand how the landscape records and preserves tectonic processes. The goal of this expert summary is to address the problem of active tectonics in a setting traditionally viewed as stable - the middle U. S. Atlantic passive margin. Specifically, two case studies will be presented, both of which deal with the problems of active tectonics in a setting either (1) not traditionally considered tectonically active (the U.S. Atlantic passive margin), or (2) not ideal for the preservation of tectonic features in the landscape (the U.S. Pacific Northwest margin). Before presenting the case studies, the paper will define some concepts and terminology central to the problem of characterizing active tectonics on a passive margin. Important components of the geomorphic system will be examined to understand how active tectonics are manifest in the landscapes and deposits of a passive margin. Lastly, the paper will propose several mechanisms for active tectonics on a passive margin and avenues for future research dedicated to quantifying those processes.

1.2 Core of the Problem

Elevation and local relief vary considerably across the different physiographic provinces of the U.S. Atlantic passive margin (Figures FP.1a and FP.1b). The Piedmont lies about 250 m above sea level. It is underlain by high-grade metamorphic rocks and exhibits an upland surface of low relief (< 20 m) punctuated by river gorges where local relief does not exceed 180 m. Similarly, the Ridge and Valley, underlain by folded Paleozoic sedimentary rocks, rises to only 650 m above sea level and exhibits about 300 m of local relief. At least 9,000 m of Paleozoic sedimentary rocks remain in the Appalachian basin (Figure FP.1b), a observation consistent with the low relief of the Ridge and Valley and long-term rates of exhumation. In contrast, the Adirondack, White, and Green Mountain ranges of the New England Appalachians, and the Blue Ridge of Virginia are considerably higher and steeper than the Ridge and Valley. High peaks in the Adirondack and White Mountains rise over 1,500 m above sea level.

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Summits of the Blue Ridge reach 1,000 meters above sea level and loom 800 m above the Piedmont along the Blue Ridge escarpment. Particularly in New England, erosion has exhumed structurally deep parts of the Appalachian orogen exposing resistant mid-crustal Proterozoic and lower Paleozoic metamorphic and igneous rocks.

The central and northern Appalachian Mountains were built by several orogenic events throughout the Paleozoic, culminating with a continent-continent collision between North America and Africa during the Permian Alleghenian orogeny. Size and relief of the Permian Appalachians may have been similar to the modern central Andes, which have a mean elevation of about 3500 - 4500 m (Slingerland and Furlong, 1989). Erosion during the Permian and Early Triassic presumably removed most of the topography created during the Alleghenian orogeny with virtually all of the detritus being shed west into and beyond the Appalachian foreland basin. An increase in topography and relief was reintroduced into the Appalachian landscape in the Late Triassic and Early Jurassic associated with continental rifting which ultimately lead to the opening of the Atlantic ocean. A reversal of Appalachian drainage from the west to the east, which continues to the present, began with the formation of late Triassic and Jurassic rift basins (Judson, 1975). The modern offshore sedimentary basins formed during and subsequent to the rift and have long served as an effective trap of detritus shed eastward from the post-rift margin.

Conventional wisdom holds that eastern North America has been relatively tectonically inactive as it entered the drift stage in the Late Jurassic, evolving as the classic Atlantic-type passive margin. Nevertheless, contemporary deformation of the middle Atlantic margin does exist in the form of a broad, flexural warp, centered across the Fall Zone, between the upwarped central Appalachians and subsided Salisbury Embayment (Pazzaglia and Gardner, 1994), and is also expressed as seismic activity (Table FP.1). Smaller tectonic features such as high-angle reverse faults are superposed on this warp (Mixon and Newell, 1977; Mixon and Powars, 1984; Newell, 1985; Prowell, 1988; Gardner, 1989). It is important to point out that these faults are small with the known maximum total displacement on the order of tens to maybe 100 m since the Late Cretaceous. Similarly, the total amount of flexural up-warping of the Appalachian Piedmont is less than 20 m in the last 15 m.y. Herein lies an important macrogeomorphic paradox: studies of offshore sedimentary basins have clearly demonstrated several dramatic increases in sediment accumulation rates (Poag, 1985; 1992; Poag and Ward, 1993). The implication is that these events are related to increases in mechanical erosion rates. There remains considerable disagreement as to whether these increases in mechanical erosion rate were caused by tectonically-driven uplift or climatically-driven changes in rock erodibility.

2 CONCEPTS

2.1 Tectonic or Non-Tectonic Features

In the strict geologic sense, a tectonic feature is a structural component of rocks; the expression of strain imposed by a stress field. This report is concerned with the expression of strain primarily as crustal faults, and secondarily as long-wavelength bending or warping of the lithosphere called lithospheric flexure. There are numerous, diverse processes in which rocks undergo strain such as orogenesis or loading from an overlying feature such as a continental ice sheet. The root of the word "tectonics" comes from the Greek *tektos*

Table FP.1 Examples of historical seismicity in or near the Susquehanna drainage basin

Event	Date	Location	Depth	Moment Magnitude	Comments
Lancaster	4/23/84	Lat. 39°54'56" Long. -76°23'92"	7 km	4.2	Feld like a thundering train at the Peach Bottom Nuclear Power Plant
Reading	1/16/94	Lat. 40°34' Long. -76°03'	5 km	4.5-5.0	Damage in the Reading area. Thought to be the largest event recorded in Pennsylvania
Wyoming	4/8/95	Lat. 40°20' Long. -76°06'	4 km	2.6	Most recent event recorded

(builder) (Moores and Twiss, 1995) and carries the connotation of strain through constructive processes such as orogenesis. In this sense, strain imparted by the addition or removal of an ice sheet for example, would not be tectonic. The U.S. Atlantic margin is presently in a state of compressive stress where s_1 has a predominate northeast trend (Zoback and Zoback, 1989). In this report, any expression of this stress as strain on preexisting faults is considered tectonic. Strain associated with the application and removal of vertical loads in the form of denuded landscapes, sedimentary basin deposition, or ice, and manifest as flexural isostatic responses is considered non-tectonic.

Tectonic processes cause the deformation of rocks. The uplift, subsidence, or horizontal translation of rocks, with respect to a datum, such as sea level, is the single most important measure of a tectonic processes. An important distinction must be made between the deformation and uplift of rocks and the deformation and uplift of geomorphic surfaces (England and Molnar, 1990). Fluvial dissection of a landscape in response to eustatic fall results in the apparent rise of interfluves with respect to sea level, but no true tectonic uplift has occurred. In contrast, efficient denudational processes may keep a landscape from attaining a high mean elevation while tectonic processes uplift rock to the land surface. The latter is the observed case for very active, small orogens such as the southern Alps of New Zealand where rates of rock uplift and rates of denudation are roughly in equilibrium at about 10 mm/yr, resulting in steep, but relatively low-standing mountains with peaks ~3000 m above sea level. This report will always refer to tectonic deformation in the context of the rates of rock uplift with respect to a datum, typically contemporary sea level.

2.2 Comparison to Active Tectonic Features in the Western United States

Much of what is known about active faults in the United States comes from studies of the tectonically active western third of the country. Two important features of the western United States make this region ideal for the analysis of

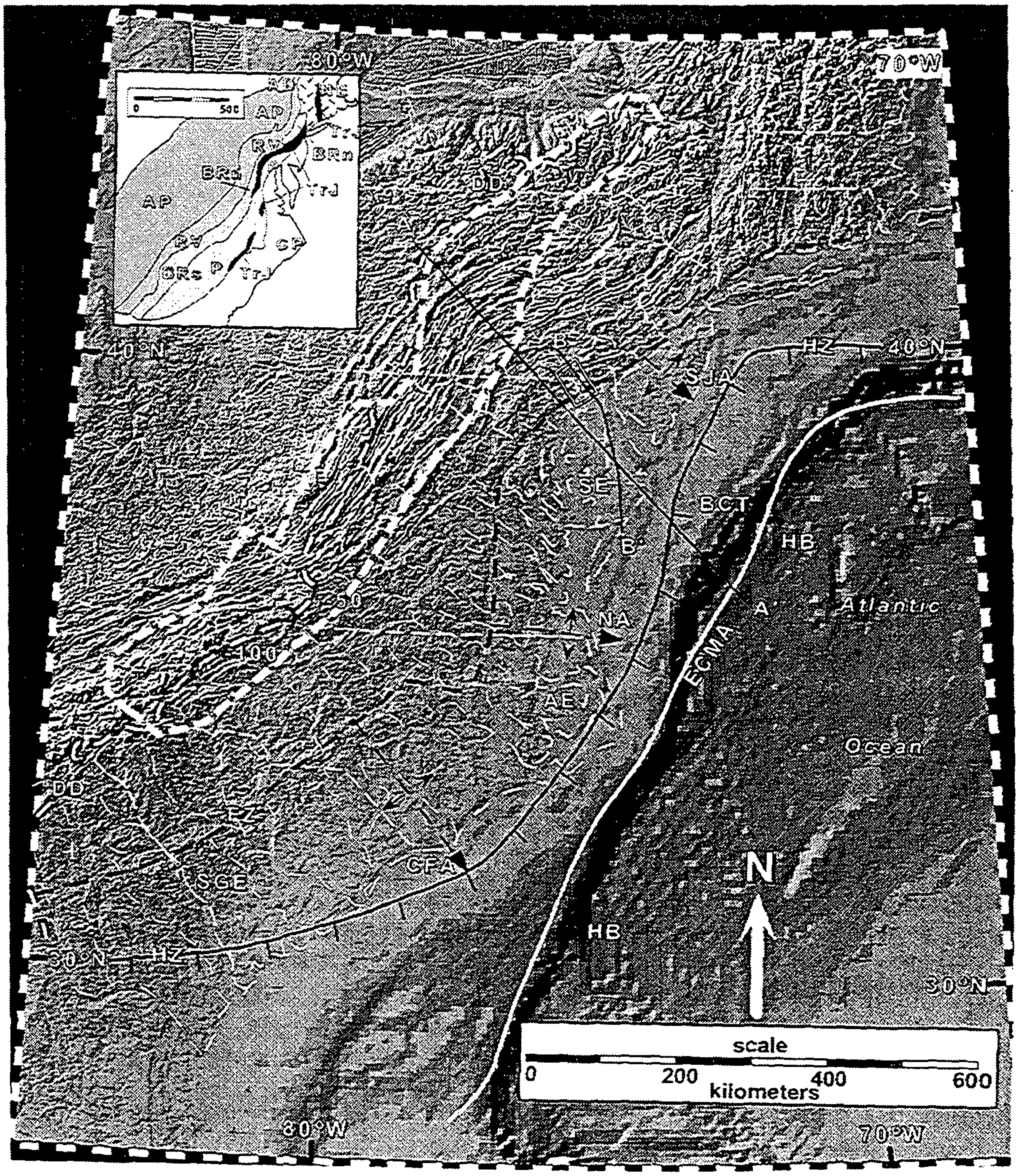

Figure FP.1 Examples of historical seismicity in or near the Susquehanna drainage basin. Modified from Pazzaglia and Gardner, Late Cenozoic Large-Scale Landscape Evolution of the U.S. Atlantic Passive Margin, M. Summerfield, ed., Geomorphology and Global Tectonics, Copyright © 1998. Reprinted by permission of John Wiley & Sons, Ltd.

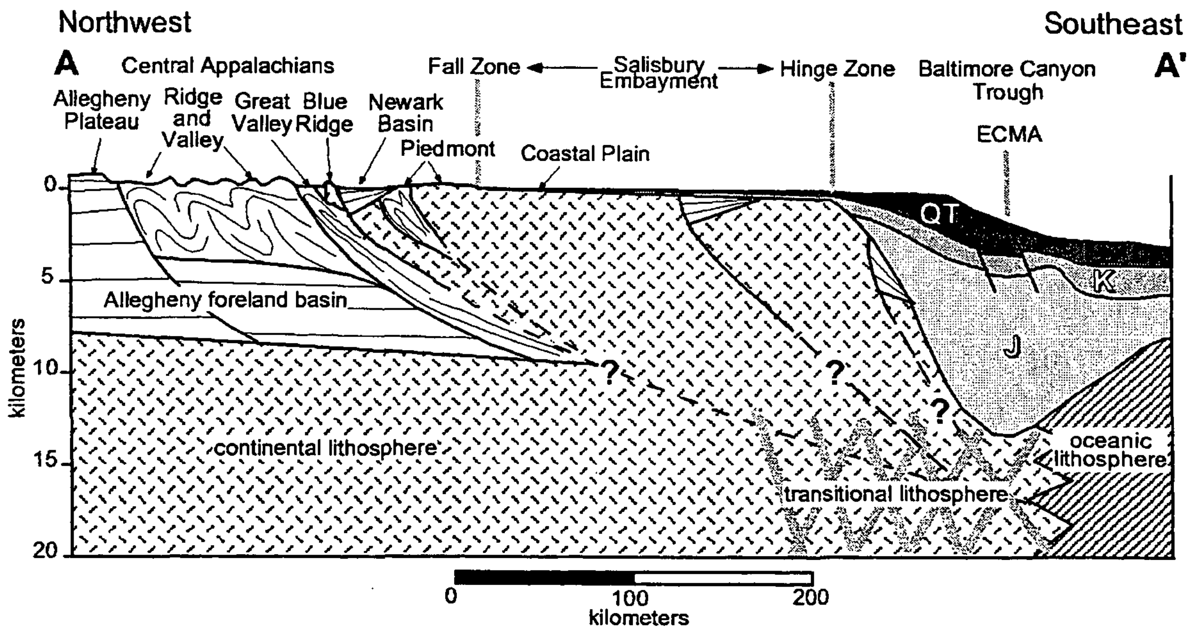

Figure FP.1b Modified from Pazzaglia and Gardner, Late Cenozoic Large-Scale Landscape Evolution of the U.S. Atlantic Passive Margin, M. Summerfield, ed., Geomorphology and Global Tectonics, Copyright © 1998. Reprinted by permission of John Wiley & Sons, Ltd.

active tectonic processes: (1) the climate is predominantly semi-arid and vegetative cover is limited leading to excellent exposure of rock and unconsolidated deposits; (2) the tectonic processes commonly involve uplift or translation of rocks at rates more rapid than the short-term rates of erosion, such that these tectonic processes are preserved in the landscape. Neither of these geologic/geomorphic conditions exist in the eastern United States which requires the geologist to focus on different landscape and stratigraphic proxies of tectonic deformation. These proxies include fluvial systems, hydrologic systems, Coastal Plain stratigraphy, carbonate valley karstsystems, lineaments, offshore sediment accumulation flux, and regional morphometric analyses. Each of these proxies are discussed individually in Section 3.

2.3 Tectonic Geomorphology

Geomorphology is the study of the earth's landforms and the processes that shape them. Landforms are the integrated expression of four important external geomorphic variables: climate, rock-type, time, and tectonics. Each of these variables imparts its own distinct signature on the landscape. It is the task of the tectonic geomorphologist to interpret and decipher the tectonic component of landscapes and from that component, attempt to understand the processes driving tectonic deformation. In an ideal setting, tectonic geomorphologists will choose a field study area where the effects of climate, rock-type, and time are either known or can be easily constrained. Most of our understanding of active tectonic processes comes from such studies. The challenge in non-ideal settings is to compile as much geologic information to constrain the effects of climate, rock-type, and time and then apply the concepts learned from studies in the ideal setting.

Oftentimes, when the landscape of the non-ideal setting simply will not surrender the obvious active tectonic proxies, the tectonic geomorphologist is forced to consider new, alternative explanations. Examples of alternative explanations are the subject of this report.

2.3.1 Reasonable Rates of Tectonic Deformation

Any geologic/geomorphic analysis of tectonic deformation and associated seismic hazards of the middle U.S. Atlantic margin must be placed in the context of the overall geologic and tectonic setting. The U.S. Atlantic margin is the "type" passive margin. It is not passive in the sense that it is completely free of tectonic activity and seismic hazards, but rather, the rates of deformation in this setting tend to be far slower than those found at convergent plate boundaries. In that sense, it is probably better to think of the U.S. Atlantic margin as a "trailing" rather than "passive" continental margin.

Compilation of diverse geomorphic and stratigraphic data representative of tectonic activity such as stratigraphic separation across fault scarps or rates of fluvial incision consistently show that short-term rates of deformation are not representative of long-term rates (Gardner et al., 1987). For very active, convergent plate margins, such as the western United States, the disparity in short- vs. long-term rates of deformation does not present a huge problem for active tectonic analyses because individual earthquake events are often frequent enough, or recorded well enough in stratigraphy and/or landscapes such that the natural variations in the deformation rates can be handled statistically. And herein lies one of the greatest challenges for active tectonic analyses in the eastern United States. In the passive margin setting, the short-term rates of deformation, represented for example by the stratigraphic offset of Quaternary sediments during a single earthquake across a Coastal Plain fault, are not at all representative of long term rates of deformation.

The middle U.S. Atlantic margin, from Virginia to New Hampshire, has a mean elevation of approximately 340 m above modern (Holocene) sea level (Figures FP.1a and FP.1b). Entire physiographic provinces such as the Piedmont and Coastal Plain have no elevations above 300 m. Such low-standing topography is simply not indicative of a landscape undergoing rapid rates of rock uplift. For example, if a fault is found to offset late Pleistocene (~20 ka) terrace gravels 2 meters, the inferred rate of slip is 0.1 mm/yr. If this rate is maintained over any significant period of geologic time, say one million years, some 100 m of rock uplift, or in other words, one third of the mean elevation of the Appalachians, would be expected along that single fault. An uplift rate of this magnitude will rapidly build higher and steeper topography than what is presently observed.

3 GEOMORPHIC EXPRESSION OF ACTIVE TECTONICS ON A PASSIVE MARGIN

In this section, geomorphic features and techniques for the identification and characterization of active tectonic features are discussed. these techniques include the use of fluvial systems, hydrologic systems, Coastal Plain stratigraphy, karst systems, continental-scale lineaments, continental scale denudation, and regional morphometric analyses.

3.1 Fluvial Systems

Rivers are among the best studied components of geomorphic systems. Geomorphologist know a great deal about fluvial processes and landforms, including the ways in which rivers are affected by active tectonics.

Rivers, in an anthropomorphic sense, are very lazy. Fluvial processes attempt to do both minimum work and uniform work (constant energy slope). These two goals are mutually exclusive and the river strikes a balance expressed in its morphology or hydraulic geometry. River channel depth, width, flow velocity, sediment distributions, sinuosity, and slope are dependent variables delicately adjusted to the independent variables of mean discharge and sediment supply. A change in either of these variables will immediately trigger a cascade of responses as all of the dependent variables mutually adjust. Active tectonic processes work in a direct way to fundamentally change the important dependent variable of slope. These slope changes are manifest in the river channel itself, in changes in the river valley, and in fluvial stratigraphy (Figure FP.2).

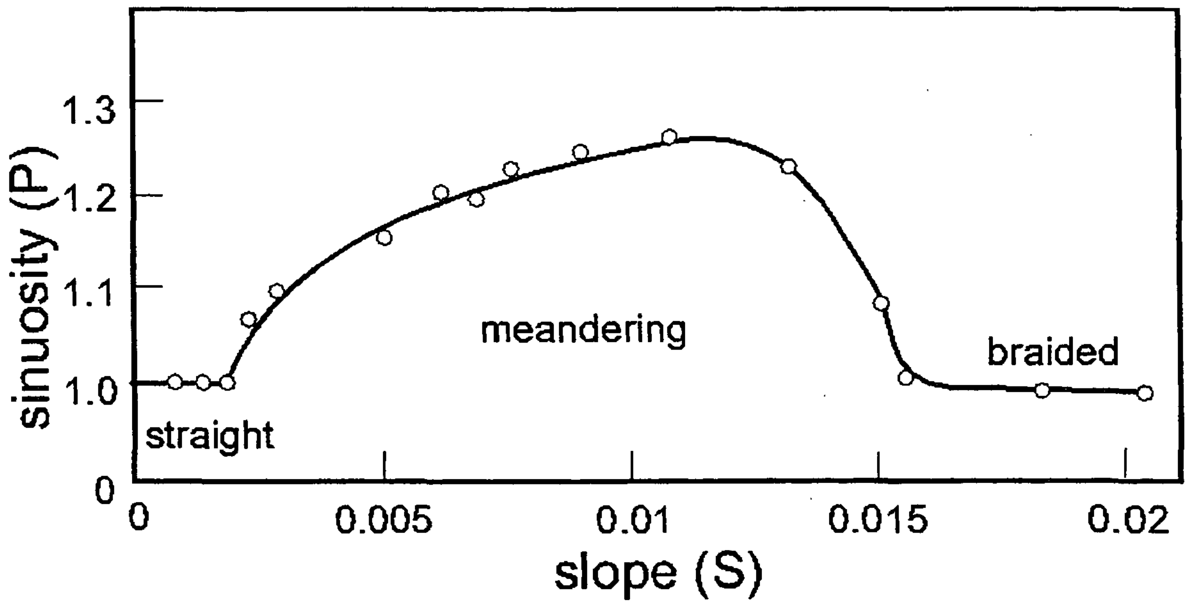

Figure FP.2 Clast identifications for lower Susquehanna River terraces and selected upper Coastal Plain and Fall Zone fluvial deposits. Modified from Schumm et al, Experimental Fluvial Geomorphology, Copyright © 1987. Reprinted by permission of John Wiley & Sons, Inc.

3.1.1 Channel Responses

Changes in sinuosity and in the width/depth ratio are most often cited as possible channel responses to active tectonic processes (Schumm et al., 1987). The vast majority of streams that have been studied in the context of active tectonics have been alluvial streams, that is, the channel is developed in the unconsolidated material being transported by river flow. Alluvial channels are fundamentally different from bedrock channels (Wohl et al., 1994). For example, processes of channel bed erosion for bedrock streams are different than those for alluvial streams so the well-understood hydraulic geometry relationships of alluvial channels do not necessarily apply to bedrock channels.

Most of the major river systems of the middle U.S. Atlantic margin such as the Susquehanna River are characterized by bedrock, rather than alluvial channels. After nearly 200 m.y. of post-Triassic subaerial denudation, the fluvial systems of the Appalachian mountains have become very well adjusted to rock-type and structure. With few exceptions, rivers have carved strike valleys into soft rock-types, or have exploited pre-existing structural weakness such as faults. In this sense, the bedrock channels of Appalachian streams are excellent locators of active faults. On the other hand, the relative slow rates of deformation associated with faults will probably not have a great effect on the hydraulic geometry of these bedrock channels.

3.1.2 Gorge Incision

Locally deep, narrow gorges cut by fluvial systems occur in the middle U.S. Atlantic margin. Deep river gorges typically are interpreted as fluvial incision in response to rock uplift. Within the glaciated regions of the Atlantic margin, most of the river gorges can be explained in terms of glacial and glaciofluvial processes. An example of a deep fluvial gorge in the Susquehanna river drainage basin attributable to glaciofluvial erosion is the Pine Creek Gorge of Tioga County, Pennsylvania. Here, an early Pleistocene proglacial lake formed between a south-facing ice dam, and a paleo north-flowing Pine Creek, a tributary of the paleo St. Lawrence drainage. The proglacial lake filled, until finding a natural outlet to the south, rapidly cutting the Pine Creek gorge, and reversing the flow of Pine Creek to the south into the Susquehanna River drainage basin.

Narrow gorges also occur where east-flowing Atlantic drainages must flow transverse to the dominant northeast-striking structural grain of the Appalachians. Near Harrisburg, Pennsylvania for example, the Susquehanna River has cut five major water gaps across resistant sandstone ridges of the Ridge and Valley physiographic province. The origin of the water gaps remains a much debated topic; however, most geomorphologists agree that they are long-lived features in the landscape and primarily reflect locally steep channel segments adjusted to rock-types of variable resistance. The one place where fluvial gorges may be directly related to tectonic activity and rock uplift is in the Appalachian Piedmont. Here, large streams such as the Susquehanna and Potomac Rivers have established their channels across very resistant, high-grade metamorphic rocks. The channel longitudinal profile of the Susquehanna River through this reach is strongly convex (see the case study below), a feature consistent with significant base level lowering. The river channel convexity is mirrored by a broad convexity in the Piedmont upland surface. Locally, rather abrupt offset of mapped upland gravel fluvial terraces (see the case study below) within the Piedmont surface convexity suggests fault deformation. Such deformation has been proposed as the source of a prominent knickpoint on the Susquehanna River at Holtwood (Thompson, 1985, 1990). The case study of Section 3 will further address the issue of the Susquehanna Piedmont gorge and attribute the base level lowering to both eustatic fall as well as flexural isostatic uplift.

3.1.3 Terraces

The geomorphic history of channel responses, incision, and aggradation are recorded as fluvial deposits along valley walls called terraces (Figure FP.3). A terrace is simply a former floodplain of a river that is stranded high enough above the modern channel such that it is no longer flooded. Terraces are best preserved in river valleys that have undergone a protracted period of incision. In these settings, terraces occur as inset fluvial deposits, stepping down to the modern floodplain. A terrace is a mappable, allostratigraphic unit. Geomorphologists concentrate on the study of

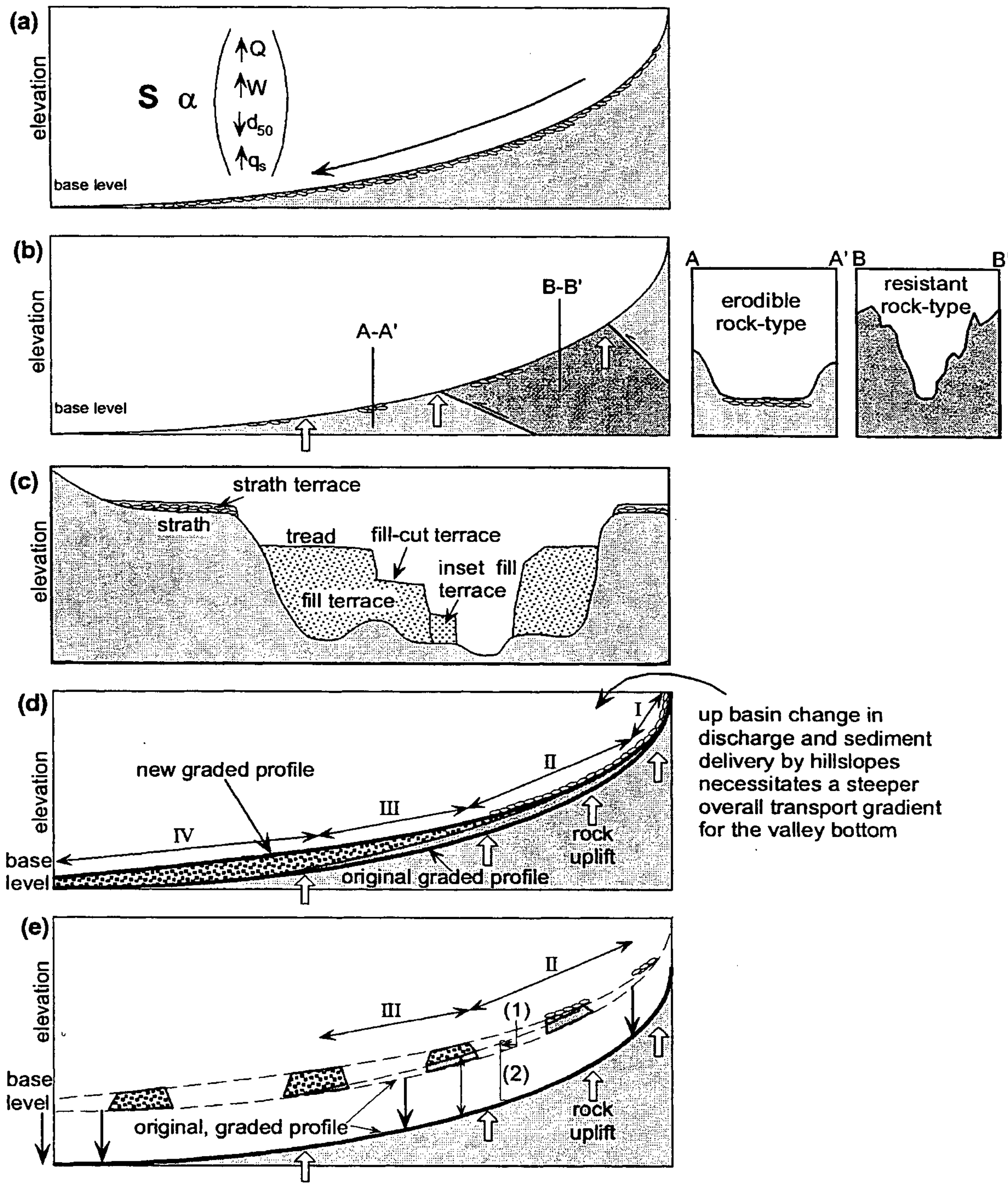

Figure FP.3 Concepts of bedrock channel graded profiles and terrace genesis. a) A graded profile for an alluvial stream develops the familiar concave-up longitudinal profile because discharge and slope are inversely proportional downstream. b) A similar graded profile shape can and does develop for bedrock streams; however, the reach slopes in this profile are delicately adjusted primarily to provide rates of incision equal and opposite to rates of rock-uplift through the channel (valley) bottom. Valley bottom widths respond to these rates of rock-uplift and incision and are adjusted locally to the amount of alluvium in the channel and variations in rock-type erodibility. c) A schematic valley cross-section illustrating the major types of fluvial terraces (from Bull, 1991). Straths are paleo-valley bottoms formed during former graded profile conditions. d) Temporal changes in overall graded profile concavity and reach-scale slope is dictated by changes in rates of rock uplift, down stream changes in base level, and/or upstream changes in basin-scale hydrology. This illustration considers an up basin increase in discharge and sediment load, and steady, uniform rates of rock uplift. An up-basin increase in discharge and sediment load is commonly driven by glacial-interglacial climatic cycles and in our example here, consider that the "original" graded profile represents the interglacial portion of that cycle, while the "new" graded profile represents the glacial portion of the profile. In general the initial response of the fluvial system is to adjust its hydraulic geometry and channel patterns without adjusting the valley gradient. In this way, valley bottoms (straths) are locally widened. Long profile gradients are adjusted as the stream attempts to accommodate and transport the increased sediment load only after all hydraulic geometry and channel pattern changes have been exhausted. The rate of fluvial bedrock incision is inversely proportional to increasing alluvial channel character because a channel at or above capacity will insulate the channel bed from erosive processes. With these considerations in mind, we represent spatial variability in the fluvial system response by zones I through IV. The channel is able to maintain a constant valley gradient and rate of incision in zone I where the valley gradient is very steep and the channel remains below capacity. Channel pattern changes may locally widen the valley bottom. An increasingly alluvial character in zone II initially widens the valley bottom, followed by a slowing in the rate of incision as the channel becomes increasing insulated. But the valley gradient is steep here and rock uplift quickly fills the accommodation space produced by the reduced rate of incision. The channel steepens beyond its original graded profile, increasing rates of incision and rapidly reaching a new graded mixed bedrock-alluvial valley bottom. As the valley gradient decreases downstream into zone III, rock uplift becomes less effective than alluviation at steepening the valley gradient. Incision ceases as the channel falls below capacity. The elevation of the valley alluvial tread defines the new graded profile. The river continues as a pure alluvial stream through zone IV. e) Return of the "new" long profile to its "original" lower-gradient form during a subsequent interglacial portion of the glacial-interglacial climate cycle isolates a paleo-valley bottom (strath) in the valley wall. We assume here that alluvial valley fills can be rapidly incised as the streams finds its "original" graded profile. Rock uplift passively carries the straths vertically where, if not removed by erosion, are mapped as terraces. Note the thick dark arrows of identical length indicating the amount of fluvial incision. The vertical distance measured in the field between the strath and the current valley long profile (thin, dark, two-headed arrow) is a measure of (1) rock uplift during attainment of the "new" graded profile and (2) rock-uplift since the strath was preserved in the valley wall. The amount of incision attributed to (1) is restricted to zones II and III and should not be interpreted as a response to variable rates of rock uplift.

terrace surfaces or treads, in part because soils developed in those treads capture the time of tread abandonment by fluvial processes. In essence, when a terrace tread is abandoned, the channel must have incised enough such that the former floodplain can no longer be flooded.

The terrace base or strath is also very important in that reconstruction of correlative straths along a river valley in essence captures the former longitudinal profile of the channel that made the terrace. Comparison of the modern channel gradient, to the terrace gradient is a direct measure of channel incision (Figure FP.3e). If that channel incision is driven by rock uplift, and an age for the terrace deposits can be determined, the terrace longitudinal profiles assume a central role in constraining both regional as well as local rates of rock uplift.

3.2 Hydrologic Systems

Subsurface ground water flow is well adjusted to rock-type and structure in the Appalachians in a manner similar to that observed for the surface drainages. Particularly in carbonate-floored strike valleys, ground water volume and flow is closely related to rock structures including fractures (joints) and faults. The 1984 Lancaster earthquake (MM 4.2; Stockar, 1989) nucleated along a north-south fracture system called the Fruitville Fault Zone. This fracture system is closely associated with a dolerite dike of Late Triassic or Early Jurassic age. The only clear geomorphic manifestation of the fracture system responsible for the earthquake is a line of travertine mounds associated with probable fracture-controlled groundwater flow (Alexander et al., 1989).

3.3 Coastal Plain Stratigraphy

The Coastal Plain of the middle U.S. Atlantic margin is underlain by gently east-dipping, Cretaceous to Recent unconsolidated marine and fluvial sand, silt, clay, and locally gravel. Given that the ages of deposits underlying the Coastal Plain are well known, tectonic deformation associated with faults will have well-constrained rates of offset. Numerous studies have documented offset, folding, or rapid changes in thickness of Coastal Plain strata associated with seismogenic or blind faults (Mixon and Newell, 1977; Newell and Rader, 1982; Mixon and Powars, 1984; Newell, 1985; Benson, 1990). Considerable debate persists as to the geomorphic expression of these faults and thus it is unclear if any of them have been active in the late Quaternary.

A persistent character of major middle Atlantic margin drainages is that they bend strongly to the southwest upon leaving the Fall Zone (Figure FP.1a). For the Potomac River, mapped Coastal Plain faults appear to have an important influence on that river's swing to the southwest. There is mounting evidence that similar faults, some which may have Quaternary offset, are located at the head of Chesapeake Bay and strongly influence the southwest bend in the Susquehanna River. At least 8 m of down to the northwest offset of early Pleistocene strata at the head of Chesapeake Bay (Pazzaglia, 1993; Figure FP.4) coincides with a steep magnetic gradient interpreted as a high-angle reverse fault (Higgans et al., 1974). Other structures, including a buried, southeast-tilted graben in northern Delaware (Spoljaric, 1973) attest to the concentration of faults in the upper Coastal Plain and Fall Zone region.

At a more regional scale, broad deformation of Coastal Plain strata suggests the character of the Atlantic margin's basin and arch architecture (Owens and Gohn, 1985; Figure FP.5). For example, the youngest Coastal Plain deposits are preserved in the center of basins such as the Salisbury Embayment. These deposits thin or are absent over the arches. In New Jersey, Cretaceous marine strata which is deeply buried beneath

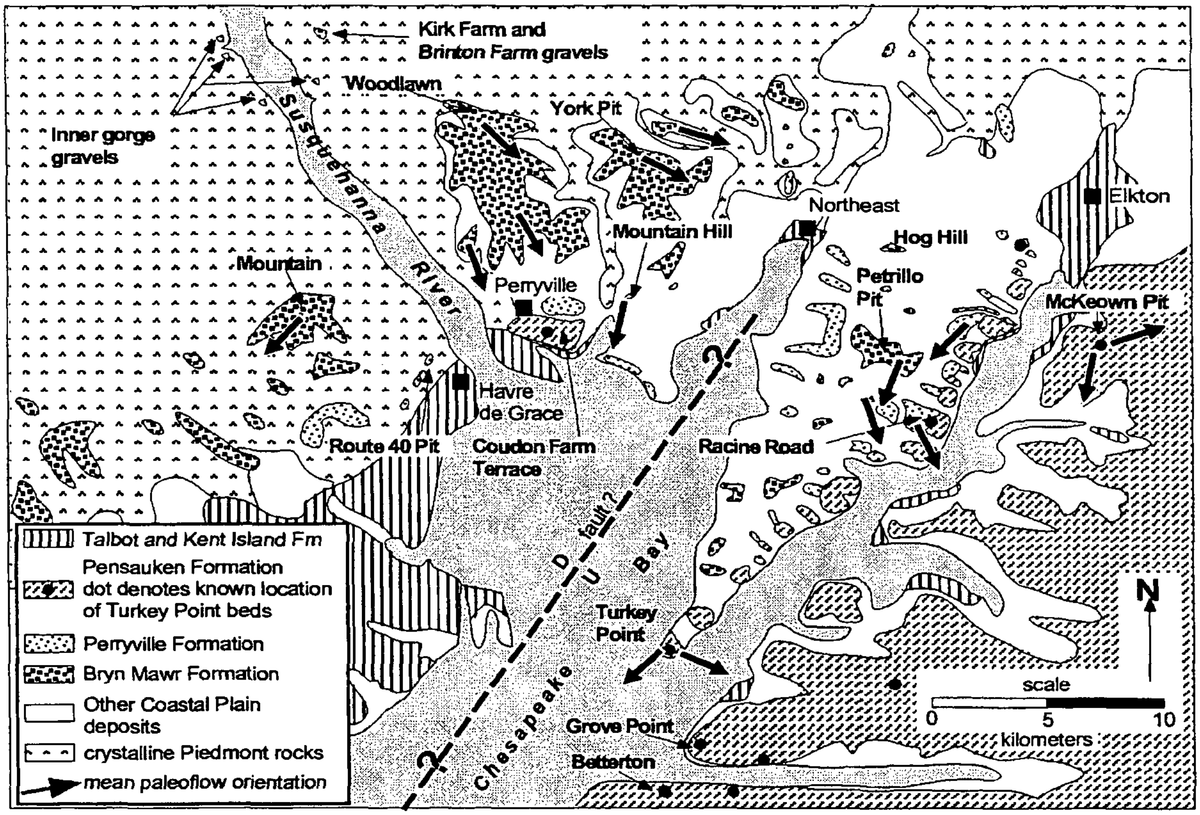

Figure FP.4 Geologic map of the upper Chesapeake Bay region showing the location of an inferred fault with Quaternary stratigraphic offset. From the Geological Society of America Bulletin, F.J. Pazzaglia. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1993 Geological Society of America.

Tertiary strata in the southern portion of the state, progressively shallow to the north, finally being exposed at Raritan bay in the northeast portion of the state. These Cretaceous beds project out and over an imaginary New England Coastal Plain,except for very limited exposures of Cretaceous strata on Long Island and Martha's Vineyard. The absence of a New England Coastal Plain may be a megageomorphic expression of regional uplift and down to the southwest tilting of the Atlantic margin in the late Cenozoic (Pazzaglia, 1993).

3.4 Carbonate Valley Karst Systems

Strike valleys of the Ridge and Valley and the Great Valley (Figures FP.1a and FP.1b) are underlain by lower Paleozoic carbonates where widespread karst has developed. Locally, where cave systems now lie above local base levels, cave stratigraphy can be used to estimate local rates of valley floor lowering which is in part reflective of regional rates of rock uplift.

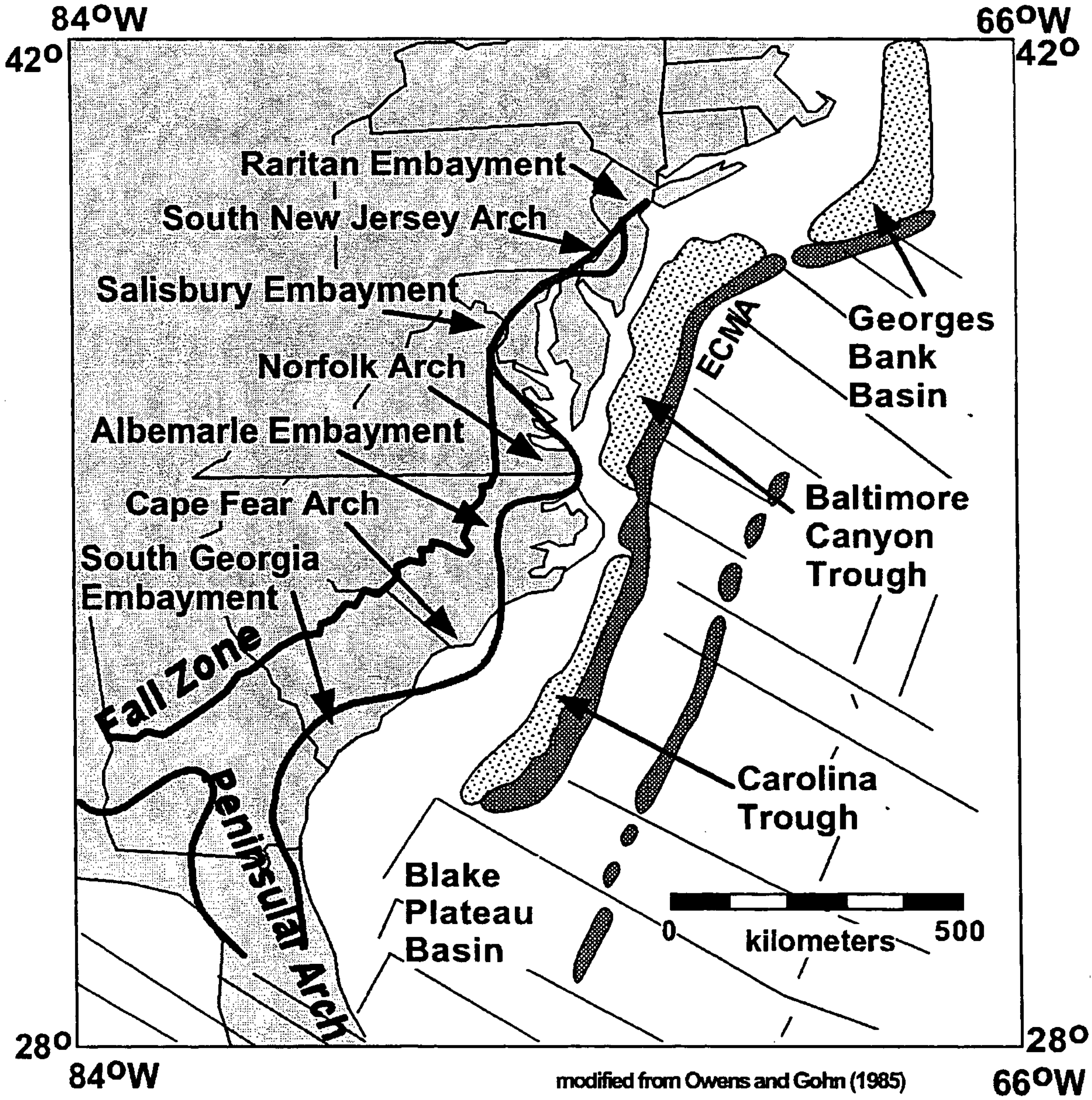

Figure FP.5 Basin and arch architecture of the U. S. Atlantic margin (from Owens and Gohn, 1985). Reprinted by permission of Kluwer Academic Publishers, Owens, J.P., and G.S. Gohn, "Depositional History of the Cretaceous Series in the U.S. Atlantic Coastal Plain: Stratigraphy, Paleoenvironments, and Tectonic Controls on Sedimentation" (C.W. Poag, ed.), Geologic Evolution of the Atlantic Margin, figure 2-1, p. 27, 1985.

The geology of the Great Valley is particularly interesting because it is flanked by some of the highest and steepest topography of the Atlantic margin represented by the Blue Ridge mountains. From Pennsylvania through Virginia, detritus shed west from the Blue Ridge has been preserved in the Great Valley as alluvial fans up to 60 m thick (Hack, 1960; Whitticar and Duffy, 1992).

Locally, the juxtaposition of these fans against the Blue Ridge mountain front is very abrupt and has been suggested to represent a fault. In numerous gravel pits, the fan deposits themselves are locallyfolded, faulted, and intruded by clastic dikes reminiscent of seismogenic soft-sediment deformation features (Gardner et al., 1993). Unfortunately these features can not be attributed solely to tectonism because the fan deposits overlie the karst-prone carbonates in the valley bottom. Nevertheless, these alluvial fan deposits represent the most extensive Cenozoic stratigraphy in the Appalachians outside of the Coastal Plain, which may record geomorphic and stratigraphic evidence of tectonic deformation.

3.5 Lineaments

Large-scale geomorphic features called lineaments have been described in the Appalachian landscape (Hobbs, 1904; Gold et al., 1973, 1974; Wise, 1974; Wheeler, et al., 1974, Wheeler, 1978; Kowalik, 1975; Lavin, 1982); yet little attention has been focused on these features as potential sources or locations of tectonic deformation. The lineaments are oriented northwest-southeast and coincide with major gravity and magnetic trends which has lead some researchers to postulate that they are very deep-seated crustal inhomogeneities separating distinct tectono-geomorphic structural provinces (Gold et al., 1974; Lavin and Alexander, 1981; Lavin, 1982; Figure FP.6). One of the more prominent of these structures, the Tyrone-Mount Union lineament, projects across the lower Susquehanna River drainage. This lineament is defined by various geomorphic features such as northwest aligned stream reaches, gaps through resistant ridges, and intensified karst development in carbonates (Figure FP.9). The lineament projects into the Pennsylvania Piedmont and essentially through the epicenter of the 1984 Lancaster earthquake. The role that this, and other lineaments, serve to concentrate regional stresses acting on the margin is unknown.

3.6 Offshore Sediment Accumulation Flux

Broad patterns of rock uplift in the Appalachians are preserved in the volume and location of detritus in the offshore sediment basins of the Atlantic margin (Poag, 1985; Poag and Sevon, 1989; Poag, 1992; Poag and Ward, 1993). Pertinent to the study of active tectonics along the Atlantic margin is the observation that the flux of detritus to the offshore basins has significantly increased throughout the late Cenozoic. In particular, the high relief portions of the margin, such as New England, have shed a disproportionately large amount of sediment into the offshore basins. These data suggest that New England may presently be undergoing broad epeirogenic uplift (Pazzaglia and Brandon, 1996).

3.7 Regional Morphometric Analyses

Related to the distribution of detritus in offshore basins is the expression of Appalachian topography as it relates to mean rates of rock uplift. Rock erodibility plays a very important role in local relief (Hack, 1982) and in general, the Appalachian landscape does not exhibit the strong correlation between mean elevation and mean local relief typical of rapidly uplifting orogens. Nevertheless, broad patterns of increased elevation and relief are present and may be indicative of broad epeirogenic processes. Differences in mean elevation and relief, when viewed at the scale of physiographic provinces, provides a bases for the subdivision of the middle Atlantic margin into tectono-geomorphic domains (Brown et al., 1972). These domains may serve as a useful spatial basis for the characterization of seismogenic activity.

4 A CASE STUDY - THE SUSQUEHANNA RIVER BASIN[1]

4.1 Executive Summary

Lower Susquehanna River fluvial terraces offer a unique opportunity to investigate the late Cenozoic tectonic, geologic and geomorphic evolution of the U.S. Atlantic passive margin (Figure FP.5). Petrography and elevation distinguish and provide a basis for correlating two groups of terraces, the upland terraces and lower terraces, through the Piedmont, Newark-Gettysburg Basin, and Great Valley. Downstream correlation to dated upper Coastal Plain and Fall Zone fluvial deposits, relative weathering, and soil profile development characteristics establish terrace age. Upland terraces (Tg1, Tg2, and Tg3), middle to late Miocene strath terraces 80 to 140 m above the present channel, occur only along the Piedmont reach. They are underlain by unstratified, texturally-mature, quartz-dominated, roundstone diamictons. In contrast the lower terraces (QTg, Qt1 - Qt6), Pliocene and Pleistocene strath and thin aggradational terraces within 45 m of the present channel, are underlain by stratified and unstratified, texturally and compositionally immature sand, gravel, and pebbly silt.

Terrace age and longitudinal profiles suggest complex interactions between relative base level, long-term flexural isostatic processes, climate, and river grade. A model for terrace genesis requires the Susquehanna River to attain and maintain a characteristic graded longitudinal profile over graded time. For the U.S. Atlantic margin, the model proposes that straths are continually cut along this graded profile during periods of relative base level stability, achieved by slow, steady, isostatic continental uplift acting in concert with eustatic rise. Change in an external modulating factor, such as eustatic fall or climate change, results in fluvial incision and subsequent genesis of strath terraces. Convex-up longitudinal profiles of lower Susquehanna River terraces, which converge at the river mouth, diverge through the Piedmont, and reconverge north of the Piedmont, stand in contrast to their hypothesized, original concave-up profiles. Progressive and cumulative flexural upwarping of the Atlantic margin accounts for terrace profile deformation suggesting flexural isostasy as a first-order, regional deformation mechanism. These results offer new interpretations of terrace age, correlation, and geologic significance that require modification of previous studies suggesting uplifted or anticlinally-warped peneplains on the U.S. Atlantic margin.

4.2 Introduction

Fluvial terraces preserved along the lower reaches of the Susquehanna River, record the late-stage geologic and geomorphic evolution of the U.S. Atlantic passive margin (Pazzaglia and Gardner, 1993). Early studies, driven by the need to understand peneplain genesis and uplift (Davis, 1889; Barrell, 1920; Bascom, 1921; Knopf, 1924; Stose, 1928; Knopf and Jonas, 1929; Ashley, 1930, 1933; Campbell, 1933; Hickok, 1933), recognized the geomorphic importance in identifying, correlating, and dating these terraces (Wright, 1892; Bashore, 1894, 1896; Stose, 1928, 1930; Campbell, 1929, 1933; Jonas and Stose, 1930; Ashley, 1933; Hickok, 1933; Leverett, 1934; Mackin, 1934; Stose and Jonas, 1939; Peltier, 1949). Previous terrace correlations

[1] Modified from Pazzaglia (1993), Pazzaglia and Gardner (1993), Pazzaglia and Gardner (1994), and Pazzaglia and Gardner, 1998.

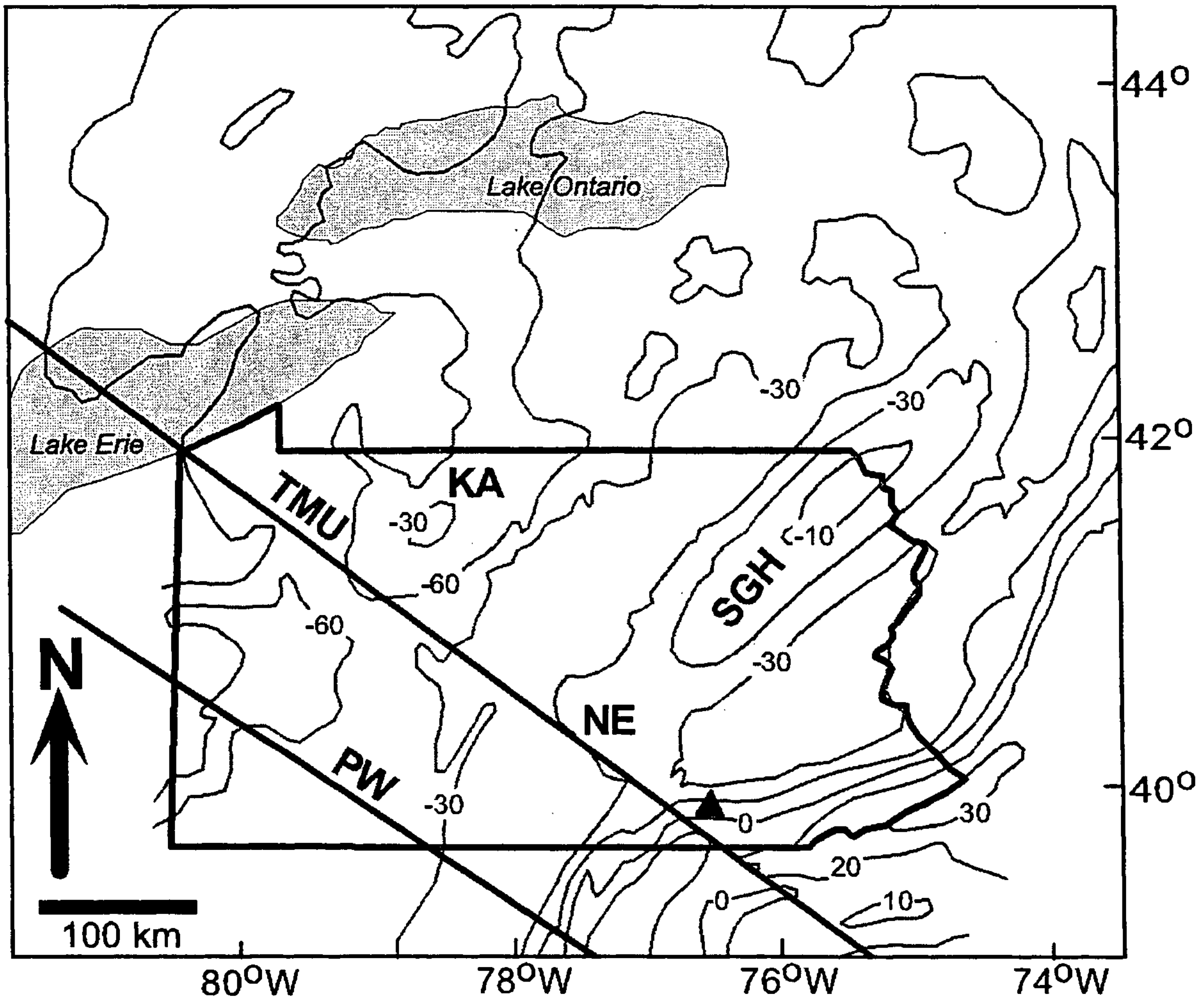

Figure FP.6 Major lineaments of the central Appalachians superposed on a simple Bouger anomaly map (contours in mgal). TMU = Tyrone-Mt. Union Lineament; PW = Pittsburgh-Washington Lineament; SGH, KA, and NE refer to the Scranton, Kane, and Newport gravity highs respectively. The trangle locates the 1984 Lancaster earthquake.

suggested uplifted, seaward-sloping peneplains (Stose, 1928, 1930), a broad northeast-southwest trending elongate dome called the Westminster Anticline (Campbell, 1929, 1933), or erosion surfaces graded to tectonically and eustatically-generated knickpoints (Hickok, 1933). These incongruous interpretations arose because: 1) systematic petrographic and textural criteria were not used to identify and correlate fluvial terraces; 2) no distinction or genetic relation was made between depositional and erosional fluvial features; 3) a relationship linking terrace genesis to the complex interaction between passive margin isostatic, eustatic, and climatic processes was not established; and perhaps most importantly, 4) terrace age was generally poorly constrained.

Recently, fluvial deposits at the mouth of the Susquehanna River have been dated by stratigraphy and petrography-based downdip correlations to marine deposits in the Salisbury Embayment (Pazzaglia, 1993). Within the context of this new age control, fluvial terraces along the lower Susquehanna River could be mapped, correlated, and dated. The terraces could then be used to develop a model for terrace genesis on a U.S. Atlantic-type passive margin suggesting the nature, magnitude, and rates of passive margin neotectonic deformation (Pazzaglia and Gardner, 1994, 1998).

4.3 Data

Offshore basins. Sediment derived from east-flowing Atlantic margin drainages has been deposited in sedimentary basins underlying the Coastal Plain, continental shelf, and continental rise (Figure FP.1 and FP.5). Poag (1985, 1992) and Poag and Sevon (1989) subdivided the offshore basin stratigraphy into 23 informal time-stratigraphic units. They subsequently collapsed these into 12 formal allostratigraphic formations (Poag and Ward, 1993). Poag's inventory accounts for all significant sedimentary accumulations including deep-sea sediments of the continental rise. Pazzaglia and Brandon (1996) have recalculated the siliciclastic volumes using the time scales of Harland *et al.* (1990) and Cande and Kent (1992) and the depth-porosity curve for the COST-B2 well (Scholle, 1977). The fluxes that produced these volumes are plotted against geologic age and are given in terms of the solid-rock volume delivered into the basins per m.y. (Pazzaglia and Brandon, 1996). Sediment loads are calculated as the product of sediment thickness represented on the isopach maps (Poag and Sevon, 1989; Poag, 1992; Popenoe, 1985) and sediment density (Scholle, 1977). Sediment thickness, in meters, for the Baltimore Canyon Trough is estimated to be 1000 times the two-way travel times shown on the isochron maps (Poag, 1992).

Continental Denudation. Rock removed from the Atlantic margin by both mechanical and chemical erosion constitutes a negative, upward-directed load that is the product of rock density and the vertical thickness of denudation. Geochemical mass balance studies of saprolite production rates suggest average total denudation rates in the Appalachian Piedmont ranging from about 5 to 50 m/m.y. (Cleaves et al., 1970; 1974; Cleaves, 1989, 1993; Pavich, 1985; Pavich et al., 1989; Pavich, 1989). Similar, drainage basin wide rates for mechanical erosion based on modern sediment yield data are reported for major Atlantic drainages such as the Susquehanna, Juniata, Delaware, and Potomac Rivers (Judson and Ritter, 1964; Sevon, 1989; Milliman and Syvitski, 1992). Considering that modern sediment yield data producing the higher end of the observed denudation range probably reflects anthropomorphic activities and may not be representative of actual long-term rates of denudation, we adopt a uniform, mean total (sum of chemical and mechanical) erosion rate of 10 m/m.y.

Piedmont Terraces. Fluvial terraces, mapped and correlated on the basis of petrography and elevation, flank the lower Susquehanna River (Pazzaglia and Gardner, 1993; Engle et al., 1996) (Figures FP.7 to FP.10; Tables FP.2 and FP.3). These terraces (Tg1, Tg2, Tg3, and QTg) are degraded strath terraces cut on the Piedmont bedrock and mantled with a thin colluvium of fluvially rounded pebbles and cobbles. The petrography of these deposits begins almost exclusively as massive vein quartz for the highest and oldest terrace Tg1, becomes progressively more heterolithic with the introduction of quartzite clasts in Tg2 and more labile sandstones and siltstones in Tg3, and very heterolithic with extrabasinal granites and gneisses of QTg. These petrographic trends mirror those of the upland gravels and Coastal Plain deposits and allow for

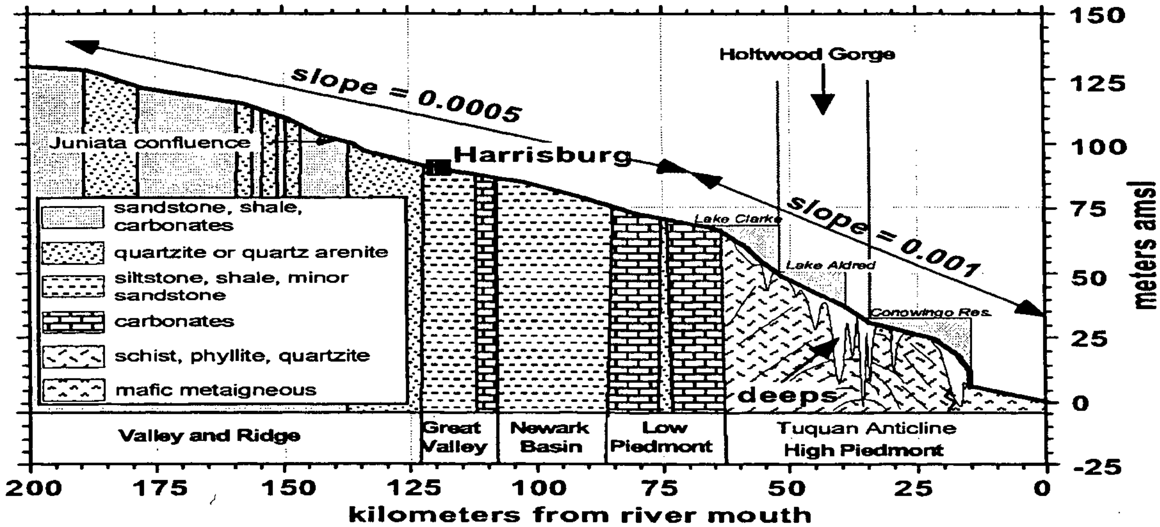

Figure FP.7 Longitudinal profile of the Susquehanna River from Northumberland to the head of Chesapeake Bay. Spoon-shaped, discontinuous thalweg "deeps" (Mathews, 1917) are illustrated in the Piedmont reach. Curves 1 and 2 represent a straight line and exponential projection, respectively, of the upper profile from the confluence with the Juniata River across the lower profile convexity. From the Geological Society of America Bulletin, E.B. Mathews. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1917 Geological Society of America.

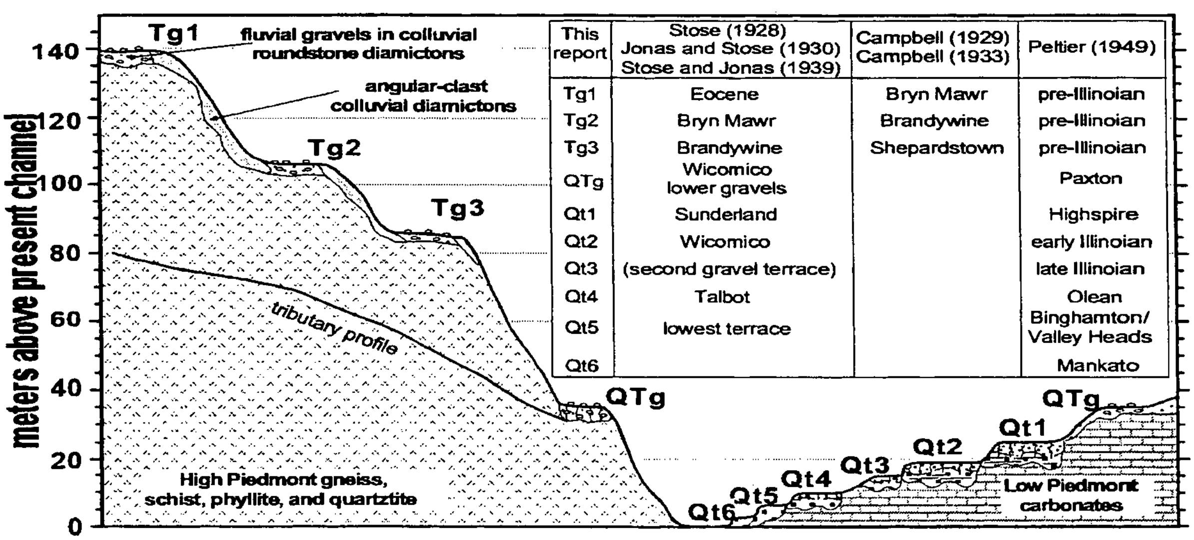

Figure FP.8 Composite cross-section of the lower Susquehanna River valley showing the relative location and elevation of terraces.

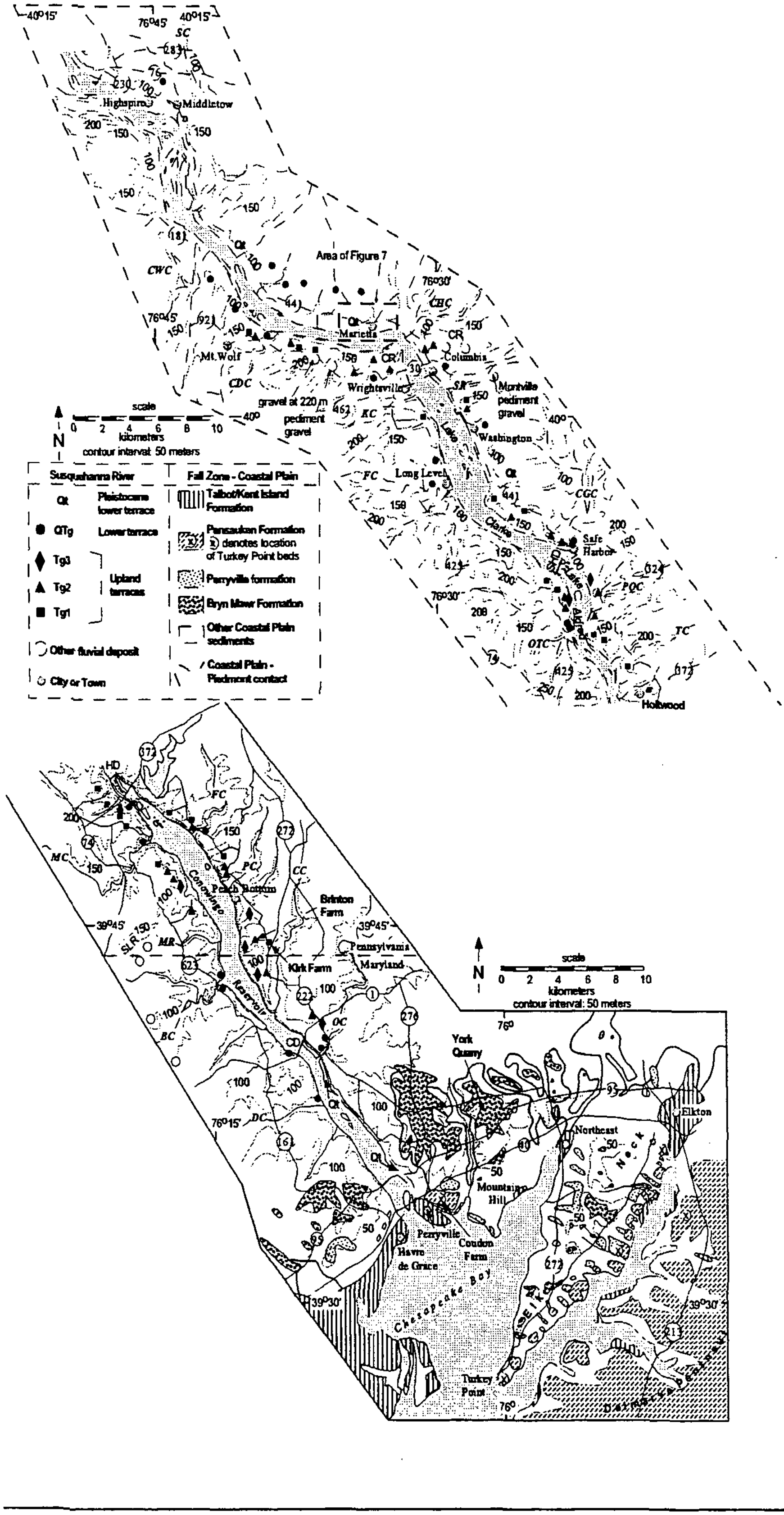

Figure FP.9

Maps of the lower Susquehanna River basin showing the location of terraces and a generalized distribution of upper Coastal Plain and Fall Zone fluvial deposits modified from Owens (1969) and Conant (1990).

BC = Broad Creek
CC=Conowingo Creek
CD= Conowingo Dam
CDC= Cordorus Creek
CGC=Conestoga Creek
CHC = Chickies Creek
CR = Chickies Ridge
CWC=Conowago Creek
DC = Deer Creek
HD = Holtwood Dam
FC = Fishing Creek
KC = Kreutz Creek
MC = Muddy Creek
MR = Muddy Run
OC = Octararo Creek
OTC = Otter Creek
PC = Peters Creek
PQC = Pequea Creek
SC = Swatara Creek
SHD=Safe Harbor Dam
SLR = Slate Ridge
SR = Strickler Run
TC = Tucquan Creek.

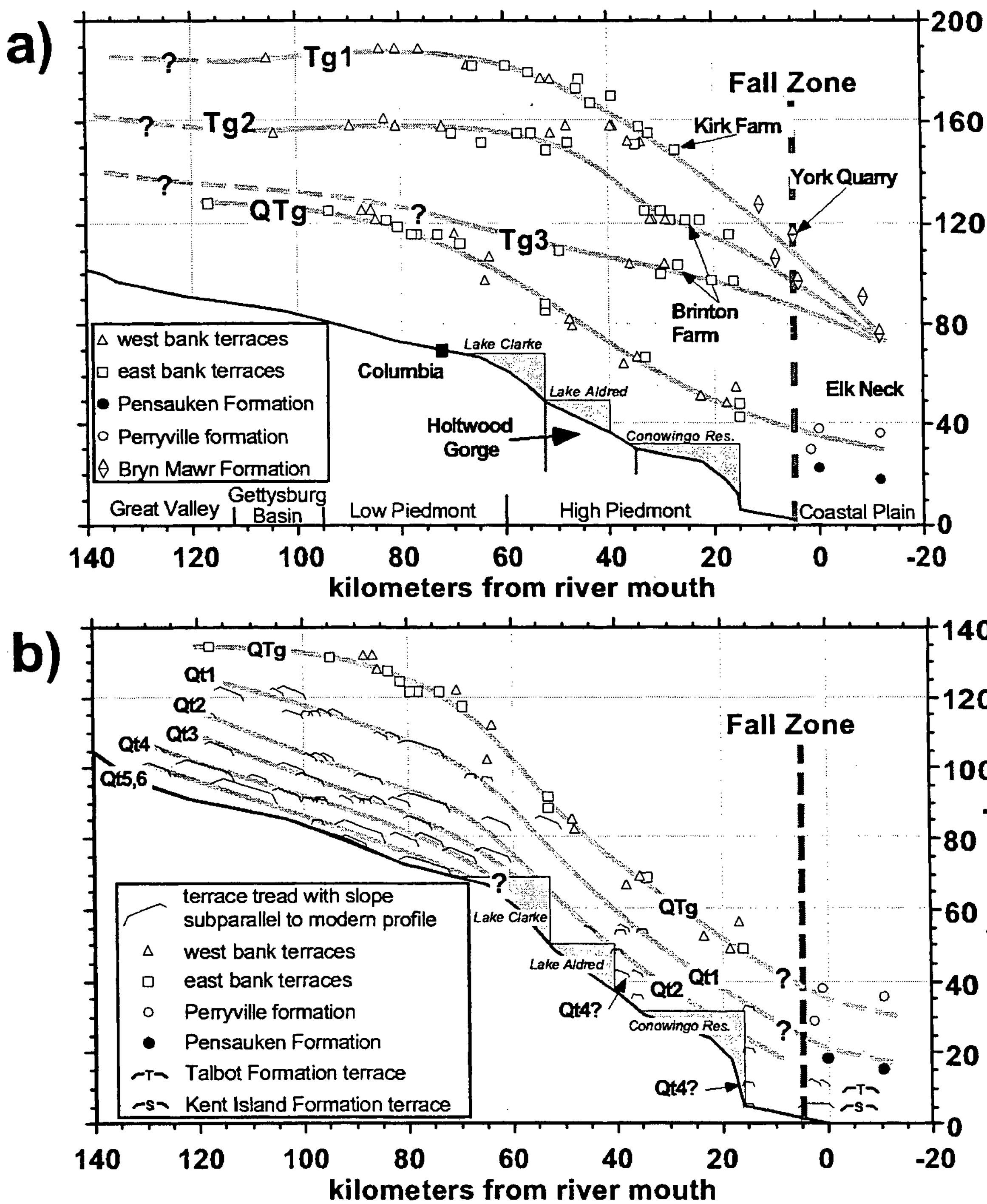

Figure FP.10 Correlation of (a) upland terraces and (b) lower terraces along the lower Susquehanna River from the head of Chesapeake Bay to Harrisburg, Pennsylvania. Profiles were produced by projection to a vertical plane located in the center of the Susquehanna River.

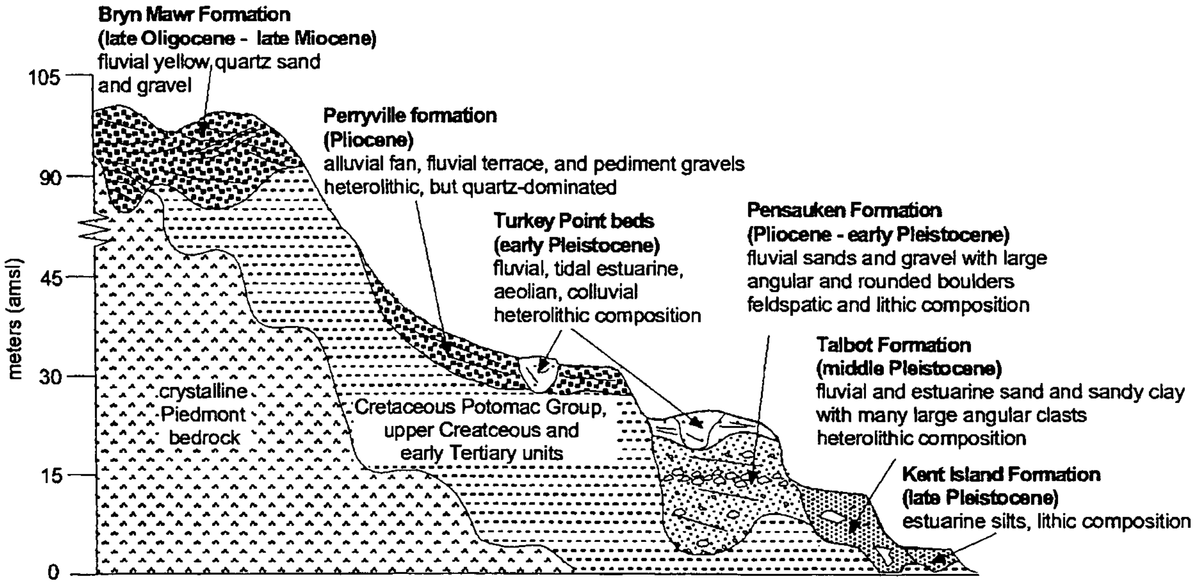

Figure FP.11 Schematic diagram of upper Coastal Plain and Fall Zone fluvial stratigraphy at the head of Chesapeake Bay. Modified from the Geological Society of America Bulletin, M.P.A. Jackson, and B.C. Vendeville. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1994 Geological Society of America.

downstream correlation to dated deposits (Figures FP.4, FP.11 and FP.12).

Our model for terrace genesis along the lower Susquehanna River requires a graded river (Leopold and Bull, 1979; Knox, 1975) with a fixed base level to attain and maintain a characteristic longitudinal profile on an isostatically dominated margin (Pazzaglia and Gardner, 1993). Terrace genesis along major streams, such as the Susquehanna, in the Appalachian landscape spans a time range commensurate with a period of Coastal Plain deposition. The minimum age of a fluvial terrace would be approximately equal in age to the unconformity at the top of a Coastal Plain unit deposited during the coincident period of terrace genesis. Thus, correlation of terraces, Fall Zone upland gravels, Coastal Plain deposits, and offshore basin deposits can be used to define stratigraphic horizons of nearly synchronous age.

Coastal Plain. The middle Atlantic Coastal Plain is the subaerially exposed portion of the Salisbury Embayment, a large arcuate-shaped basin approximately 300 km in diameter flanked on the north by the South Jersey Arch, on the south by the Norfolk Arch, and on the west by the Fall Zone (Owens and Gohn, 1985; Figures FP.1 and FP.5). Detailed stratigraphic reconstructions obtained from field mapping and borehole analyses have demonstrated that the Salisbury Embayment has a complex Cenozoic depositional history attributed to isostatic, tectonic, and eustatic processes (e.g. Brown et al., 1972; Newell and Rader, 1982; Newell, 1985; Owens and Gohn, 1985; Ward and Strickland, 1985; Figure FP.12).

Appalachian Piedmont rocks continue as Salisbury Embayment basement seaward of the Fall Zone, dipping seaward more steeply than anywhere else along the Atlantic Coastal Plain (Hack, 1982). Cretaceous to Eocene predominantly marine deposits unconformably overlie the basement,

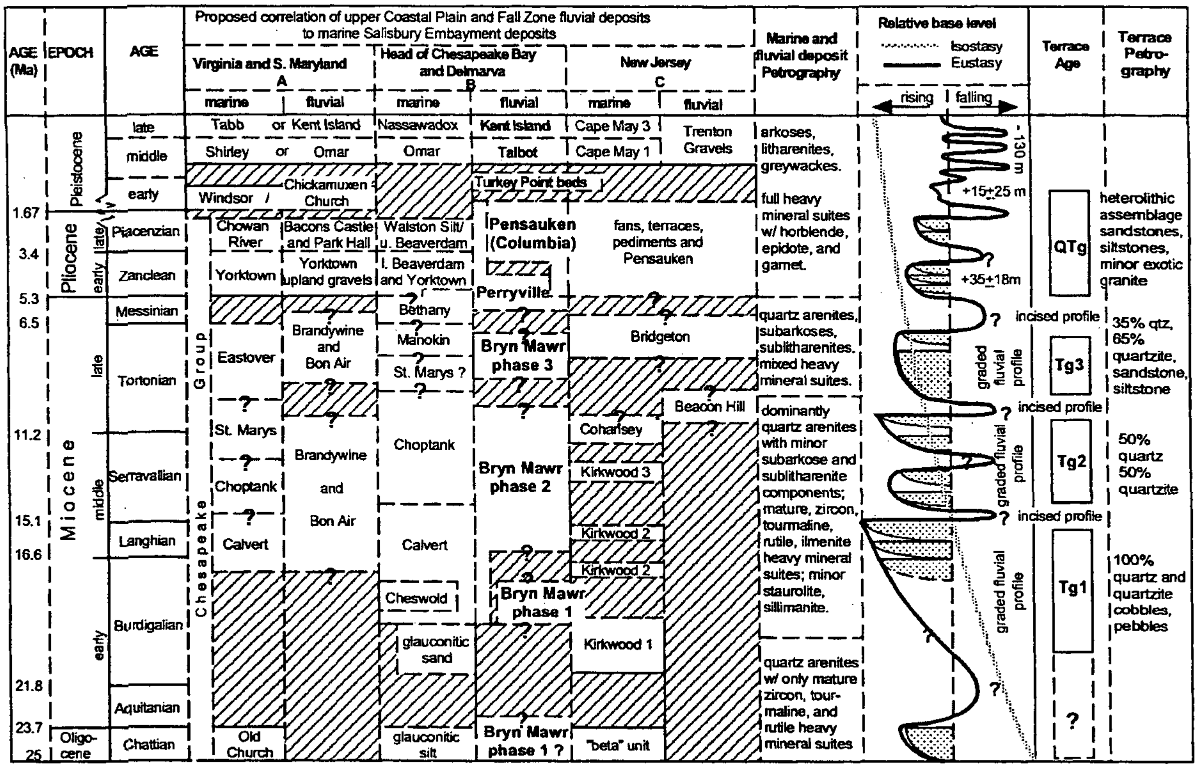

Figure FP.12 Regional correlation of upper Coastal Plain and Fall Zone fluvial stratigraphy between the James and Hudson Rivers and relation to lower Susquehanna River fluvial terraces. Stratigraphy compiled from various sources provided in Pazzaglia (1993). Modified from the Geological Society of America Bulletin, F.J. Pazzaglia. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1993 Geological Society of America. Eustasy from Haq and others, 1987; Ward and Powars, 1991; Dowsett and Cronin, 1990; Cronin and others, 1981; Greenlee and Moore, 1988).

filling the basin in a seaward thickening wedge (Owens and Gohn, 1985; Ward and Strickland, 1985). A basin-wide unconformity of middle Oligocene age (Ward and Strickland, 1985) represents a sustained period of subaerial erosion, which separates the early Cenozoic and Cretaceous deposits from the overlying Chesapeake Group.

Several upper Coastal Plain and Fall Zone fluvial deposits, like those present at the mouth of the Susquehanna River, represent the proximal, updip equivalents to the marine Chesapeake Group facies (McGee, 1888; Jordan, 1964; 1974; Owens and Minard, 1979; Owens and Denny, 1979; Figures FP.11 and FP.12). Petrography-based lithostratigraphic correlations along the Fall Zone and downdip into the Salisbury Embayment establish a regional chronostratigraphic framework for these fluvial deposits (Pazzaglia, 1993).

Time Lines. We correlate four stratigraphic horizons along cross section A-A' (Figure FP.1) and use them as time lines (Figure FP.13) to constrain the progressive flexural deformation of the margin to offshore sediment loading and continental denudation. The present elevation of a stratigraphic time line, with respect to modern sea level ($E_{TL}(x)$) is the sum of original land-surface or depositional gradient ($E_{TL0}(x)$), the change in eustatic sea level (DSL), and change in elevation attributed to isostatic deformation (I(x)):

$$E_{TL}(x) = E_{TL0}(x) + DSL + I(x). \quad (1)$$

Slope and change in eustatic sea level can be obtained by regional geomorphic and stratigraphic relationships and from published sources (Pazzaglia and Gardner, 1994). Values for isostatic deformation will be generated by the geodynamic model.

4.4 Geodynamic Models

Geodynamic models simulate flexural deformation assuming that the U.S. Atlantic passive margin is in isostatic equilibrium (Karner and Watts, 1982), time lines can be reconstructed from geologic and geomorphic data, and the passive margin lithosphere can be simulated as a uniformly thick, perfectly elastic plate, without horizontal stresses, that will respond flexurally to strike-averaged, vertically-applied loads.

The one-dimensional model (Figure FP.14) treats the passive margin lithosphere as an infinite, unbroken, elastic plate of uniform thickness. The one-dimension approximation of flexure for a thin, unbroken elastic plate under a line load is given in Turcott and Schubert (1982) and outlined in Pazzaglia and Gardner (1994); the interested reader is referred to these sources for a complete mathematical representation. The one-dimensional model is composed of 17 equally spaced, 50-km-wide cells, aligned parallel to cross section A-A' (Pazzaglia and Gardner, 1994; Figure FP.14). The unbroken, elastic plate is allowed to flex under vertical stresses applied as positive sediment loads offshore, and negative erosional loads for the continent. We run four models, using as input, offshore sediment loads and negative erosional loads for the upper Oligocene-lower Miocene, middle Miocene, upper Miocene, and Plio-Pleistocene. Flexural deformation of the plate under these loads is then compared to stratigraphic horizons corresponding to each of the four time lines (Figure FP.13).

The two dimensional model assumes a similar infinite elastic plate of finite, uniform thickness, but differs from the one-dimensional model in that vertical loads are applied (offshore sediments) or removed (onshore denudation) across a two-dimensional gridded data set. The gridded vertical load data is assembled from isopach maps of offshore sediment (Poag, 1992), and for spatially uniform subaerial denudation at a rate of 10 m/m.y. The flexural response to the gridded vertical loads is calculated by the software package GMT.

4.5 Results

The results of the geodynamic models have already been presented and discussed in detail in Pazzaglia and Gardner (1994) and Pazzaglia and Gardner (1998). Here we provide a short synopsis of these results that serves as a basis for understanding long-wavelength flexural deformation of the margin and generation of contemporary margin stress. The one-dimensional results show the best fit to the stratigraphic time lines occurs when the model is parameterized by a flexural rigidity of $4*10^{23}$ Nm (effective elastic thickness, h, of 40 km) and an erosion rate of 10 m/m.y. (Figure FP.15). These results agree with several geophysically based studies for which estimate the elastic thickness of the lithosphere underlying the U.S. passive margin ranges between 20 and 60 km (Karner and Watts, 1982; Bechtel et al., 1990). The best fit for our one-dimension models was calculated as the sum of the squares of the residuals (ΣR^2) between the model-generated flexural profiles (thick, dark lines in Figure FP.15) and the stratigraphic time lines (shaded background and terrace data in Figure FP.20). When the one-dimensional model was

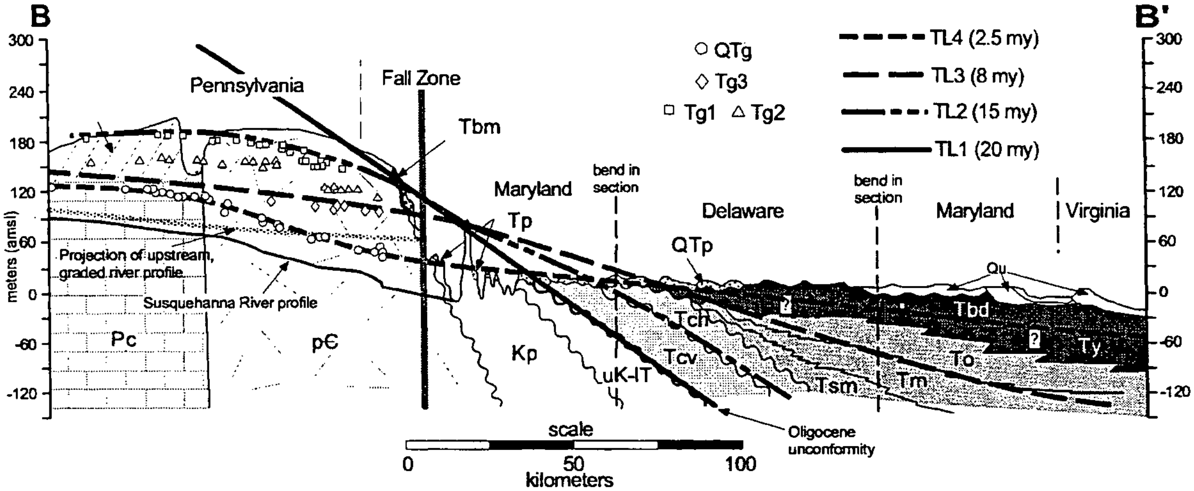

Figure FP.13 Cross section B-B' of Figure FP.1a showing the four stratigraphic time lines constructed by correlation of Susquehanna River terraces (Figure FP.14) through Inner Coastal Plain and Fall Zone fluvial deposits (Figure FP.16) and into Coastal Plain deposits of known age in the Salisbury Embayment (Modified from Pazzaglia and Gardner, 1994). pC = Precambrian bedrock, Pc = Conestoga Group, Pch = Chilhowee Group, Kp = Cretaceous Potomac Group, Tbm = Bryn Mawr Formation, Tp = Perryville formation, Qtp = Pensauken Formation, lT = lower Tertiary deposits, Tcv = Calvert Formation, Tch = Choptank Formation, Tsm = St. Marys Formation, Tm = Manokin formation, Tb = Bethany formation, Tbd = Beaverdam Formation, Ty = Yorktown Formation, Qu = Quaternary deposits undivided.

parameterized with higher flexural rigidities (effectively thicker elastic plates), the flexural profiles tended to flatten. Conversely, with lower flexural rigidities (effectively thinner elastic plates), the flexural profiles steepen. In either case, these profiles produced rather poor fits to the Coastal Plain portions of the stratigraphic time lines. In contrast, erosion rates strongly influence the fit of the flexural profiles to the terrace portion of the stratigraphic time lines. Variance in the erosion rates of just 5 m/m.y. produced flexural profiles either tens of meters above or tens of meters below the mapped terrace elevations. The sensitivity of this simple one-dimensional model is such that an effective elastic thickness of 40 km and an erosion rate of 10 m/m.y. represents a rather unequivocal, and geologically reasonable parameterization for flexure of the middle Atlantic margin. The total amount of rock uplift through the Pennsylvania Piedmont in the past 15 m.y. ranges from 35 m at the Fall Zone to 130 m at the Great Valley.

The one-dimensional model captures, to a first-order, flexural deformation of the Atlantic margin as long as the flexural profiles are aligned orthogonal to the coast and in or near the center of Coastal Plain depositional basins such as the Salisbury Embayment (i.e. the cross section of Figure FP.1). Application of the one-dimensional model on a cross section coincident with a known Coastal Plain arch (such as the Norfolk arch of Figure FP.1 and FP.6) could not produce flexural profiles that closely matched stratigraphic

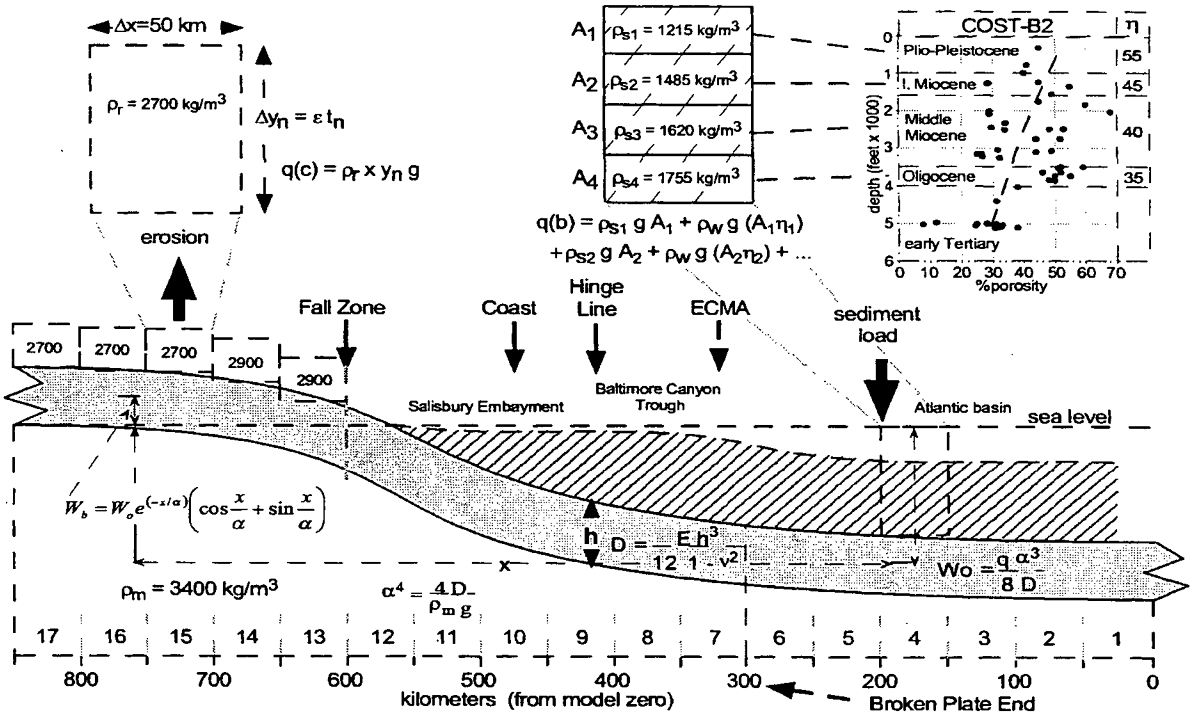

Figure FP.14 Geodynamic model (from Pazzaglia and Gardner, 1994).

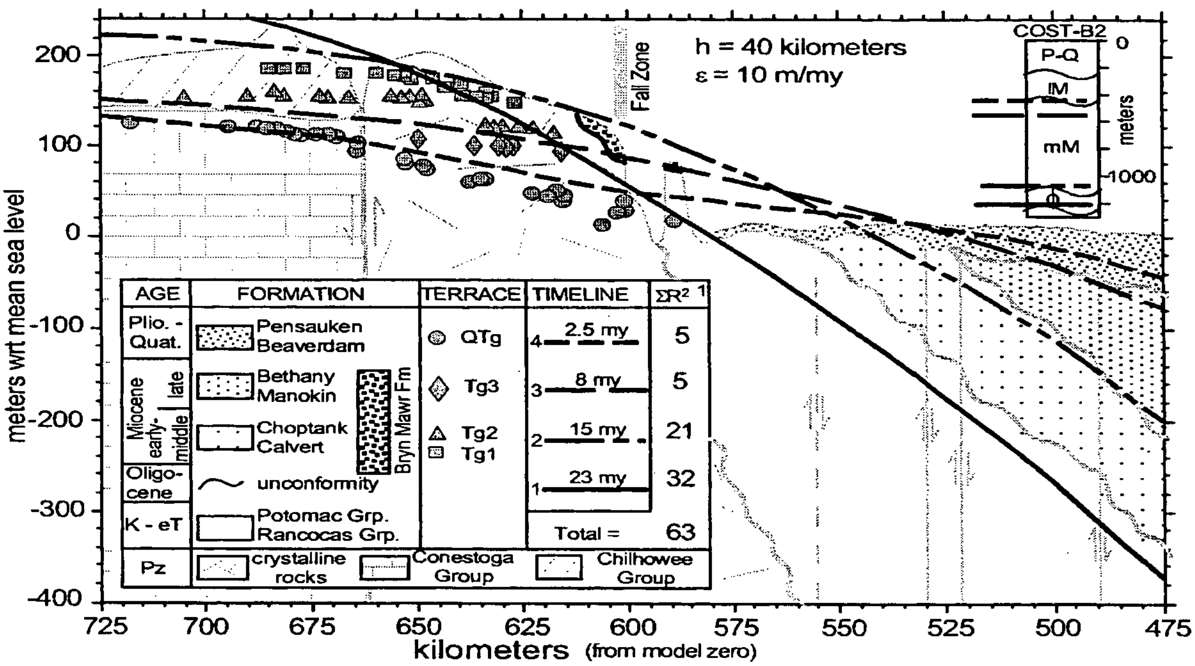

Figure FP.15 Model results for an erosion rate equal to 10 m/m.y. and plate elastic thickness equal to 40 km.

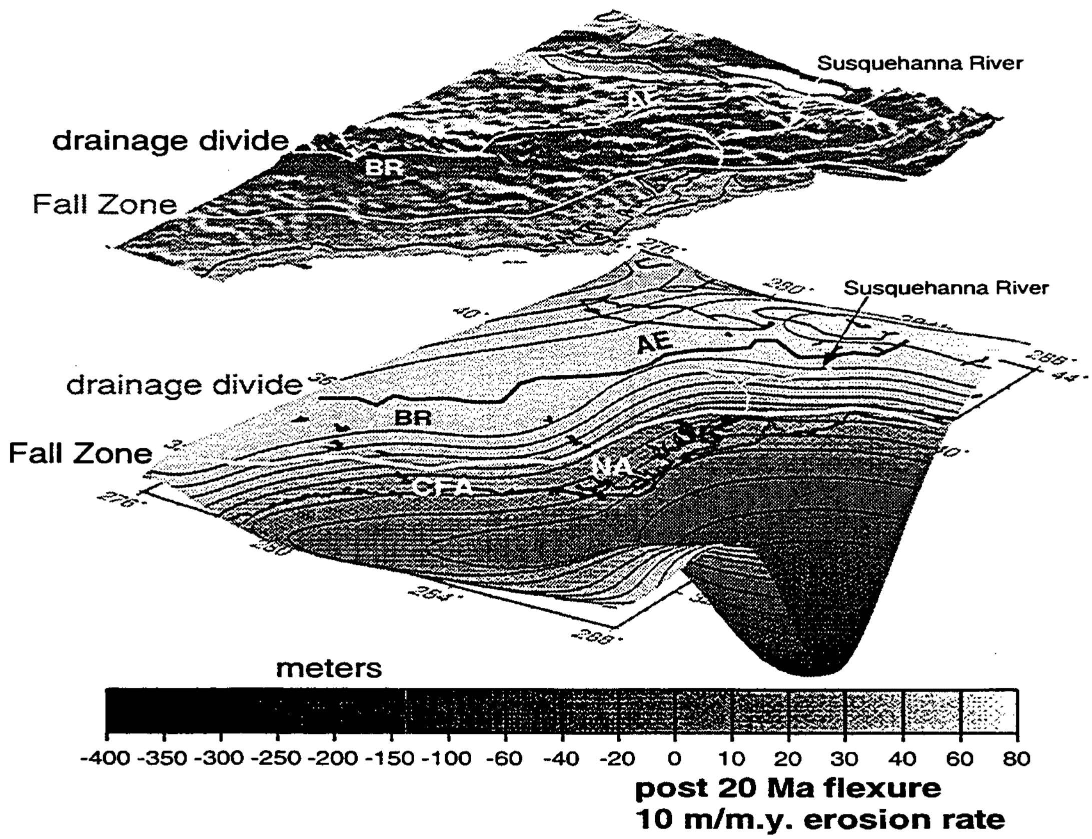

Figure FP.16 Two-dimensional flexural model of the U.S. Atlantic passive margin. a) Oblique digital shaded relief of Appalachian topography for reference. b) Model results where D = $4*10^{23}$ Nm, and erosion rate = 10 m/m.y. Notice the steep flexural gradients landward of the Salisbury Embayment. Stresses related to this flexure may be a source for earthquakes in the Susquehanna drainage basin (see location in (a) above). CFA = Cape Fear Arch, NA = Norfolk Arch, AE = Allegheny Escarpment, BR = Blue Ridge Escarpment, SE = Salisbury Embayment. Modified from Pazzaglia and Gardner, Late Cenozoic Large-Scale Landscape Evolution of the U.S. Atlantic Passive Margin, M. Summerfield, ed., Geomorphology and Global Tectonics, Copyright © 1998. Reprinted by permission of John Wiley & Sons, Ltd.

timelines. The obvious implications of this observation are that (1) the arches are not the product of margin flexural deformation, but rather reflect tectonic or dynamic topography processes, or (2) the arches are in fact related to flexural deformation that simply cannot be captured by a simple, one-dimensional line load model. Shorter wavelength structural features, such as the Norfolk arch, are better addressed with the two-dimensional model which considers a much larger portion of the margin and the combined effects of offshore sediment loading for both the middle and southern Atlantic margin.

The two-dimensional flexural model (Figure FP.16), parameterized with a flexural rigidity of $4*10^{23}$ Nm and an erosion rate of 10 m/m.y (from Figure FP.15) provides a first-order estimate for the amount of post-late Oligocene rock uplift and margin deformation that could be attributed to flexural processes. The effects of continental denudation are important in that up to 80 m of rock uplift is predicted for a broad region stretching across the Piedmont and into the Blue Ridge for the southern portion of the margin, or into the Ridge and Valley for the more central portion of the margin. The flexural hinge coincident with the Fall Zone is exacerbated by an overall steeper profile. The flexural hinge is seen as a very abrupt and steep feature for the Salisbury Embayment, and a broader, more gentle feature for the southern portion of the margin. At the only location for which there exist well constrained, stratigraphic time lines (cross section A-A', Figure FP.1 and Figure FP.13), the two dimensional model comes within a factor of 2 of predicting the observed subsidence of Coastal Plain sediments and uplift of Piedmont fluvial terraces. The most important implication of these preliminary two dimensional modeling efforts is that the steep flexural gradients observed landward of the Salisbury Embayment may be a source of lithospheric stress, capable of generating earthquakes in the Susquehanna River basin.

4.6 Summary and Conclusions

Stratified and unstratified sand and gravel terrace deposits flanking the lower Susquehanna River have been remapped on the basis of petrographic characteristics and correlated with deposits of known age at the head of Chesapeake Bay. Reconstructed longitudinal terrace profiles provided data for developing a model of terrace genesis controlled by erosion, isostatic uplift, and eustatic fluctuations. This model supports flexural isostasy as the first-order late Cenozoic, passive-margin, deformation mechanism and allows for the reconstruction of the lower Susquehanna River history for the past 20 my. Total uplift of the central Appalachian Piedmont, determined by terrace age and correlation, is at least 120 m since the middle Miocene (15 ma) resulting in a long-term uplift rate of 8 m/my. Long-wavelength flexural bending is a source of complex compressive and tensional stress in the crust of the U.S. Atlantic margin. In addition to the overall northeast-directed compressional stress (Zoback and Zoback, 1989), flexural isostasy plays an important, but poorly-quantified role in partitioning of these stresses along margin structural anisotropies such as lineaments and faults. Contemporary seismicity of the U.S. Atlantic margin may be related to the complex interaction between the regional and flexural stress field and pre-existing faults.

5 RECOMMENDED APPROACHES FOR ASSESSING ACTIVE TECTONICS IN THE PASSIVE MARGIN SETTING

- The assessment of active tectonics on the U.S. Atlantic passive margin should include tectonic geomorphologic studies focused on:

(1) The fluvial system, specifically terrace stratigraphy.

(2) The hydrologic system, specifically linear traces of spring travertine.

(3) Regional considerations of epeirogenically uplifted topography and theconcomitant denudational response.

- Conceptual models for how strain is apportioned and accommodated in a passive margin should be constructed, following from the examples of similar models developed for active margins.

- Active faults are probably associated with area of late Cenozoic tectonic deformation. These faults are likely reactived Appalachian faults, fractures, and lineaments. An effort should be made to consider how existing faults will accommodate the:

(4) Regional, northeast-oriented compressive stress field

(5) The local stress field concentrated at the Fall Zone by margin flexure

- Tectonic geomorphologic analyses, in concert with structural, geophysical, and other analyses, should investigate potential seismogenic sources not previously identified as such. These sources include, but are not limited to:

(6) Karst and non-karst related features in the Great Valley and Ridge and Valley.

(7) Lineaments transverse to the Appalachian structural grain.

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Table FP.2 Clast identifications[1] for lower Susquehanna River terraces and selected upper Coastal Plain and Fall Zone fluvial deposits

Terrace or Coastal Plain deposit	Name and Location[2]	Vein quartz	Meta quartz[3]	Quartzite	Sandstone, siltstone, chert, limestone	Ironstone, granite, gneiss, schist
	Qt4, at 3 km south of Washington	3.5	-	19	74	3.5
lower terraces	Qt4, at Marietta (Peltier, 1949)	2	-	12	81	5
(Qt1 - Qt6)	Qt2, at Highspire	1	2	31	64	2
	Qt2, at Marietta (Peltier, 1949)	5	-	17	75.5	2.5
and	Qt1?, at Coudon Farm	15	40.5	30.5	12	2
	Pensauken, at Turkey Point	18.5	39.5	16	22.5	3.5
Pensauken Fm.	Pensauken, at Turkey Point	7	20	10	54	9
	Pensauken on Delmarva (Jordan, 1964)	46	-	36	16	2
lower terraces	QTg, at Broad Creek	16	16	41	24	3
(QTg)	QTg, combined count for entire terrace	25	12	37.5	25	<1
and	Perryville, at Mountain Hill	12	45	41	1	1
Perryville fm.	Perryville, north of Havre de Grace	10	15	31.5	31.5	12
Upland terraces (Tg1 - Tg3)	Tg3, at Brinton Farm, Lancaster Co., PA	16	21	52.5	10.5	<1
	Tg2, combined count for entire terrace	26	15	55	4	<1
and	Tg1, at Kirk Farm, Lancster Co., PA	33	51.5	14.5	<1	1
	Tg1, combined count for entire terrace	5	55	30	9	1
Bryn Mawr Fm	Bryn Mawr, at York quarry, (2 - 4 cm)	18.5	42	37.5	2	<1
	Bryn Mawr, at York quarry, (4-10 cm)	25	30	40	4	1
	Bryn Mawr, Fall Zone, (Owens, 1969)	95	-	<1	3	2

Table FP.2 Clast identifications[1] for lower Susquehanna River terraces and selected upper Coastal Plain and Fall Zone fluvial deposits (continued)

Bryn Mawr, on Elk Neck Peninsula	29	61	2	5	3

1 = clast size: 2-8 cm unless otherwise specified
2 = see Figure FP.3 for locations.
3 = dash means that vein and metamorphic quartz were not separately identified. Total non-quartzite quartz is listed in the vein quartz column.

Table FP.3 Compositional, textural, and weathering characteristics of lower *Susquehanna River terraces.* Data compiled from this study and from Peltier (1949).

Terrace	Composition	*Shape*	*Size* (cm)	*Soil characteristics*[1]	Oxidation depth	Clast rind thickness
Qt6				10 YR Bw: 0.1 m	1	0
Qt5				10 YR Bw: 0.3 m	< 5	0
Qt4					4 - 6	0 - 0.25
Qt3	heterolithic; at least 50% quartzite, sandstone, and siltstone; up to 3% granite and gneiss	subangular to well-rounded	2 - 200	7.5 YR Bt: 0.5 m		
Qt2				- 7.5 YR Bt1, 5 YR and 2.5 YR 2Btb: 1 - 1.5 m*	- 6 m	- 0.5 - 1 infrequent saprolitized clasts
Qt1				-	-	0.5 - > 1 frequent saprolitzed clasts
QTg	30% vein and meta quartz 50% quartzite and sandstone 20% red siltstone and chert	subangular to well-rounded	2 - 8 with boulders 1 meter across	2.5 YR Btb: 6 m*	> 6 m	> 1 frequent saprolitized clasts
Tg3	35% vein and meta quartz 65% quartzite, sandstone, red siltstone, and chert	subrounded to well-rounded	2 - 20 with cobbles up to 50 cm	-	-	white, leached rinds on quartzites 0.25 to 1 cm
Tg2	50% vein and meta quartz 50% quartzite and sandstone	well-rounded	2 - 20 with cobbles up to 50 cm	-	-	white, leached rinds on quartzites 0.25 to 1 cm
Tg1	75% vein and meta quartz 25% quartzite	well-rounded	2 - 20 with cobbles up to 50 cm	10 YR Bt1, 7.5 YR 2Btb, 5 YR 3Btb*	at least 3 m	white, leached rinds on quartzites 0.25 to 1 cm

1 = soil symbols: B = illuvial zone; w = cambic horizon; t = illuvial clay; 2, 3 = change in parent material; b = buried (Soil Survey Staff, 1975).
* = soil developed in colluvial roundstone diamictons derived from originally-stratified terrace gravel

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PALEOSEISMOLOGY AND SEISMIC HAZARDS EVALUATIONS IN EXTENSIONAL VOLCANIC TERRAINS

by

Richard P. Smith and Suzette M. Jackson[1]
William R. Hackett[2]

1 INTRODUCTION

Magma intrusion is an important component of worldwide crustal extension (Parsons and Thompson 1991; Gans, 1987; Coney, 1987; Forslund and Gudmundsson, 1991; Lepine and Hirn, 1992; Rubin and Pollard, 1988). Seismicity, surface faulting, magma intrusion and volcanism are expressed within many tectonic settings, and extension of the brittle crust is accommodated by a combination of normal faulting and magmatic (dike intrusion) processes (Bursik and Sieh, 1989; Parsons and Thompson, 1991, 1993; Forslund and Gudmundsson, 1991). Although areas of ambiguous or complex interplay of tectonic and magmatic processes exist, in this paper we emphasize faulting and seismicity that seem clearly associated with magma intrusion, particularly dikes. The dike intrusion process is emphasized because dike intrusion is a widespread process within the upper crust, especially in extensional settings (Emerman and Marrett, 1990; McKenzie et al., 1992).

Because earthquakes are commonly associated with magma intrusion, volcanic zones must be considered as potential seismic sources in seismic hazards assessments. However, the mechanism of formation of magma-induced fault scarps and their post-deformational geomorphic evolution differ from those of tectonic normal faults. As a result, conventional paleoseismic approaches based on excavation of surficial sediments along fault zones and investigation of colluvial wedges may be inadequate to furnish earthquake recurrence and magnitude information, and may therefore lead to inaccurate conclusions. Innovative approaches are required; approaches that emphasize the geochronology of cogenetic volcanic materials for recurrence estimation, and analogy with historically active volcanic rift zones and field measurements in the region of interest for maximum magnitude estimation.

In this paper, we suggest some useful approaches and field methods for estimation of recurrence and maximum magnitude of seismicity in extensional volcanic terrains. A more complete discussion of the approaches and methods is provided in Smith et al. (1996) and Hackett et al. (1996). The described approaches can be applied to extensional volcanic terrains worldwide if field characteristics indicate that dike intrusion is the driving mechanism of deformation. The principles presented in this paper can affect the outcome of studies of these terrains in two ways. First, the paleoseismic investigations will be focused towards techniques that provide the greatest chance of establishing meaningful recurrence and magnitude data. And second, the assessed level of seismic hazard will likely be lower (lesser maximum magnitudes and perhaps longer recurrence intervals) than would be indicated by techniques that

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overlook the role of magma intrusion in the faulting and fissuring process. We report some preliminary results of their application in the eastern Snake River Plain of Idaho, and their significance to seismic hazards assessment at the Idaho National Engineering Laboratory.

2 EXTENSIONAL VOLCANIC STRUCTURES

2.1 Process

When magma pressure in the upper part of a conduit or reservoir exceeds the strength of the surrounding rocks, blade-like dikes propagate outward from the reservoir along self-generated fractures (Rubin and Pollard, 1987) at depths of 2 to 4 km (Ryan, 1987; Gudmundsson, 1984). These dikes have heights (vertical dimensions) of several kilometers, and lengths that may extend tens of kilometers from the magma conduit. Dikes intrude normal to the direction of least-compressive stress, which is controlled by the regional extension direction in volcanic provinces. Repetition of the dike intrusion process forms subsurface dike swarms and complex, overlapping surface-deformation features.

Volcanic rift zones are the surface expression of dike swarms, and dike intrusion causes extensional deformation and seismicity in volcanic rift zones. Supporting evidence includes surface geodetic measurements, field observations, and the characteristics of co-intrusive seismicity during volcanic cycles (Klein et al., 1987; Einarsson, 1991), exposed dike swarms in deeply eroded volcanic rift zones (Walker 1987; Gudmundsson 1983, 1984), drilling data (Eichelberger et al., 1985), geophysical anomalies (Schoenharting and Palmason, 1982; Flanigan and Long, 1987), the characteristics of volcanism along eruptive fissures (Macdonald, 1972), and alignment of volcanic vents along fissures (Hackett and Smith, 1992). Migration of seismicity, geodetic changes, and fault offsets observed during and after dike intrusion events in Iceland and Hawaii (Einarsson and Brandsdottir, 1980; Karpin and Thurber, 1987) conclusively link the dike-intrusion process to seismic activity. Dikes propagating at rates of 0.03-1.7 m/s (Hauksson, 1983; Klein et al., 1987; Einarsson, 1991) incrementally form normal faults and fissures and give rise to swarms of small, shallow earthquakes that migrate in concert with advancing dike-tips.

Numerical modeling of elastic media and geodetic monitoring of regions above intruding dikes provide an understanding of the relationship of dike geometry to stress and strain distributions and to surface deformation (Figure RS.1) (Tarantola et al., 1979; Sigurdsson 1980; Pollard et al., 1983; Tryggvason, 1984; Marquart and Jacoby, 1985; Rubin and Pollard, 1988; Mastin and Pollard, 1988; Stein et al., 1991; Rubin, 1992, 1993; Roth, 1993). Dike intrusion produces a broad zone of uplift (up to 10 km wide), with a narrow zone of subsidence centered above the propagating dike. The broad uplift results from the magma-induced compressional stress field alongside the dike, and the narrow zone of subsidence above the dike results from the tensile stress developed there (Figure RS.1A). The normal faults and fissures that develop where the tensile zone interacts with the earth's surface (Rubin, 1992) have also been investigated by physical analog models (Mastin and Pollard, 1988). In these experiments, fissures appear in two symmetrical zones above the simulated dike (Figure RS.1B), and form progressively inward with dilation of the dike. Most fissures are oriented parallel to the dike plane. As dilation continues, dip-slip movement occurs on the fissures, producing normal faults and a central graben. Nearly vertical normal faults at the surface decrease to 70-degree dips at depth. The down-dip extent of faults is small; they extend only slightly deeper than the dike top. The total horizontal

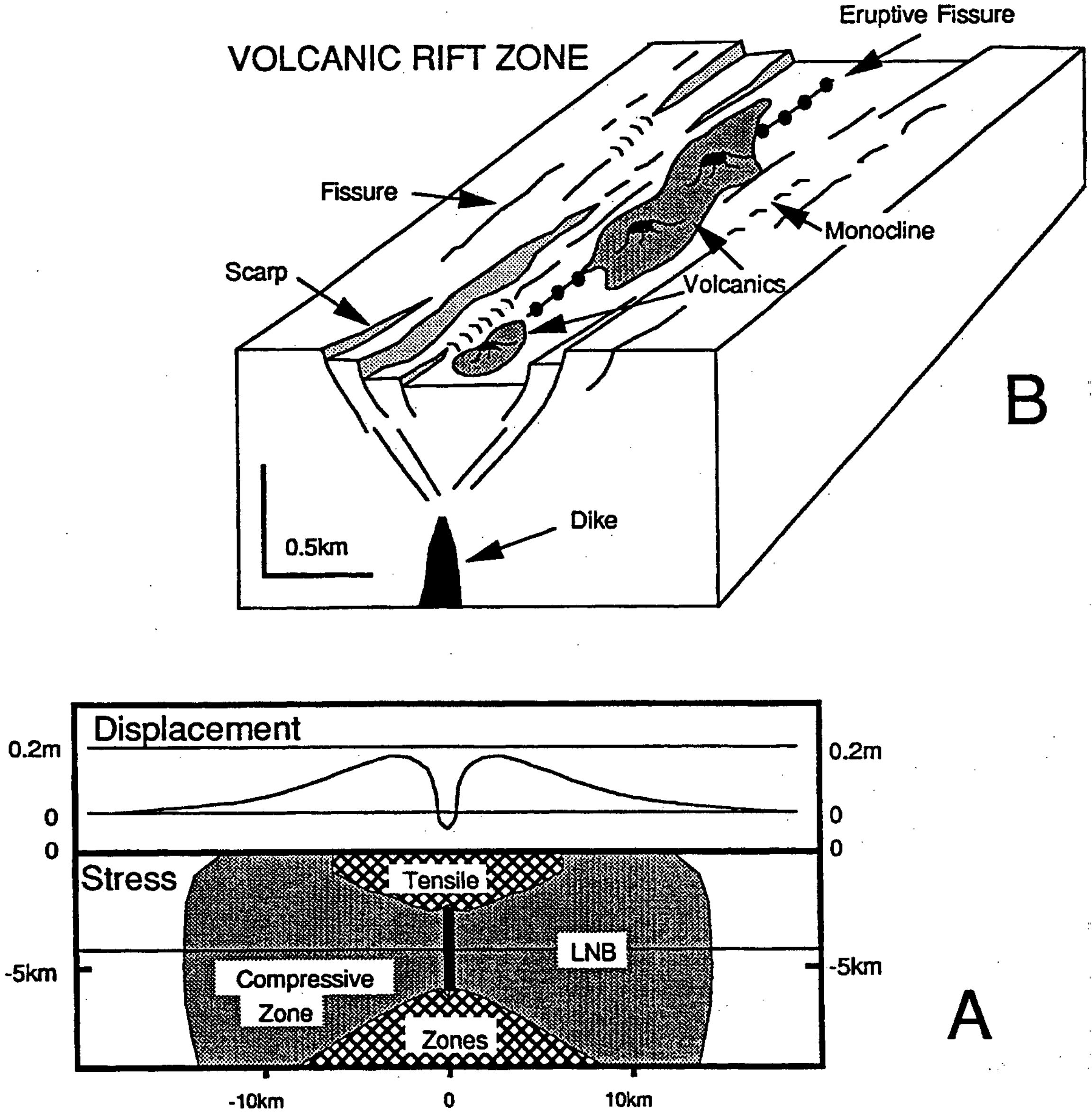

Figure RS.1 Schematic diagrams summarizing the configurations of extensional volcanic structures and their relationship to magmatic processes. (A) Results of numerical elastic deformation model of dike intrusion (Rubin, 1992) are consistent with observed brittle deformation features shown in (B). Upper part of the diagram shows a vertical displacement profile above a dike of 1meter thickness, extending from 1 - 6 km depth. Lower part of the diagram shows the compressive and extensional zones that develop around a dike as a result of magma pressure. LNB = level of neutral buoyancy. (B) Block diagram showing dike-induced structures along a volcanic rift zone. From Smith et al., "Paleoseismology and Seismic Hazards Evaluations in Extensional Volcanic Terrains," *Journal of Geophysical Research,* 101-B3:6277-6292, 1996, © by the American Geophysical Union. Used with permission.

component of strain (i.e., extension on faults and fissures) is about 60% to 75% of the dike thickness, and the width of the extensional zone at the surface is related to dike depth. The inelastic structures of the physical models are generally consistent with: (1) the elastic-strain profiles of numerical experiments (Rubin, 1992, 1993); (2) geodetic-inversion models showing that faults extend only above and ahead of propagating dikes (Du and Aydin, 1992); and (3) field, geodetic and seismic observations on active volcanic-rift zones (Bjornsson et al., 1977, Forslund and Gudmundsson, 1991; Gudmundsson, 1984, 1987; Rubin and Pollard, 1988; Rubin, 1990; Mastin and Pollard, 1988).

2.2 Characteristics/Geometry

Volcanic rift zones are characterized by linear belts of eruptive fissures, tensile fissures, normal faults, flexural monoclines, and graben (Stein et al., 1991; Sigurdsson, 1980; Rubin and Pollard, 1988; Rubin, 1990, 1992; Pollard et al., 1983; Nairn and Cole, 1981; Mastin and Pollard, 1988; Hackett and Smith, 1992). The largest volcanic rift systems (mid-ocean ridges) are thousands of kilometers long by a few tens of kilometers wide. The smallest rift systems are a few kilometers long by several hundred meters wide and are marked by isolated, monogenetic volcanoes fed by single dikes. The Kings Bowl lava field in the eastern Snake River Plain of Idaho (Kuntz et al., 1992) is an example of such a monogenetic system. Eruptive fissures, places where dikes breach the surface, are commonly marked by aligned vents and spatter ramparts that cover all or part of the fissure. Eruptive fissures are commonly flanked on each side by fissure swarms and/or graben faults (Figure RS.1B). Tensile cracks, or fissures, tend to occur in swarms that range up to ~10 km wide and several tens of kilometers long. Individual fissures vary from a few meters to several kilometers long and from a few centimeters to several meters wide. Fissures are formed by purely tensional processes, and the fine-scale configuration of their walls is commonly controlled by pre-existing joint patterns in the host rocks. Monoclines and vertical scarps are the surface expressions of normal faults. Monoclines tend to form where thin lava flows or well developed horizontally jointed slabs drape over the fault. In places where thick lava flows and/or vertical jointing predominate, vertical fault scarps form. Slickensides or other structural indications of shear do not typically occur on these fault scarps because fissuring separates the walls of the fracture before vertical offset occurs. Most dike-induced fault scarps are <1 m high, but emplacement of thick dikes or reactivation of preexisting structures by multiple dike intrusions may produce fault scarps >10 m in height.

Graben in volcanic rift zones usually have symmetrical vertical offsets and commonly occur symmetrically disposed about a central eruptive fissure (Figure RS.1B). Worldwide, graben range up to 10 km in length and from 0.4 to ~3 km in width (Hackett et al., 1996; Jackson, 1994). Vertical offsets of eastern Snake River Plain graben do not exceed about 10 m, but reach several tens of meters in some volcanic rift zones, such as those of Iceland and Asal, where numerous dikes have been injected (Gudmundsson, 1983, 1987; Stein et al., 1991).

3 CRITERIA TO DIFFERENTIATE DIKE-INDUCED STRUCTURES FROM TECTONIC FAULTS

Although discrete, dike-induced structures are morphologically similar to non-magmatic extensional structures, several criteria can be used to recognize them in the field.

3.1 Geologic/Stratigraphic Criteria

Association With Cogenetic Volcanic Deposits. The best criterion is the inferred or demonstrated relationship to cogenetic volcanic materials (Figure

RS.1B). In near-vent areas, structures formed by dike intrusion are partially to completely buried by cogenetic volcanic products (usually lavas) when the dike breaches the surface.

Small Fault Displacements. Vertical offsets on faults range from less than 1 m, when associated with single basaltic dikes, to several tens of meters, when associated with thick (silicic) dikes or with repeated injection of closely spaced basaltic dikes.

Variations In Along Strike Displacements/Fault Segmentation. Vertical displacements typically vary abruptly along strike, and individual faults are short (hundreds of meters to about 10 km), commonly grading into monoclines or purely tensional fissures. In contrast, vertical displacements on tectonic normal faults and fault segments usually vary systematically along strike (Dawers et al., 1993; Wu, 1993; Bruhn and Wu, 1993) and individual fault segments are commonly longer than 20 km.

3.2 Geomorphologic Criteria

Subdued Topography. On the regional scale, extensional magmatism produces diffuse belts of volcanism, fissuring and subdued normal fault scarps. Even after millions of years of extension accompanied by basaltic volcanism and dike intrusion as in Iceland and the eastern Snake River Plain of Idaho, the terrain is topographically subdued. This is in contrast to extensional provinces that lack substantial magma flux into the upper crust (Parsons and Thompson, 1991), where recurrent faulting is a primary mountain-building process that produces several kilometers of vertical offset, substantial topographic relief, and significant rotation of crustal blocks.

3.3 Structural Criteria

Monoclinal Folding. Monoclinal flexures are common, reflecting offset of jointed and layered volcanic rocks, usually lava flows. Monoclines in volcanic terrains are observed to grade into vertical fault scarps along strike. Monoclinal flexures are uncommon in tectonic normal faults because they generally intersect the surface in unconsolidated alluvial fan deposits along mountain fronts. However, in places where youthful tectonic normal faults offset lava flows, similar monoclinal flexures occur, and other criteria are required to differentiate tectonic faults from dike-related deformation.

Width of Deformation Zone. Extensional volcanic structures occur in diffuse belts up to several kilometers wide instead of narrow zones, usually less than a few hundred meters wide, associated with the surface expressions of normal faults.

Symmetry of Cogenetic Volcanic Structures. A graben or two zones of non-eruptive fissures commonly occur symmetrically disposed about an eruptive fissure (Figure RS.1B). Graben are also commonly associated with extensional tectonic faults, but there they usually occur in the hanging wall block, just downslope of the main fault, and a genetic relationship to the main fault is indicated.

Dilational Features. Tensional fissures are the most abundant structure, there is little net vertical displacement across graben, and slickensides are seldom observed on fault scarps, indicating that most of the deformation is purely dilational. In contrast, normal faults in extensional tectonic environments generally exhibit evidence for more vertical than dilational deformation.

Symmetric Geophysical Anomalies. Structures may be associated with symmetrical geophysical anomalies, such as aeromagnetic or gravity highs, that indicate the presence of subsurface dike swarms. Examples include a linear aeromagnetic anomaly associated with the northern end of the Great Rift volcanic rift zone in Idaho (See Figure RS.2), aeromagnetic anomalies associated with the Kilauea and Mauna Loa volcanic rift zones in Hawaii (Flanigan and Long, 1987), and gravity anomalies in eastern Iceland (Schoenharting and Palmason, 1982). In contrast, asymmetric geophysical anomalies reflecting accumulations of basin-fill sediments adjacent to uplifted bedrock ranges are typical of tectonic normal faults.

4 PALEOSEISMIC ANALYSIS

4.1 Implications of Mechanism of Formation

The conceptual understanding of the relationship of extensional volcanic structures to cogenetic magmatic processes and the results of numerical and physical models have important implications for paleoseismologic field investigations.

1. The magnitude of cumulative fault throw at the surface above intruding dikes, and the horizontal extension, are proportional to the dike thickness, with thicker dikes producing more pronounced graben, normal faults and fissures. Basaltic dikes are generally 1 to 6 m thick. Hence, the dike-induced vertical and horizontal strain measured in the field for single basaltic dikes should range up to several meters. Thicker dikes (rhyolitic dikes can reach thicknesses of 10 m) should produce proportionately greater strain, and this is consistent with the more pronounced structures observed above silicic dikes (with total throws of 10 m or more) relative to basaltic dikes (Mastin and Pollard, 1988).

2. Down-dip extents (fault widths) of dike-induced faults and fissures are only slightly greater than the depth to the dike-top (Rubin, 1992). Worldwide, dikes intrude at shallow depths within the crust, and dike tops usually reach <5 km from the surface (Ryan, 1987; Gudmundsson, 1984).

3. The small offsets and rupture areas of dike-induced faults, and their incremental growth in tandem with dike propagation, suggest that the magnitudes of associated earthquakes will be small to moderate.

4. Graben width is related to dike depth, such that deeper dikes produce wider zones of extension than shallower dikes. Few dike-induced graben in volcanic rift zones are greater than 2 km wide, suggesting that dikes generally do not induce surface faulting until they are within a few kilometers of the surface.

5. Colluvial wedges generally do not form because the fault scarps are commonly developed in volcanic bedrock.

6. The close genetic association of dike intrusion, volcanic rift zone structures, and volcanism, suggests that seismic recurrence can be based on the volcanic rock record rather than on deposits associated with the degradation of fault scarps in unconsolidated materials.

4.2 Estimation Of Recurrence

Recurrence information for tectonic faults is typically sought in poorly consolidated colluvial deposits that are relatively quickly modified and redistributed after fault scarp development. In contrast, volcanogenic structures are commonly developed in volcanic bedrock which is composed of strongly lithified materials that do not produce well developed colluvial deposits. In addition, many volcanogenic structures are monogenetic (non-recurrent) and have small displacements, whereas others are polygenetic and represent many decimeter-scale displacements, none of which produce colluvial wedges or other recognizable deposits. Therefore, recurrence estimation must rely on other approaches. These include the dating of cogenetic volcanic materials, the dating of eolian or alluvial sediments accumulated in fissures and along fault scarps, and the determination of the time elapsed since fissure walls and scarps were exposed to cosmic rays (see Smith et al. 1996 for a more detailed discussion of these methods).

4.2.1 Establishment of Recurrence Intervals

Because earthquakes in volcanic rift zones occur as a consequence of dike injection during magmatic cycles, earthquake recurrence can be estimated by establishing the recurrence interval of volcanic cycles. In addition to geochronologic data on volcanic rocks and structures, this requires thorough knowledge of volcanic processes and the regional patterns of volcanism, and should take into account the nature of vent clusters. Conservatism in assessments is introduced by assuming one dike injection episode for each volcanic vent. More realistic assessments are based on field observations that the injection of a single dike can produce several volcanic vents along or near its trace and that dike-intrusion events can occur without volcanism. Thus, there is some judgment involved in estimating the number of volcanic cycles. Criteria for grouping vents to provide a best-estimate of the number of dike intrusion events include temporal affinity, spatial proximity, geometric alignment, and petrographic and chemical similarities.

Geochronology of rocks in drill holes can also provide valuable constraints for volcanic recurrence intervals determined for volcanic rift zones. Borehole recurrence data may provide a check on the assumption of constant event rate because boreholes commonly furnish information that is applicable over time frames longer than those represented at the surface.

Even when precise and sufficient age determinations are available from volcanic rocks to confidently establish volcanic recurrence intervals, the information is not analogous to that established by paleoseismic studies of individual normal faults. The recurrence intervals resulting from the analysis define periods of quiescence between volcanic cycles. Seismicity associated with volcanic cycles is characterized by numerous earthquake swarms as opposed to a mainshock-aftershock sequence typical of tectonic faults. Within a volcanic cycle one or more dikes may be intruded and each will be accompanied by a migrating swarm of seismicity. Recurrence intervals between dike intrusion events within a volcanic cycle could range from hours to centuries. The intrusion of each dike will take hours to days and during that time scores to hundreds of earthquakes can occur, with recurrence intervals ranging from less than a second to several minutes.

4.3 Estimation Of Maximum Magnitude

In general the maximum magnitude of earthquakes associated with dike intrusion in volcanic rift zones should be lower than that for most tectonic earthquakes. This is because the faults and fissures are developed within the relatively shallow zone above dike tops (Figure RS.1B). Normal faults and fissures above dikes in volcanic rift zones generally have rupture areas of less than 25 km^2 (Jackson, 1994), whereas tectonic normal faults may have rupture areas of up to 1000 km^2 (Wells and Coppersmith, 1994). The normal faults and fissures produced by dike intrusion form in the shallow crust at depths less than 4-5 km, where differential stress and rigidity are lower than at depth. Cracks that nucleate at shallow depth lack sufficient strain energy to propagate to deeper depths and higher strength regions (Daz and Scholz, 1983).

Two methods are suggested for estimating the maximum magnitude of dike-induced seismicity (Jackson, 1994; Hackett et al., 1996). The first method bases maximum magnitude on analogy with the largest earthquakes observed in active volcanic rift zones worldwide. Observational seismicity indicates that the largest earthquake magnitudes associated with dike injection in volcanic rift zones range from 3.0 to 5.5 and have a mean and one standard deviation of 3.8±0.8 (Smith et al., 1996; Hackett et al., 1996). Einarsson (1991) reports that earthquakes of magnitude <4.0 are usually associated with dike injection from Krafla, whereas those with magnitudes ≥5.0 are associated with caldera deflation. The second method estimates a maximum earthquake magnitude using empirical relationships among fault dimensions and magnitudes (e.g. Wells and Coppersmith, 1994). Structural parameters, such as surface lengths of normal faults and fissures, and graben widths, can be used to estimate fault dimensions in volcanic rift zones (Jackson, 1994). Although surface dimensions can be measured directly, fault widths (down-dip extent) must be estimated by the relationship between graben width and depth to the dike top (Mastin and Pollard, 1988; Pollard et al., 1983), by geophysical methods, such as shallow high-resolution reflection profiling, by drilling, or by estimation of the depth at which dikes are likely to propagate: the level of neutral buoyancy (LNB) (Ryan, 1987) or the depth extent of low tensile strength rocks (Gudmundsson, 1984).

Wells and Coppersmith (1994) report a comprehensive empirical data set relating earthquake magnitude to a number of dimensional characteristics of tectonic faults, including surface length, subsurface length, downdip extent (width), rupture area, and fault displacement. Application of these relationships to dike-induced faults in extensional volcanic terrains suggests that some dimensional parameters are better than others for estimating observed maximum magnitudes (Figure RS.3A).

The empirical relationships for average and maximum displacement are not recommended for several reasons. First, deformation associated with earthquakes having moment magnitudes less than approximately 5.7 may be a secondary effect from ground shaking and may not be due to primary rupture of the fault (Wells and Coppersmith, 1994). Second, displacements along dike-induced normal faults generally do not form during single-event ruptures, but commonly result from many incremental displacements during dike intrusion or from multiple dike-intrusion events (Rubin, 1992; Hackett et al., 1996). Under these conditions, 1-2 meter-high scarps form as a result of several decimeter-scale displacement events over periods of minutes to hours during dike propagation. At Krafla, 1-m displacements on normal faults and extensive fissuring were observed to occur over a period of several hours during which time the largest associated earthquake had a magnitude less than 4 (Brandsdottir and Einarsson, 1979; Hauksson, 1983). In contrast, scarps of the same height along tectonic faults commonly form nearly instantaneously as large areas of the fault rupture (Bullen and Bolt, 1985; Wells and Coppersmith, 1994). And finally, abrupt fluctuations in vertical displacement along the lengths of dike-induced fractures suggest that local variations in lithologic, structural, and stress conditions exert strong control on the amount of displacement seen at the surface.

Comparison of observed vs. empirically estimated maximum magnitudes in active volcanic rift zones (Figure RS.3B) suggests that maximum magnitudes calculated using fault width are most consistent with the instrumentally recorded magnitudes that are associated with dike injection. Fault width in quiescent volcanic rift zones is best estimated from inferred depths to the tops of associated dikes. The width of graben or distance between parallel fissure swarms (Figure RS.1B) allow estimation of depth to dike tops. In the absence of graben or fissure swarm pairs, an upper-bound fault width can be obtained from estimated depth to the level of neutral buoyancy or depth extent of low tensile-strength rocks, on the order of 5 km (Ryan, 1987).

5 CASE STUDY: EASTERN SNAKE RIVER PLAIN (ESRP) OF IDAHO

5.1 Regional Setting

The ESRP is a northeast-trending linear volcanic province that traverses the northeastern part of the Basin-and-Range province of the western United States (Figure RS.2). It is a low-elevation basin of subdued relief within the mountainous terrain of the northern Basin and Range and the Northern Rocky Mountains. The concept that the ESRP represents the track of the Yellowstone mantle plume as it was over-ridden by southwestward movement of the North American tectonic plate has been advanced by numerous investigators (Morgan, 1972; Armstrong et al., 1975; Pierce and Morgan, 1992). Since passage of the hotspot from the ESRP at about 4 Ma, the Plain has been the site of persistent basaltic volcanism (Kuntz et al., 1992) which has filled the subsiding basin with >1 km of basalt lava flows (Hackett and Smith, 1992). Much of the volcanism has emanated from volcanic rift zones that trend north-northwest across the ESRP, consistent with the east-northeast direction of basin-and-range extension in the region (Zoback and Zoback, 1989; Pierce and Morgan, 1992). In contrast to the surrounding Basin and Range, most of the extension on the ESRP has been accommodated by injection of basaltic dikes beneath the volcanic rift zones (Rodgers et al., 1990), a process that reduces differential stresses and supplants tectonic normal faulting and associated large earthquakes (Parsons and Thompson, 1991; Hackett and Smith, 1992).

The Idaho National Engineering Laboratory (INEL) is located on the ESRP, near its boundary with the northern Basin and Range province (Figure RS.2). Active basin and range normal faults die out near the margin of the ESRP and volcanic rift zones occur on the ESRP. Since tectonic normal faults and volcanic rift zones occur near or on the INEL, seismic hazards assessments must consider both as seismic sources. Additional sources are also considered, including "random" or background seismicity in both the Basin and Range province and the ESRP. The seismic hazard from the tectonic normal faults is addressed by trenching investigations, mapping of Quaternary deposits along the faults, and detailed structural mapping of the faults (Hemphill-Haley et al., 1994; Olig et al., 1994; Knuepfer, 1994; Bruhn and Wu, 1993; Wu, 1993). In order to assess the seismic hazard from volcanic rift zones, we have used the methods previously described for estimating both the maximum magnitude and the recurrence of volcanic rift zone seismicity.

5.2 Maximum Magnitude

A M 5.5 is appropriate for the maximum earthquake magnitude which could be associated with future dike intrusion in ESRP volcanic rift zones. Maximum magnitudes calculated using the dimensions of normal faults and fissures in ESRP volcanic rift zones, and the empirical relationships of Wells and Coppersmith (1994) are shown in Table RS.1. Maximum magnitude estimates, with means ±1 standard deviation, are tabulated for a variety of fault dimensional parameters, including surface length, fault width based on estimated depths to the dike tops and to the level of neutral buoyancy (LNB), and fault area based on surface length multiplied by depth to the dike top and to the LNB.

The maximum magnitudes for fault width and rupture area based on estimated depth to dike tops have the most significance because available information suggests that they are the best estimators of maximum magnitude in active volcanic rift zones (Figure RS.3). To account for the possibility of single-event ruptures along normal faults or fissures with relatively large fault areas (e.g. 6-7 km surface lengths with

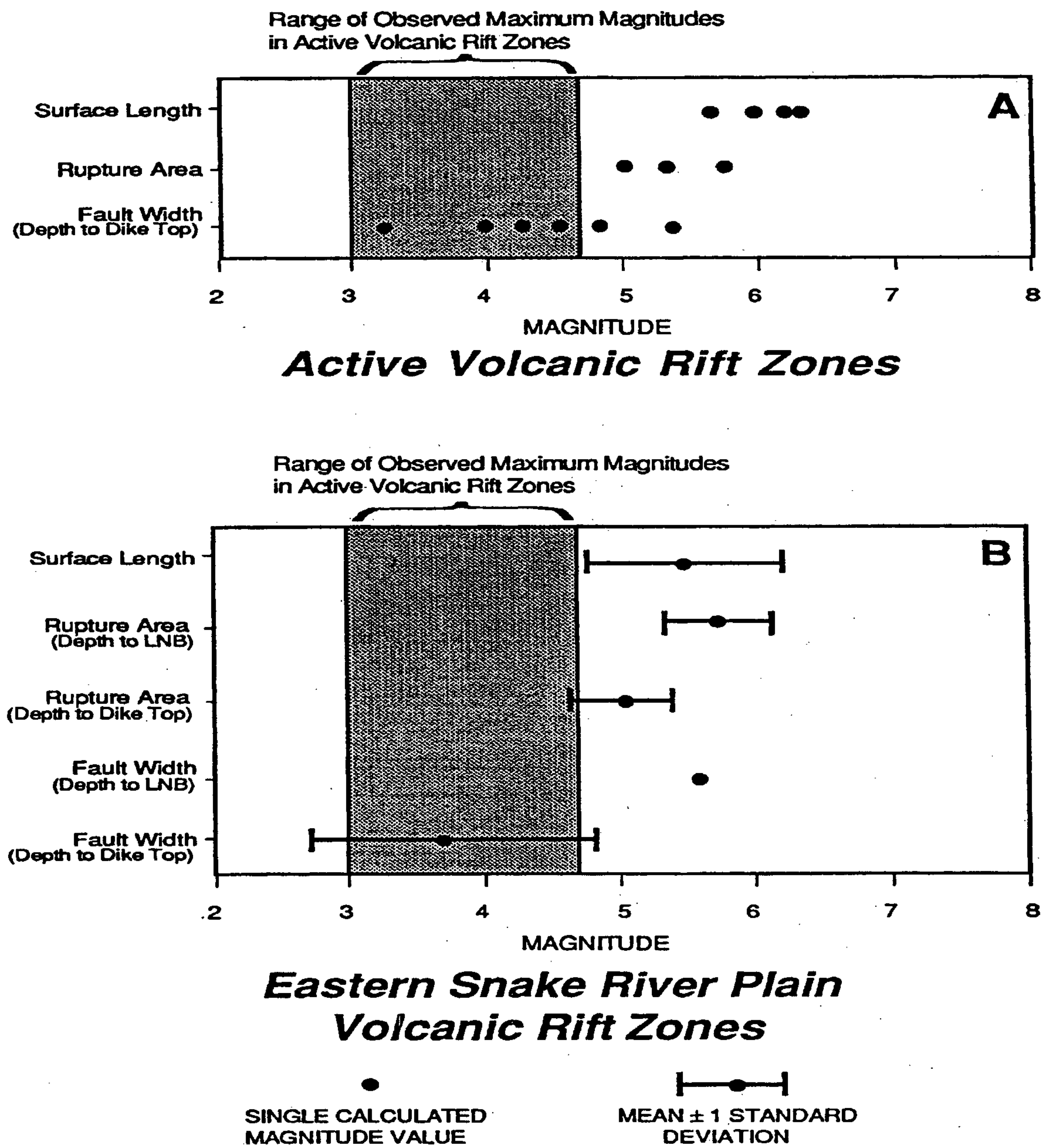

Figure RS.2 Graphical representation of maximum magnitudes calculated using Wells and Coppersmith's (1994) empirical relationships (see text) and fault dimensions of normal faults and fissures in (A) active volcanic rift zones, and (B) Eastern Snake River Plain volcanic rift zones. Depth to the dike top and level of neutral buoyancy (LNB) are used to estimate fault widths. Shaded band represents the range for the mean and one standard deviation of maximum magnitudes (M3.8 ± 0.8) observed for dike intrusion episodes worldwide. Figure courtesy of the Idaho National Engineering and Environmental Laboratory. Permission to use this copyrighted material granted by The Seismological Society of America.

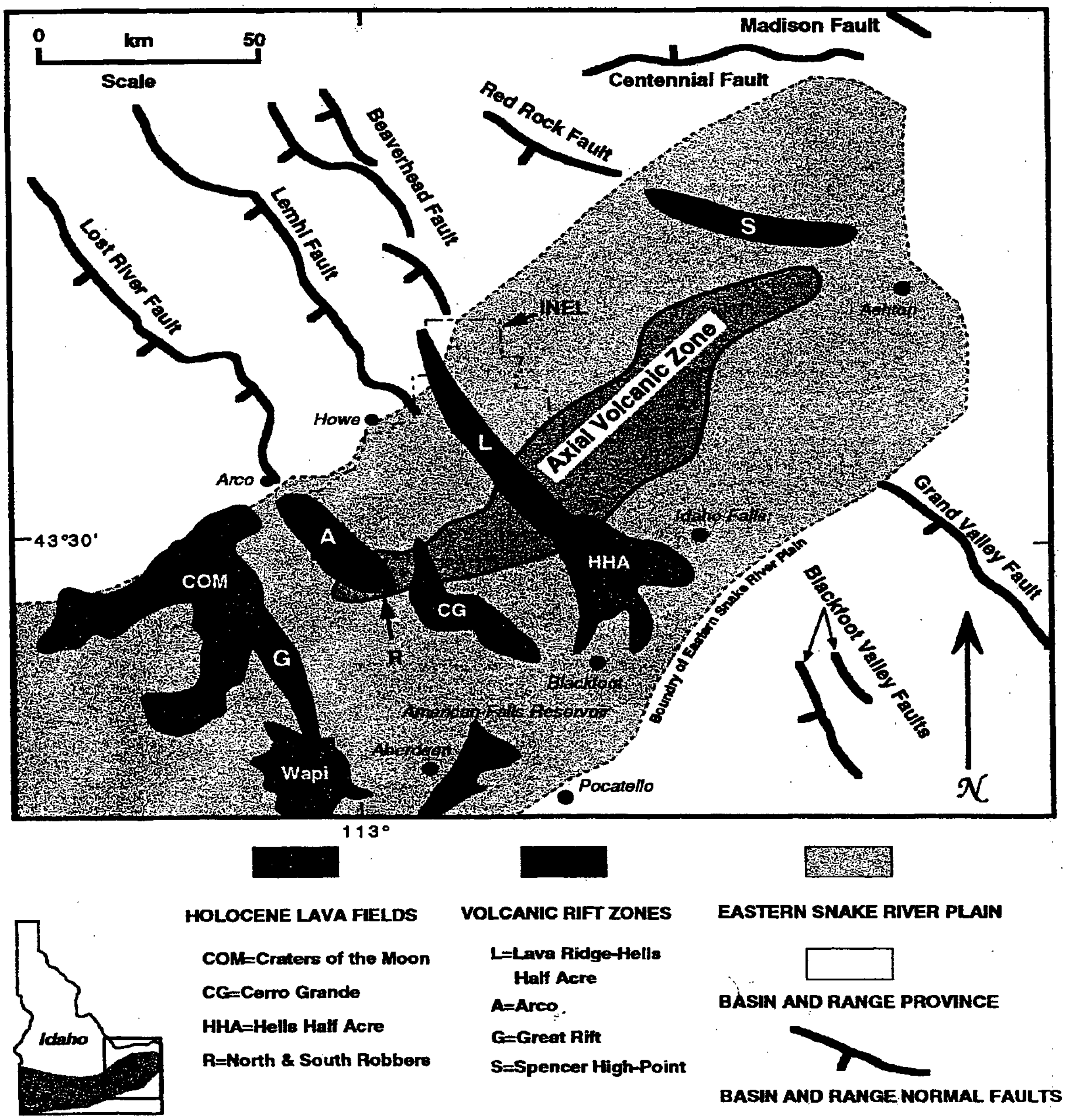

Figure RS.3 Map of the eastern Snake River Plain and surrounding region, showing tectonic and volcanic features and the location of Idaho National Engineering Laboratory (INEL). Figure courtesy of the Idaho National Engineering and Environmental Laboratory.

fault widths that may extend to the depth of the LNB), a **M** 5.5 was chosen for the maximum magnitude that may be associated with ESRP rift zone volcanism. **M** 5.5 is an appropriate upper bound for two reasons. First, the average of all calculated moment magnitudes is 5.2$\pm$0.9 (Table RS.1). And second, for those estimators that produce the best agreement with observed magnitudes during dike injection (i.e., fault width and rupture area based on estimated depth to dike tops) the value of 5.5 covers the mean plus one standard deviation (Figure RS.2B).

5.3 Recurrence

Recurrence intervals for volcanic cycles in volcanic rift zones of the INEL area range from about ~15 ka to 100 ka (Table RS.2). Existing whole rock K/Ar, radiocarbon, and paleomagnetic age determinations of ESRP basalts (Kuntz et al., 1988, 1994) and an independent TL age (Forman et al., 1993) of a relatively young lava flow in the Arco volcanic rift zone (Figure RS.2) are used to determine the time period during which each of the volcanic rift zones was active. This information is augmented by existing whole rock K/Ar and paleomagnetic age determinations from drill holes (Champion et al., 1988).

The recurrence intervals are based on grouping of vents and fissures into geologically reasonable sets according to several criteria, including: (1) the observation that vent-fissure groups resulting from discrete ESRP fissure eruptions attain a maximum areal extent of ~2x5 km; (2) consistency with absolute and relative age assignments; (3) consistency with the ubiquitous north-northwest alignment of vents, fissures, and faults; and (4) textural and compositional similarity of volcanic products in vent groups. Experience developed from mapping spatial and temporal relationships of vents and fissures in the context of ESRP magmatic processes is also useful in assignment of vents to cogenetic clusters. The recurrence interval estimates for the ESRP volcanic rift zones represents a progress report and are likely to be modified as ongoing geochronologic and petrologic investigations are completed.

5.4 Implications For Seismic Hazards Assessment

Credible assessments of seismic hazards must incorporate all potential earthquake sources. In the INEL area, these sources include both large-magnitude earthquakes along major basin and range faults, and near-field, small to moderate earthquakes that may be associated with future magma intrusion and volcanism (Figure RS.3). Volcanic recurrence intervals in the INEL area (Table RS.2) are substantially longer (10^4-10^5 years) and the estimated maximum magnitude (**M**5.5) of volcanic seismicity is substantially lower than that for seismicity on nearby basin-and-range normal faults (10^3 - $2x10^4$ years; **M** 7.0). Estimated ground motions and annual probabilities of occurrence in the INEL area are therefore lower for volcanic rift zone seismicity than for potential nearby tectonic earthquakes and volcanic rift zone seismicity does not contribute significantly to the overall seismic hazard at INEL (Woodward Clyde Consultants, 1992).

6 CONCLUSIONS/SUMMARY

Seismic hazards associated with dike intrusion in extensional volcanic terrains can be assessed using innovative approaches, and by recognizing the genetic relationship of surface structures to shallow dike intrusion and to cogenetic volcanic products. The most successful approach involves several independent methods, some of them atypical of paleoseismic assessment of tectonic faults. Volcanic recurrence, as determined from dating the volcanic products, is used in place of earthquake recurrence determined from conventional paleoseismic investigations of colluvial wedges and fault-scarp morphology in unconsolidated deposits. Procedures for dating individual fissures and scarps in volcanic rift zones are largely untested and include dating of sediments accumulated in fissures or against fault scarps, and dating of the exposure time of deformation-generated geomorphic surfaces. Potential problems involve the complexities of sediment accumulation and fissure/fault development.

The limited down-dip extent of dike-induced faults and their incremental rupture during dike intrusion suggest that the magnitudes of intrusion-related earthquakes will be small to moderate. Maximum magnitudes are best estimated using empirical relationships between fault width or fault area and magnitude. Estimates based on fault length consistently overestimate the maximum magnitudes instrumentally observed during dike intrusion in active volcanic rift zones.

Application of these described approaches to volcanic rift zones on the ESRP is the basis for our assessment of their seismic potential. The geochronology of lavas cogenetic with structures in volcanic rift zones suggests long recurrence intervals between volcanic cycles (15-100 ka). Low to moderate magnitudes of volcanic earthquakes in the ESRP ($M<5.5$) are established by use of empirical relationships involving estimated fault width (down-dip extent), and by analogy with maximum magnitudes of instrumentally recorded seismicity in active volcanic rift zones similar to those on the ESRP.

The concepts presented here can be applied to any extensional area where magma intrusion rates are sufficient to accommodate the deformation. Application of the approaches focuses the effort towards techniques that provide the greatest chance of establishing meaningful recurrence and magnitude data. The assessed level of seismic hazard may be lower than would be indicated by techniques that do not recognize the role of magma intrusion in the faulting and fissuring process.

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Table RS.1 Maximum magnitudes calculated from normal fault and fissure dimensions in eastern Snake River Plain volcanic rift zones

	Moment Magnitudes M^a			
Normal faults or fissure	**Surface length**	**Fault width[b]**	**Rupture area SL X DDT**	**Rupture area SL X LNB[c]**
		Great Rift		
Fissure	6.4	3.6	5.0	5.8
Fissure	6.6	4.1	5.4	6.0
Fissure	6.3	4.1	5.1	5.7
		Arco		
Fault	5.9	2.5	4.1	5.3
Fault	5.7	nc	nc	6.4
Fault	5.6	2.5	4.8	5.6
Fault	4.9	nc	nc	5.6
Fissure	4.9	nc	nc	5.6
Fissure	5.5	nc	nc	6.2
Fissure	5.2	nc	nc	5.9
Fissure	4.5	nc	nc	5.2
Fissure	4.7	nc	nc	5.4
Fissure	5.1	nc	nc	5.8
		Howe-East Butte		
Fissure	4.8	nc	nc	5.5
Fissure	5.3	nc	nc	6.0
		Lava Ridge-Hells Half Acre		
Fissure	6.4	3.8	5.1	5.8
Fissure	5.2	nc	nc	5.9
		Spencer-High Point		
Fault	6.3	3.4	4.8	5.7
Fault	6.2	4.3	5.1	5.6
Fault	6.2	4.1	5.1	5.6
Fault	6.3	4.6	5.3	5.7
Fault	5.7	2.9	4.1	5.2
		LNB[c]		
	na	5.4	na	na
Mean + 1σ	5.6 + 0.7	3.8 + 0.9	4.9 + 0.4	5.7 + 0.3
Range of values	4.5 – 6.6	2.5 – 4.6	4.1 – 5.4	5.2 – 6.4

SL – Surface length; FW – Fault width; DDT – Depth to the dike top; LNB – Level of neutral buoyancy; nc – Not calculated because only one fissure or fault exposed, and therefore, the depth to the dike top could not be estimated (Jackson, 1994); na – Not applicable; a – Maximum magnitudes calculated by Jackson (1994) using empirical relationships of Wells and Coppersmith (1994) for surface length, fault width, and rupture area; b – Fault widths were calculated using graben widths to estimate the depths to the dike top (Jackson, 1994); c – A fault width of 4.0 km (Jackson, 1994) was used for the level of neutral buoyancy (Ryan, 1987) or depth extent of low tensile strength where dikes propagate within the upper crust (Gudmundsson, 1984).

Table RS.2 Estimated volcanic recurrence intervals for volcanic rift zones and borehole sites in the INEL area

Volcanic zone or borehole	Data sources	Time interval of volcanism years B.P.	Number of vents, fissures, or flow groups	Comments	Estimated recurrence interval (years)	Annual probability of occurrence
Great Rift (25 km southwest of INEL)	Kuntz et al. (1986, 1988)	2100-15,000 (radiocarbon dating)	> 100 vents; eight Holocene eruptive periods (each lasting a few decades or centuries and each including multiple flows and cones).	no impact on INEL; most recently and frequently active of all ESRP rift zones; thus provides minimum-recurrence for entire ESRP; most probable area of future ESRP volcanism	2,000	5×10^{-4}
Axial volcanic zone (southern INEL)	Kuntz et al. (1986, 1994)	5000-730,000 (K-Ar dating; radiocarbon; paleomagnetic data)	73 vents and fissure sets; four Holocene lava fields, three of them shared by volcanic rift zones; 45 cogenetic vent/fissure groups	could affect much of southern INEL; most recently and frequently active of all volcanic zones that could impact INEL	16,000	6.2×10^{-5}
Arco volcanic rift zone (southwestern INEL)	Kuntz (1978) and Kuntz et al. (1994)	10,000-600,000 (radiocarbon, K-Ar and TL dating; paleomagnetic data)	83 vents and fissure sets; two Holocene lava fields; 35 cogenetic vent/fissure groups	volcanism could affect southwestern INEL	17,000	5.9×10^{-5}
Lava Ridge-Hells Half Acre volcanic rift zone[a] (north and east INEL)	Kuntz et al. (1986, 1994)	5000-1,200,000 (K-Ar dating; radiocarbon; paleomagnetic data)	48 vents and fissure sets; one Holocene lava field; Hells Half Acre; 30 cogenetic vent/fissure groups	could affect northern and eastern INEL; extremely long eruptive history; includes oldest and youngest basalts in the INEL area	40,000	2.5×10^{-5}
Howe-East butte volcanic rift zone (central INEL)	Kuntz (1978) and Kuntz et al. (1992)	230,000-730,000 (K-Ar dating; paleomagnetic data)	7 vents and fissure sets, no Holocene features, 5 cogenetic vent/fissure groups	old, poorly exposed and sediment-covered; identified in part by subsurface geophysical anomalies	100,000	1×10^{-5}
Borehole NPR site E (south central INEL)	Champion et al. (1988)	230,000-640,000 (K-Ar dating; paleomagnetic data)	9 lava-flow groups (each group contains multiple flows, erupted over a short time)	dates from 600-foot (183-m) interval of subsurface lavas give recurrence estimate consistent with surficial geology of the area	45,000	2.2×10^{-5}
Borehole RWMC 77-1 (southwestern INEL)	Kuntz (1978) and Anderson and Lewis (1989)	100,000-565,000 (K-Ar and TL dating; paleomagnetic data)	11 lava flow groups (each group contains multiple flows, erupted over a short time)	dates from 600-foot (183 m) interval of subsurface lavas give longer recurrence interval than nearby Arco and Axial zones, reflecting flow group (subsurface) versus vent-counting (surface geology) approaches	45,000	2.2×10^{-5}

[a] Includes Circ Butte/Kettle Butte volcanic rift zone

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LARGE-SCALE GRAVITY GLIDING AND SPREADING ABOVE SALT OR SHALE

by

Bruno Vendeville[1]

1 INTRODUCTION

This paper briefly summarizes the current state of knowledge about salt and shale tectonics with special emphasis on (1) the role of large-scale gravity gliding and spreading, the two main processes driving halokinesis and growth faulting, and (2) the seismogenic potential of salt/shale structures to generate moderate to large magnitude earthquakes (Magnitude >5).

Rock salt and overpressured shale are weak mobile rocks that can deform and flow under low tectonic or gravitational stress levels. In addition to their ability to flow and spread, both rocks commonly act as basal lubricating layers above which the overlying rocks can glide, spread, extend, and be translated over large distances (e.g., Bally et al., 1981; Ewing, 1983). Salt/shale tectonics are intimately associated with regional extension, during which gravity forces (large-scale gravity spreading or gliding) play a major role, whether extension is thin-skinned or thick-skinned.

In this report, salt and shale tectonics are analyzed and described jointly because of their similarities in terms of the structural settings in which they take place, of deformation style, and of processes driving deformation. The main difference between salt and shale structures is their longevity and the degree of maturity they can reach: unlike rock salt, whose mechanical properties, hence mobility, is solely due to its mineralogic composition (and therefore does not vary dramatically with time and amount of deformation), shale mobility is essentially due to abnormally high fluid pressure. Shale layers that are not overpressured are not mobile enough to produce diapirs or allow for large-scale gravity deformations. Overpressured shale can lose excess pore pressure when deformed and mobilized into diapirs or rollers and hence loose their extreme mobility. Moreover, the high density contrast between sediments and rock salt favors its diapiric rise, whereas there commonly is little or no density contrast between shale and the overlying brittle sediments. Shale diapirs commonly rise less, do not emerge, extrude and spread at the surface as do salt diapirs.

2 TECTONIC SETTINGS

A survey of the world's salt-diapir provinces by Jackson and Vendeville (1994) showed that salt or shale structures (diapirs or growth faults) are nearly always associated in time and space with extensional tectonics. Areas that have all the characteristics traditionally believed to allow for salt or shale movement (e.g., thick source layer, density inversion between sediment and source layer or differential loading caused by uneven deposition of lobes of clastic sediments) but never underwent extension during their geologic history did not develop salt or shale structures. Examples of basins having thick (>500m) salt but no salt structures in North America include the Williston basin, the Mid-continent basin, the Colorado basin, the Western Canada basin and the Michigan basin (references in Jackson and Vendeville,

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1994). By contrast, nearly all the basins where salt or shale diapirs and growth faults are present underwent extension at some point during their tectonic evolution. Jackson and Vendeville (1994) also showed that the onset of diapirism coincided with the timing of extension and formation of growth faults in these basins.

Diapirs and growth faults can be found in two kinds of extensional settings (Figure BV.1). First, they occur in divergent continental rifts or basins affected by thick-skinned extension, such as the North Sea, the Scotian Basin, Jeanne-d'Arc basin, the Whale-horseshoe Basin, or the Nordkapp Basin (Norway), where salt was deposited before the initial or final episode of rifting began (Figure BV.1a and b). For most of these basins, growth faults formed and salt began to flow and rise up during rifting. In many of these basins, the location and orientation of faults above the salt and that of the diapirs has been attributed to the influence of basement faults underneath. However, many seismic data, experimental results, and theoretical considerations suggest that, unless the salt or shale layer has been dramatically thinned (say, 100 m or thinner), it acts as an insulating cushion between the basement and its cover (Figure BV.2) and does not allow faults in the basement to propagate upward into the suprasalt sediments (Vendeville and others, 1995). This implies that even though the basement, salt, and suprasalt sediments might undergo the same amount of extension, the location of salt structures and cover faults remains mostly independent of that of the basement faults. Moreover, the large number of diapirs and growth faults occurring above undeformed basement in basins where salt and shale structures formed after rifting clearly demonstrates that basement faulting is not required for such structures to form.

Second, diapirs and growth faults occur in basins and on passive continental margins where the salt or shale layer was deposited after rifting ended (Figure BV.1c and d). In this case, extensional tectonics is strictly thin-skinned, affects only the salt/shale layer and overlying rocks, and leaves the base of the salt/shale layer undeformed. Examples include the Gulf of Mexico (from SW Texas to NE Florida; Jackson and Galloway, 1984; Bally et al., 1981; and Nelson, 1991) and the south Atlantic margins (Angola and Gabon in West Africa [Duval et al., 1992]; Campos and Santos basins in Brazil; Niger delta [Weber and Daukoru, 1975]). In all these basins, early thin-skinned extension and salt flow were recorded by stratigraphic thinning and thickening, and onlaps and truncations of seismic reflectors across growth faults and above rising salt pillows. In this type of setting, salt and shale structures are commonly caused by two processes, gravity gliding and gravity spreading, which may act together or independently (Figure BV.3).

Gravity gliding is the transport or translation of rock masses down a basal slope above a weak detachment plane or décollement layer (Figure BV.3a; from Ramberg, 1981). In this report, the term gravity gliding refers to large-scale movement of a sedimentary section moving above a salt or shale décollement layer. Unlike small-scale gravity gliding or sliding (e.g., landslides), that occur in a geologically short period of time and are often triggered by earthquakes, large-scale gravity gliding is a process that occurs continuously for periods of time of a few million years.

Gravity gliding requires that both the base and the topographic surface of the gliding unit be tilted. Gravity gliding typically forms normal faults in the upper part of the slope, and folds or thrusts in the lower part of the slope. The moving blocks generally remain undeformed. Although gravity gliding was commonly used to explain the formation of growth faults, many observations, mainly from seismic data from the Texas Gulf Coast and other hydrocarbon-bearing salt basins, suggest that its applicability is limited. For example, subsidence of growth faulted continental margins caused by repeated deposition of thick wedges of clastic sediments has commonly tilted the base of the salt or shale layer landward (Figure BV.4). Because gravity gliding requires that the base of the detachment layer be tilted seaward, it cannot be regarded as the trigger for growth faults along such margins. Gravity gliding also requires that the basal detachment plane cuts through the topographic surface both updip, as normal growth faults, and downdip, as folds or thrust planes. However, the detaching salt layer in which most listric normal growth faults along the Texas Gulf Coast are rooted at depth much deeper (about 10 km; 33,000 ft) than the adjacent abyssal plain (depths less than 3 km; 10,000 ft) (Worral and Snelson, 1989).

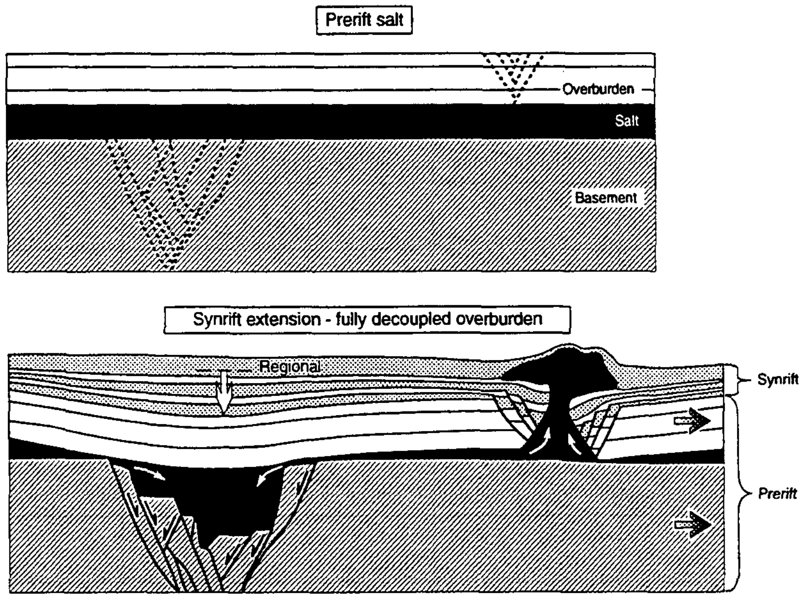
Gravity spreading (Figure BV.3 b) is the vertical thinning and horizontal widening of rock masses driven by the topographic slope and is regarded to be the main trigger for growth faulting and diapirism along passive continental margins. Gravity spreading tends to lower topographic highs and thicken topographic lows by (1) extending the shelf and the upper slope, (2) translating the middle slope seaward, and (3) shortening sediments or salt on and in front of the lower slope (Figure BV.5). Gravity spreading can take place in the absence of a basal slope, but requires (1) a surface slope, and (2) space on the lower slope or in front of the lower slope to allow for lateral shortening. This space can be provided by either the absence of sediments in front of the spreading sediment wedge (Figure BV.5a), formation of folds or thrusts in thin sediments on or in front of the lower slope (Figure BV.5b), or shortening of preexisting diapirs or salt/shale massifs on or in front of the lower slope (Figure BV.5c). In the Gulf of Mexico, gravity spreading is likely to have been the main drive for thin-skinned extension and formation of most of the growth faults and diapirs.

3 PROCESS OF FORMATION AND RESULTING GEOMETRIES OF SALT/SHALE STRUCTURES

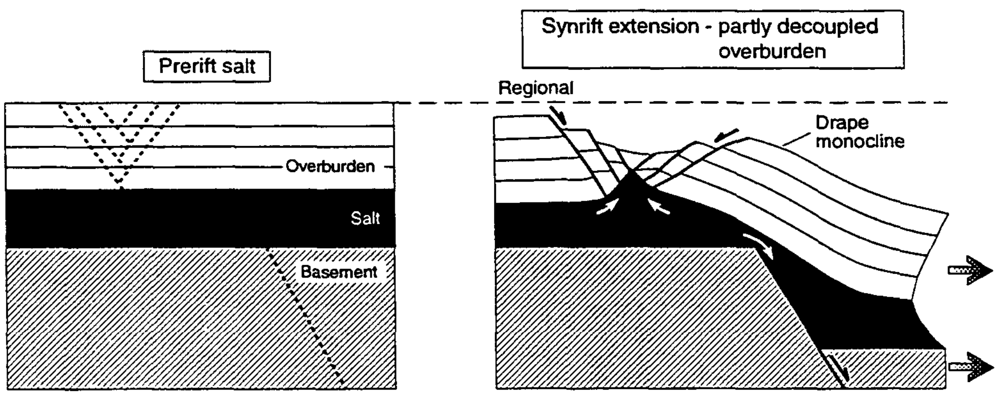
3.1 Diapirs and Associated Faults

Diapirs were traditionally regarded as driven solely by buoyancy. According to this view, the dense sediment overburden pressurizes the viscous salt/shale layer, which would then buoyantly flow upward by deforming its roof. In the past ten years, seismic, numerical, and experimental data have indicated that the finite strength of sedimentary rocks overlying salt or shale layers is too high to allow for such diapiric instabilities to grow spontaneously unless triggered by either regional extension or depositionally-induced differential loading. In tectonic settings subjected to gravity spreading or gliding, normal faulting of the overburden promotes diapir rise regardless of the thickness and density of the sedimentary rocks. Extension-induced diapirs follow three evolutionary stages, reactive, active, and passive stage (Figures BV.6 and BV.7). Each stage is associated with different types of faulting.

During the first stage (reactive diapirism), extension triggers graben faulting of the overburden (Figures BV.6a and d; Figure BV.7a; Vendeville and Jackson, 1992a). As extension proceeds, new graben faults form inward, progressively slicing the graben block into smaller and smaller blocks. Normal faulting creates space for salt underneath to rise. Diapirs at this stage have angular profiles in vertical section, and linear, ridge-like planforms in map view. The growth of reactive diapirs is strictly controlled by the rate of extension. If extension stops, diapir rise



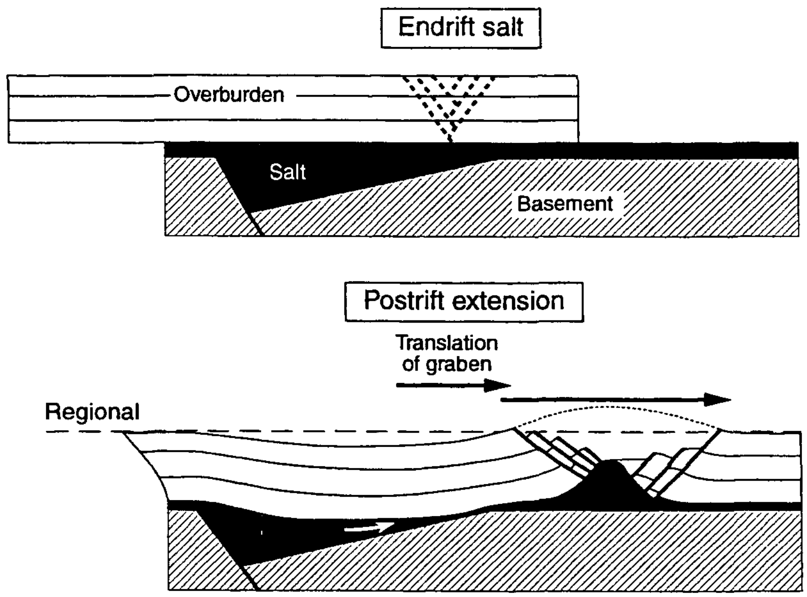
A. Fully decoupled synrift extension in both cover and basement causes independent structural evolution above and below the salt layer. The diapir formed below the overburden faults rather than above the basement faults.



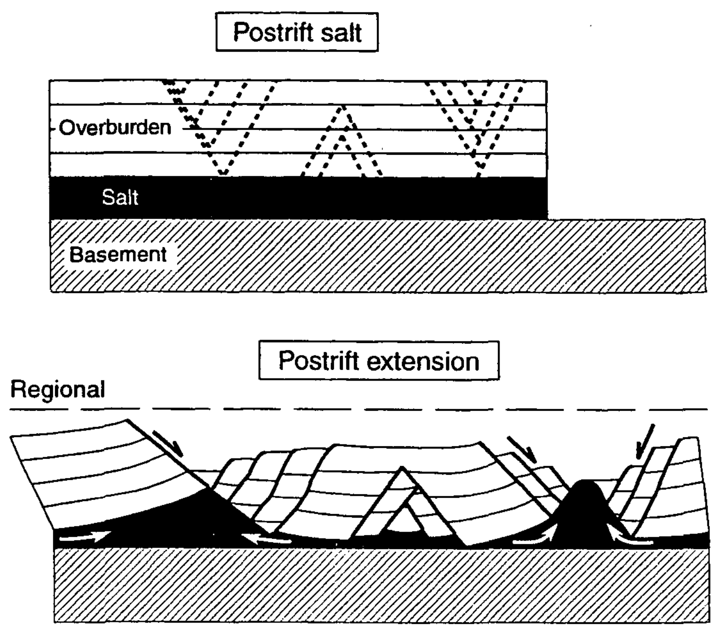
B. Partly decoupled synrift extension in which the cover drapes over the basement fault; stretching of the upper hinge of the monocline localized extension and normal faulting, which initiated a reactive diapir.

Figure BV.1 Four extensional systems that trigger diapirism. In all four systems, stretching of the overburden directly initiates diapirism; any basement faults only indirectly affect diapirism by locally stretching the overburden. All situations have been models in experiments and observed in natural examples. From the Geological Society of America Bulletin, M.P.A. Jackson, and B.C. Vendeville. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1994 Geological Society of America.

(Figure BV.1, continued)



C. Gravity glide of endrift salt and overburden above a basement half graben. Rightward gliding of the overburden and salt flow forms an anticline in the overburden. Local stretching of the crest of the anticline localized normal faulting and subsequent diapir rise.



D. Entirely thin-skinned extension above a flat basement. Diapirs form only where normal faulting has locally thinned the overburden. From Jackson and Vendeville (1994).

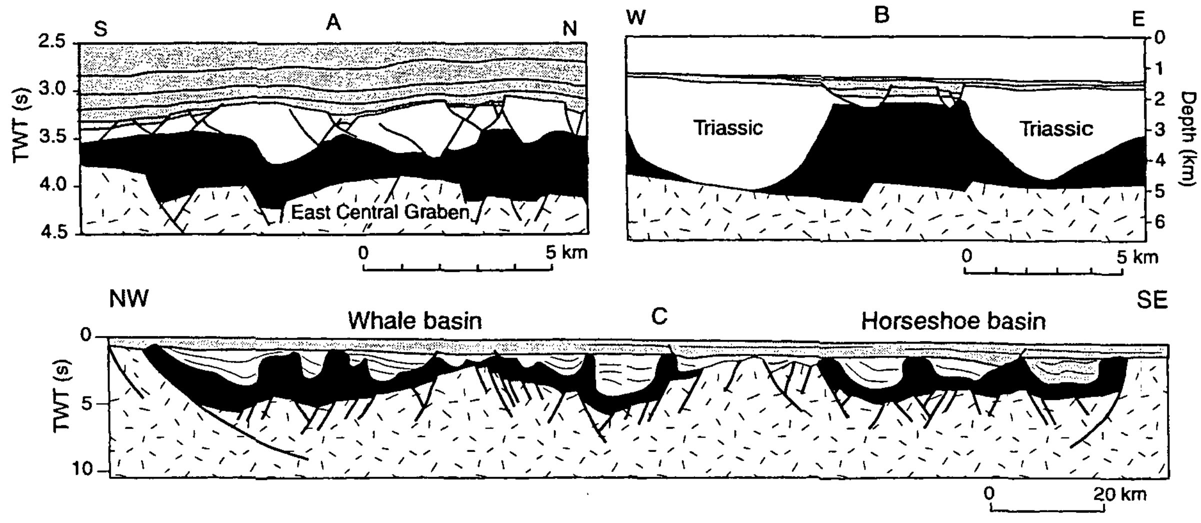


Figure BV.2 Line drawings of three seismic examples where overburden faults and diapirs formed independently of the location of basement faults. (a) East Central Graben (Penge et al., 1993 with permission, from the author and from the Annual Review of Earth and Planetary Sciences, Volume 22 © 1994, by Annual Reviews.); (b) Ula Field (Stewart et al., 1992). Permission to use this copyrighted material is granted by the Norwegian Petroleum Society; (c) Whale and Horseshoe Basins (H.R. Balkwill, and F.D. Legall, AAPG Memoir Series No. 46, AAPG © 1989, reprinted by permission of the American Association of Petroleum Geologists).

stops. As extension thins the graben floor, the diapir rises and syntectonic sediment deposition fills the graben and thickens the overburden, retarding diapirism (Figure BV.8a). Such faults are commonly asymmetric and dip downslope or seaward (Figure BV.9 left and bottom). Rapid extension and slow sedimentation therefore promote diapirism (Figure BV.8c); whereas slow extension and rapid sedimentation lead to fault-bounded depocenters that widen and deepen with ongoing sedimentation and extension. Faults associated with reactive diapirism have linear traces oriented perpendicular to the regional or local direction of extension. In vertical sections, most faults dip toward the diapir; older faults die out upward and are replaced by younger faults closer to the diapir (Figure BV.9).

The second stage, active diapirism (Figures BV.6e; 7b), occurs after extension and reactive diapirism have thinned the graben floor (i.e., the diapir roof) so much that its strength can be overcome by the pressure in the underlying salt. The reactive diapir can thus actively lift and pierce its thinned roof and rise, and emerge. Both natural examples and experimental models indicate that only parts of the early reactive salt ridge reach active piercement. Active diapirs have circular to subcircular planforms and root at depth into the older, linear reactive salt ridge. Faults associated with active diapirism are usually arranged in a radial pattern centered around the diapir (Figures BV.7b and 11 ; Nelson, 1991). In addition, arching and stretching of the diapir roof during active diapirism form keystone grabens above the diapir crest.

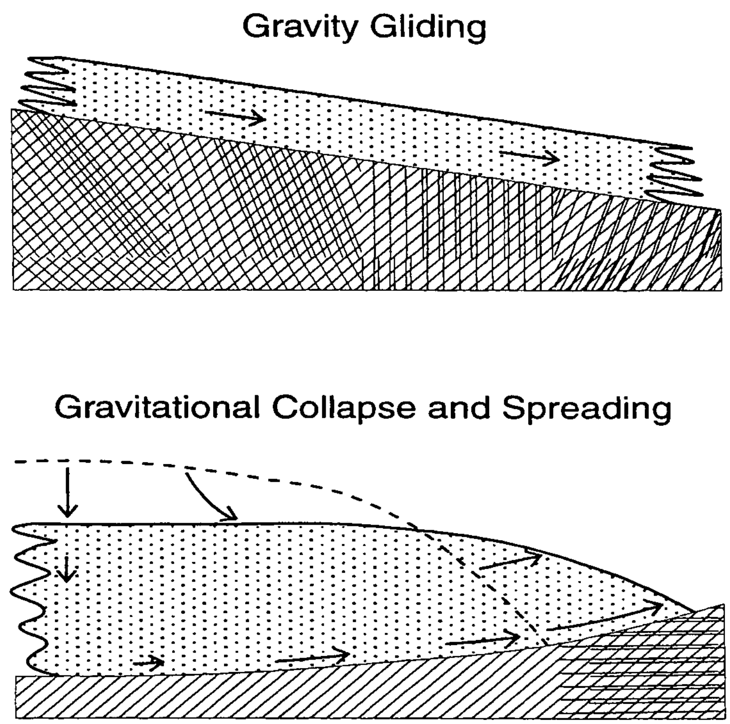


Figure BV.3 Two gravitational transport mechanisms. Reprinted from Deformation of the Earth's Crust in Theory, Experiments, and Geological Application (2^{nd} Ed.), H. Ramberg, p. 20, © 1981, by permission of the Academic Press.

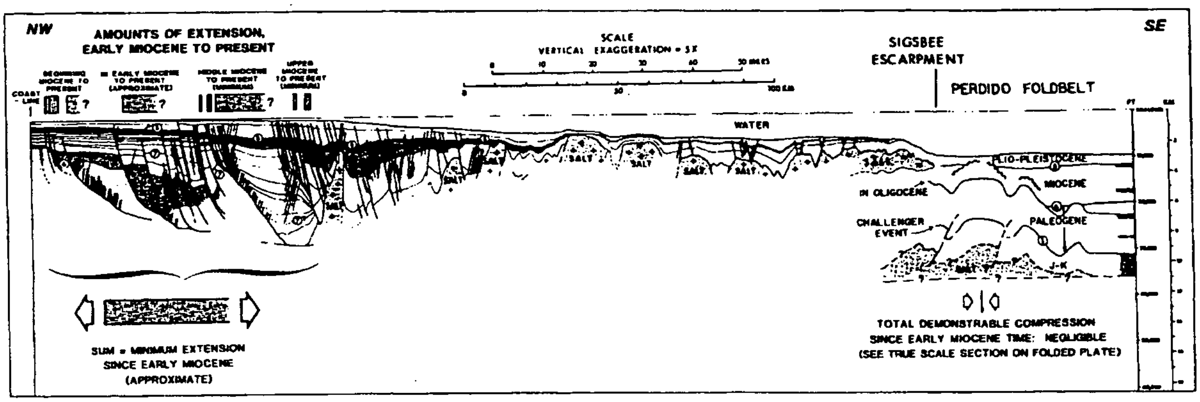


Figure BV.4 Structural cross section of the northern Gulf of Mexico, from the Texas coastline to the Sigsbee abyssal plain showing Miocene to Recent listric normal growth faults on the Texas shelf. Vertical exaggeration = 5X. From the Geological Society of America Bulletin, D.M. Worral, and S. Snelson. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1989 Geological Society of America.

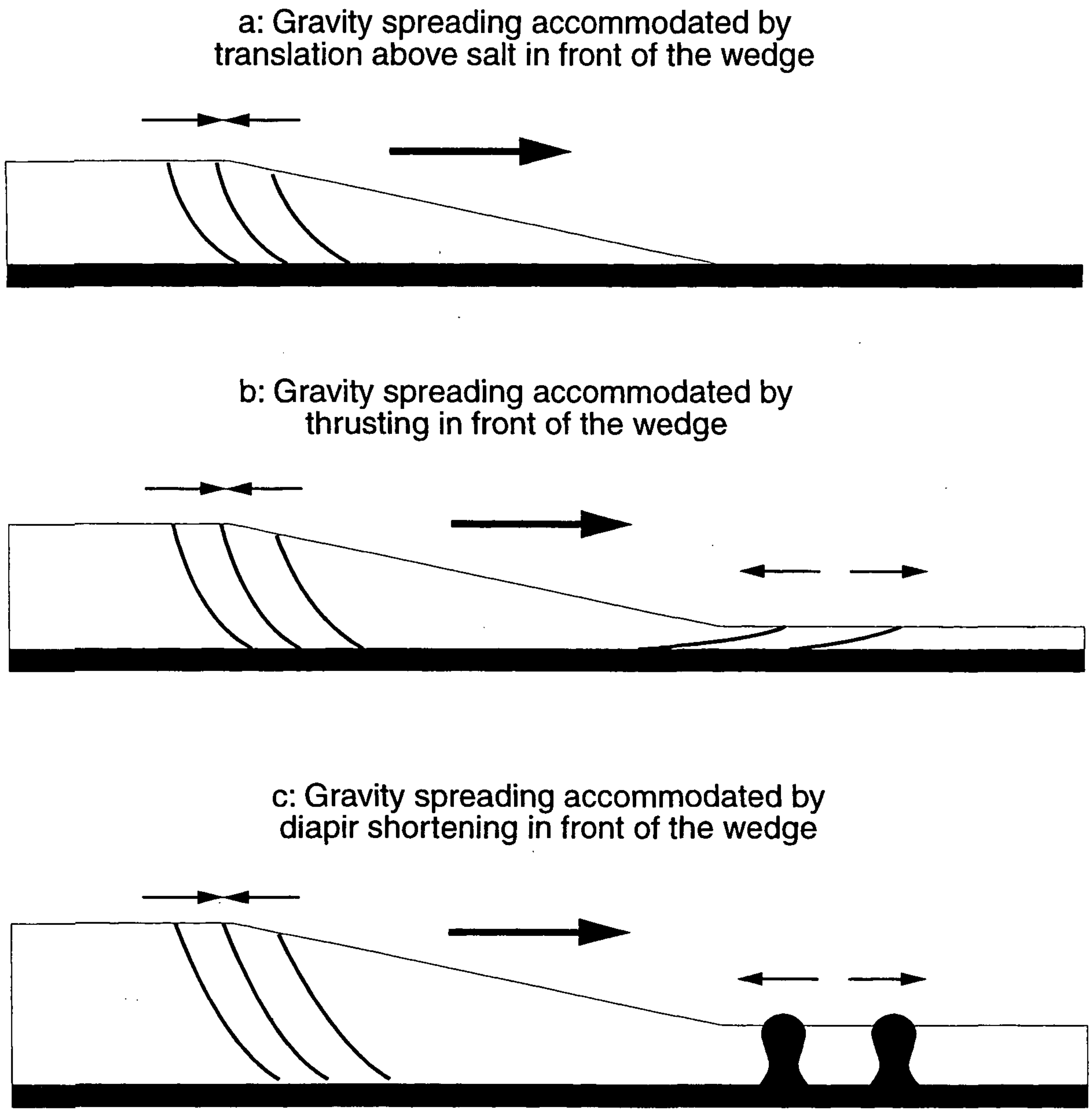


Figure BV.5 Three main types of progradational settings in which gravity spreading of the sediment wedge triggers extension in the shelf and the upper slope.

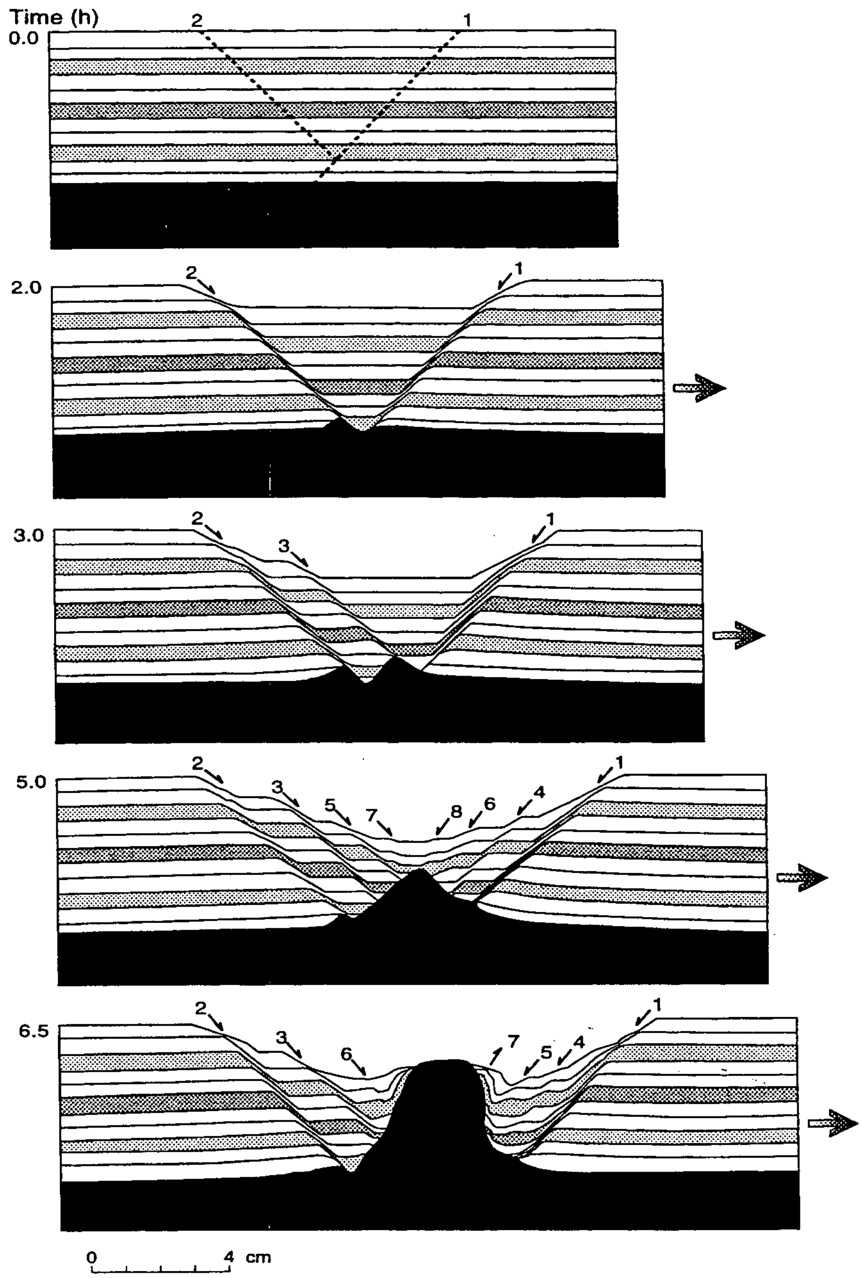


Figure BV.6 Tracings of vertical sections through models of reactive diapirism during thin-skinned extension. (a) initial stage; (b, c, and d) reactive stage; (e) active stage. Black: viscous silicone polymer simulating salt; white and gray: sand layers simulating brittle sediments. The diapir rose reactively until it became tall enough and its roof was thin enough to be pierced actively. From the Geological Society of America Bulletin, M.P.A. Jackson, and B.C. Vendeville. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1994 Geological Society of America.

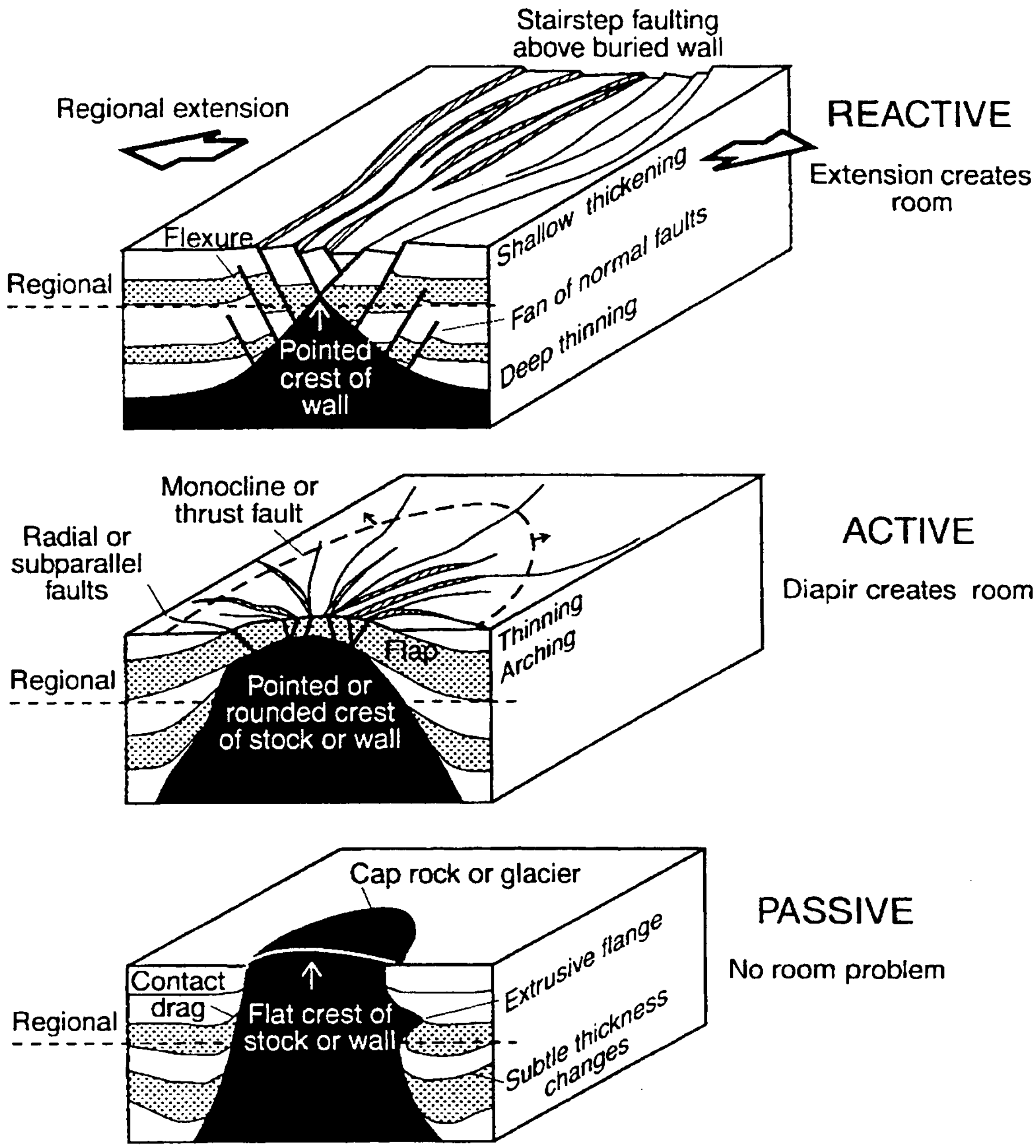


Figure BV.7 Three piercement modes for salt diapirs (black) and their characteristics structures. From Jackson et al. (1994), with permission, from the Annual Review of Earth and Planetary Sciences, Volume 22 © 1994, by Annual Reviews.

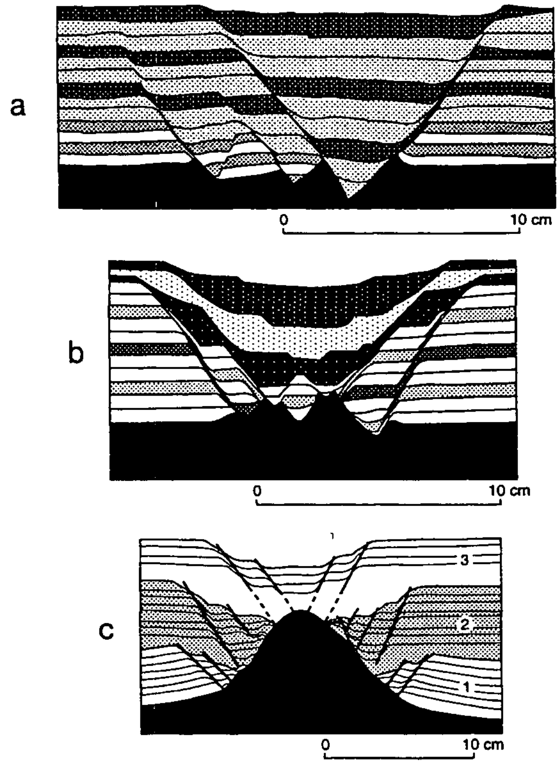


Figure BV.8 Experimental models illustrating the influence of the ratio between the rates of sedimentation and regional extension. Black: viscous silicone polymer simulating salt; white and gray: sand layers simulating brittle sediments. (a) Model with rapid deposition and slow extension deformed by growth faulting. The base of the graben has subsided below the regional datum. (b) Model with moderately rapid sedimentation deformed by growth faulting but the base of the graben did not subside below the regional datum. Stratigraphic thickening caused by sedimentation and tectonic thinning caused by normal faulting canceled out each other. (c) Model with slow deposition and rapid extension. Three overburden layers were deposited episodically. Each had once a continuous top across the entire model at the tie of deposition. A reactive diapir rose below the graben because normal faulting thinned the overburden faster than sedimentation filled it. From Vendeville and Jackson, Rates of Extension and Deposition Determine Whether Growth Faults or Salt Diapirs Form, in Rates of Geologic Processes, GCSSEPM Foundation 14th Annual Research Conference, (1993). Reprinted with permission of the Gulf Coast Section Society of Economic Paleontologists and Mineralogists Foundation.

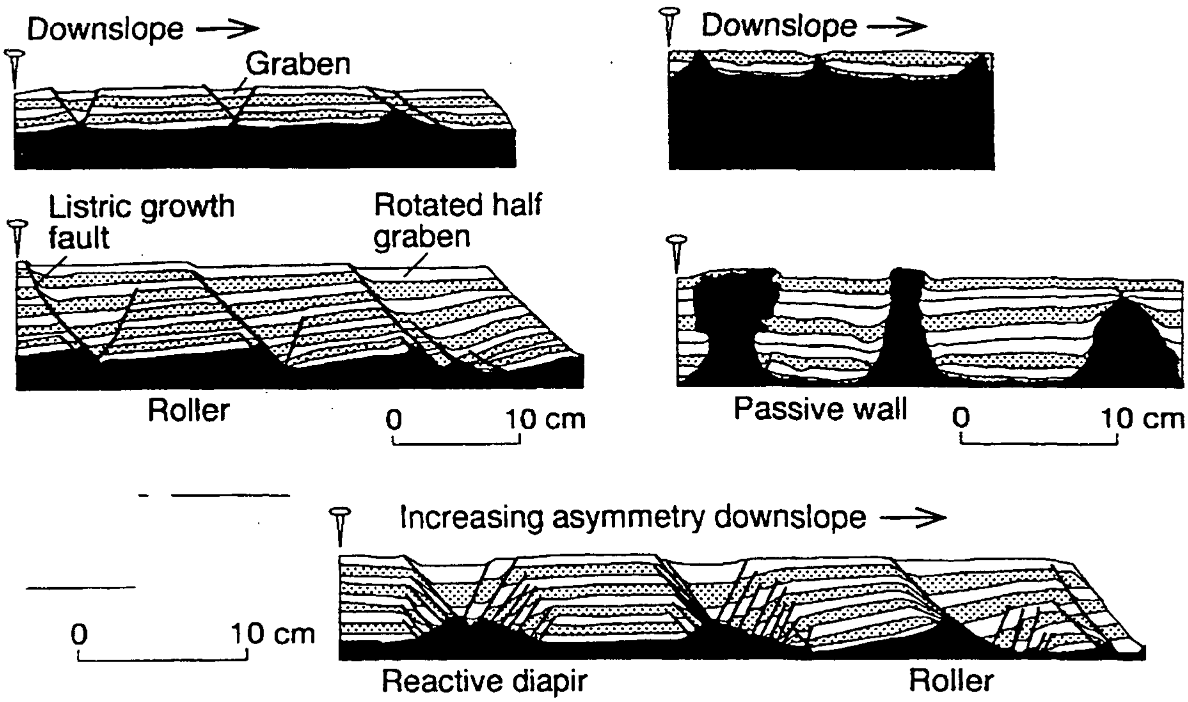


Figure BV.9 Influence of the ratio between sedimentation rate and extension rate in models having multiple faults and salt structures. (Upper left) Rapid sedimentation and slow extension results in buried salt rollers and large listric normal growth faults. (Upper right) Slow sedimentation and rapid extension results in shallowly buried diapirs and negligible faulting. (Bottom right) Medium ratio between sedimentation and extension results in buried reactive diapirs overlain by multiple growth faults. The structural asymmetry increases downslope. From Vendeville and Jackson (1992) and Jackson et al. (1994), with permission, from the Annual Review of Earth and Planetary Sciences, Volume 22 © 1994, by Annual Reviews.

Once a diapir has actively pierced its roof, it emerges at the surface and grows passively (Figure BV.7c). During passive diapirism, also known as downbuilding, the diapir maintains its crest at the seafloor while strata accumulate on its sides. Because a passive diapir has no roof, it can rise without displacing strata, hence does not induce significant folding and faulting of the adjacent sediments, except for localized drag zones along the salt-sediment contact.

faulting (Figures BV.6a and b, 9 top left and bottom right, 12). The best examples of large growth faults detaching on a thin décollement layer are along the Texas portion of the US Gulf Coast (Figures BV.4 and 13) and have been described in detail in Worral and Snelson (1989) and Nelson (1991). Growth faults cut steeply and deeply into the upper continental slope and have traces that extend regionally for tens to hundreds of miles subparallel to the present shoreline. Because growth faults initiate at or near the shelf break, the youngest faults form seaward of the older faults, as the prograding sedimentary wedges advance (Figure BV.14). In vertical section, these faults have a listric profile, dip seaward, and their dip decreases with depth down to a décollement layer of salt or shale 7-9 km deep (Figure BV.13). Commonly, Texas growth faults have had

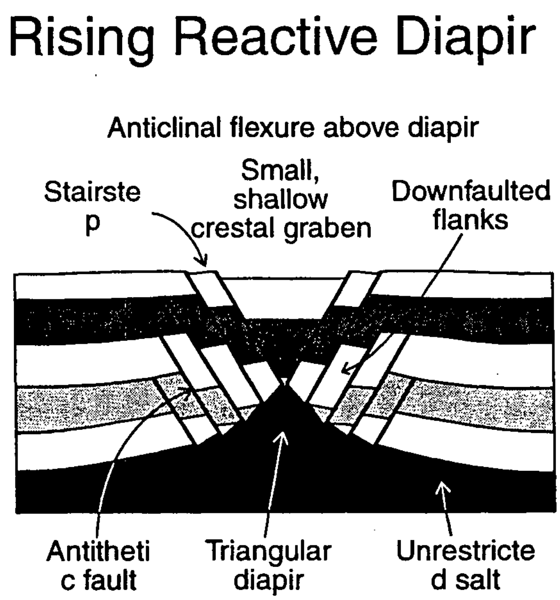


Figure BV.10 Schematic summary of the characteristics of reactive rising diapirs. Reprinted from Marine and Petroleum Geology, v. 9, Vendeville and Jackson, The Fall of Diapirs During Thin-Skinned Extension, p. 370, © 1992, with permission from Elsevier Science.

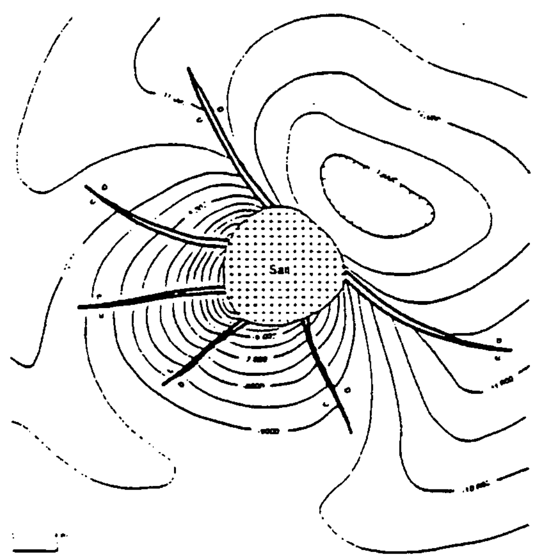


Figure BV.11 Schematic structure map of sedimentary horizons around a passive diapir. From the Geological Society of America Bulletin, T.H. Nelson. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1991 Geological Society of America.

considerable amount of slip (1000's m); strata deposited during growth faulting thickens greatly across the fault plane; strata in the hangingwall deform by rollover folding of the strata in their hangingwall. Where the décollement layer is thicker (Figure BV.15), as along the Louisiana portion of the US Gulf Coast, growth faulting and vertical rise of salt domes combine to produce more complex structures. Fault traces are commonly arcuate (concave seaward), a few km long, and commonly terminate against diapirs. Fault orientation and vergence can also vary greatly. The variation in fault orientation is especially pronounced where the prograding clastic sediments loading the salt were not deposited as a linear, continuous front along the entire shelf edge, but more locally as circular or hemi-circular delta lobes. Delta lobes tend to spread radially, forming a complex set of intersecting radial and concentric faults.

4 ASSOCIATION WITH SEISMICITY

Published literature on the seismicity associated with the growth of salt structures (as opposed to seismicity associated with salt mining), is scarce. Published data on seismicity associated with shale structures is even rarer. Nearly all published articles conclude that salt structures are not associated with moderate or high levels of seismicity. Frolich (1982) analyzed data from a 1978 earthquake in the Central Gulf of Mexico (magnitude 5.0) and located the hypocenter at a depth of about 15 km, far deeper than the segment of the crust currently involved in salt tectonics and gravity gliding (i.e., 7-8 km or less). Although the central Gulf of Mexico comprises some of the most vigorously growing geologic structures on earth (e.g., more than 15,000 ft of vertical rise/subsidence during Plio-Pleistocene times for some diapirs and the adjacent sediment depocenters; see Figure BV.16), Frohlich concluded that the Gulf of Mexico is virtually aseismic. That a basin containing some of the most active salt structures in the world shows little or no significant seismicity clearly demonstrates that salt tectonics alone does not generate stresses capable of triggering large earthquakes.

At a more local scale, A Safety Evaluation Report related to the South Texas Nuclear Plant project (U.S. Nuclear Regulatory Commission, 1986) evaluated the seismic risk associated with growth faults onshore Texas (Matagorda County) and concluded that the faulted rocks involved in growth faulting were not capable to store enough strain energy to produce moderate or large earthquakes and hence significant ground motion.

Another set of earthquake studies related to salt tectonics are those about the 1984 Carbondale earthquake in Colorado (Goter and Presgrave, 1986; Goter and others, 1988). Results of these studies suggest that the earthquake-generating deformation may have been associated with slip along the interface of or within an evaporitic layer (the Eagle Valley evaporite, at depths of about 3-6 km). However, the magnitude of the seismic events within the earthquake swarm were low (1.9 to 3.2).

A third set of published studies have focused on seismicity associated with salt structures in compressional tectonic settings, such as the Salt Ranges and the Potwar Plateau in Pakistan, the Zagros in Iran, or the Kulyab area in Tadjikistan. Results and conclusion from these studies vary. On one hand, Yeats and Lillie (1991) and Seeber and others (1981) suggested that movement along the evaporitic detachment was unlikely to generate significant seismicity and attributed earthquakes observed in the region to folding were the evaporitic detachment might be absent. On the other hand, a study of the seismicity of the Kulyab region, Tadjikistan by Leith and Simpson (1986) reported moderate 1972-1973 earthquakes (M>5) clustering around emergent and buried diapirs.

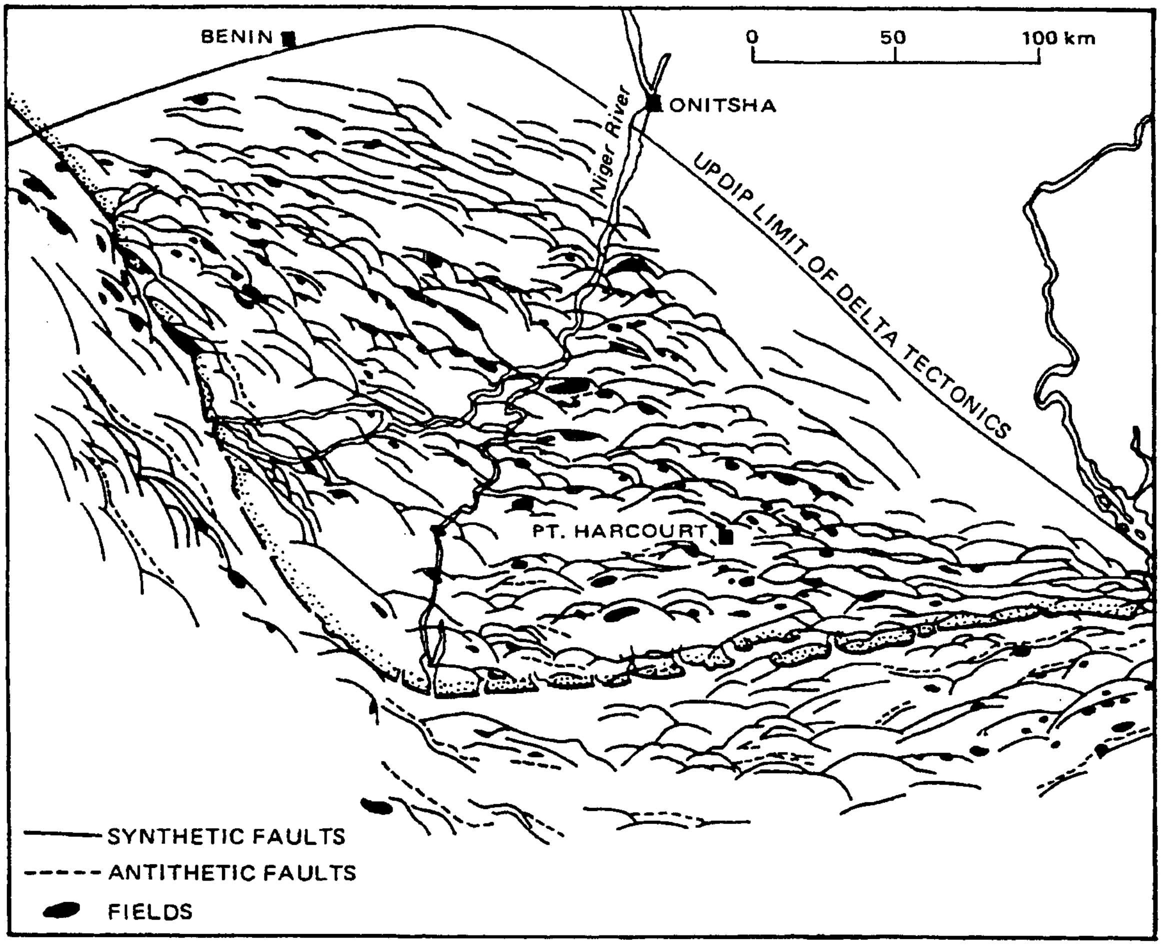


Figure BV.12 Regional map of the Niger delta growth faults. Growth faults formed in response to gravity spreading-gliding above shale. Reprinted with permission from the World Petroleum Congresses, K.J. Weber, and E. Daukoru, Petroleum Geology of the Niger Delta, © 1975.

The Kulyab region is located North of the Pamir range and is undergoing regional shortening. Unlike Leith and Simpson (1986), who attribute the earthquakes to active rise of the salt diapirs, I hypothesize that the observed seismicity reflects folding and/or thrust-faulting of the sediments above and between the diapirs in response to lateral shortening. Because rock salt is much weaker than sedimentary rocks, diapirs in the cover can readily and rapidly deform by viscous flow. By contrast, the adjacent non-evaporitic sediments must deform by faulting or folding and can potentially store much more strain energy, hence induce seismicity. Deformation and seismicity of such geologic settings therefore needs to be regarded as similar to that occurring in fold-and-thrust belts detaching above a salt or overpressured shale décollement layer.

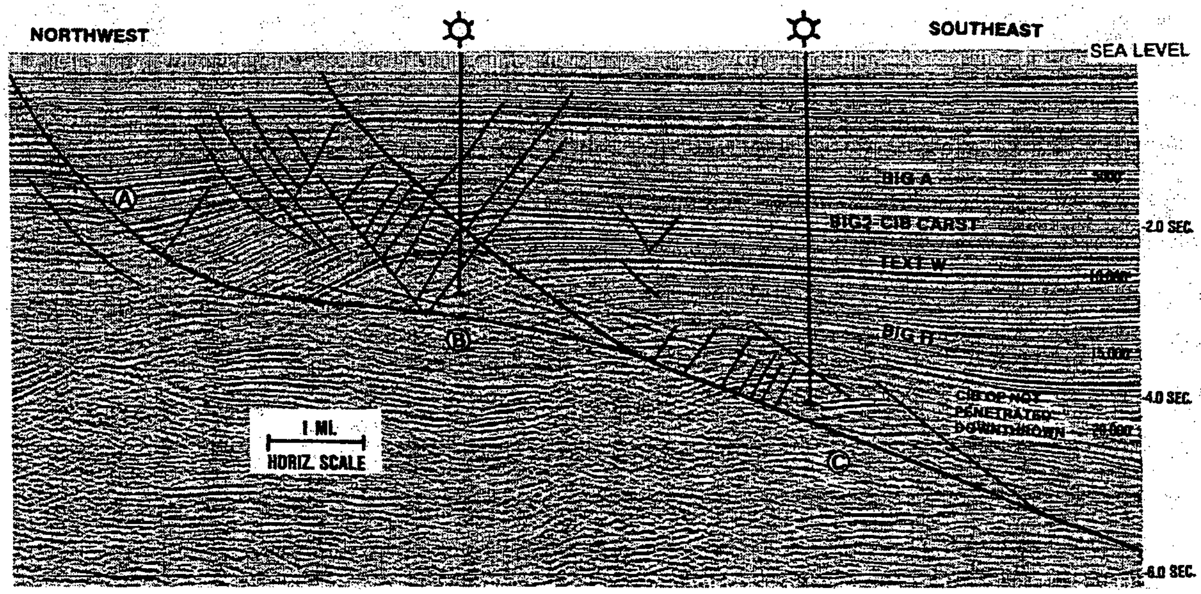


Figure BV.13 Seismic section across the Corsair fault trend, Offshore Texas. (H.A. Vogler, and B.A. Robinson, AAPG Bulletin, v. 71-7, AAPG © 1987, reprinted by permission of the American Association of Petroleum Geologists).

5 CRITERIA FOR IDENTIFYING POTENTIALLY SEISMOGENIC SALT OR SHALE-RELATED STRUCTURES.

As indicated above, salt or shale typically are associated with low seismicity. There are two exceptions.

First, where salt movement occurs in rifts currently undergoing basement-involved extension, slip along the basement faults in the basement beneath the salt may produce significant seismicity. Because salt can decouple the basement from its cover, active faults above the salt are not necessarily located above active basement faults (Vendeville and others, 1995). Therefore, identification of faults in rocks above the salt layer may or may not assist in identifying and characterizing seismogenic tectonic faults below the salt.

Second, diapir-bearing basins subjected to regional compression display significant levels of seismicity (see the above example in the Kulyab region). In such settings, shortening may be accommodated by laterally squeezing the diapirs (Figure BV.17). Similar feature have been described in the Nordkapp Basin, Barents Sea by Nilsen et al. (1995) (Figures BV.18 and 19). Themain clue indicating whether diapirs in such basins are being subjected to horizontal shortening is the presence of a thick, deformed diapir roof having evenly thick strata above the diapir (Figure BV.17). A thick roof indicates that diapiric rise ceased or at least slowed down considerably, a tell-tale sign of source-layer depression, and that the diapir was later rejuvenated by shortening (see Vendeville and Nilsen, 1995). Otherwise, diapirs buried under thick roofs would remain extinct in

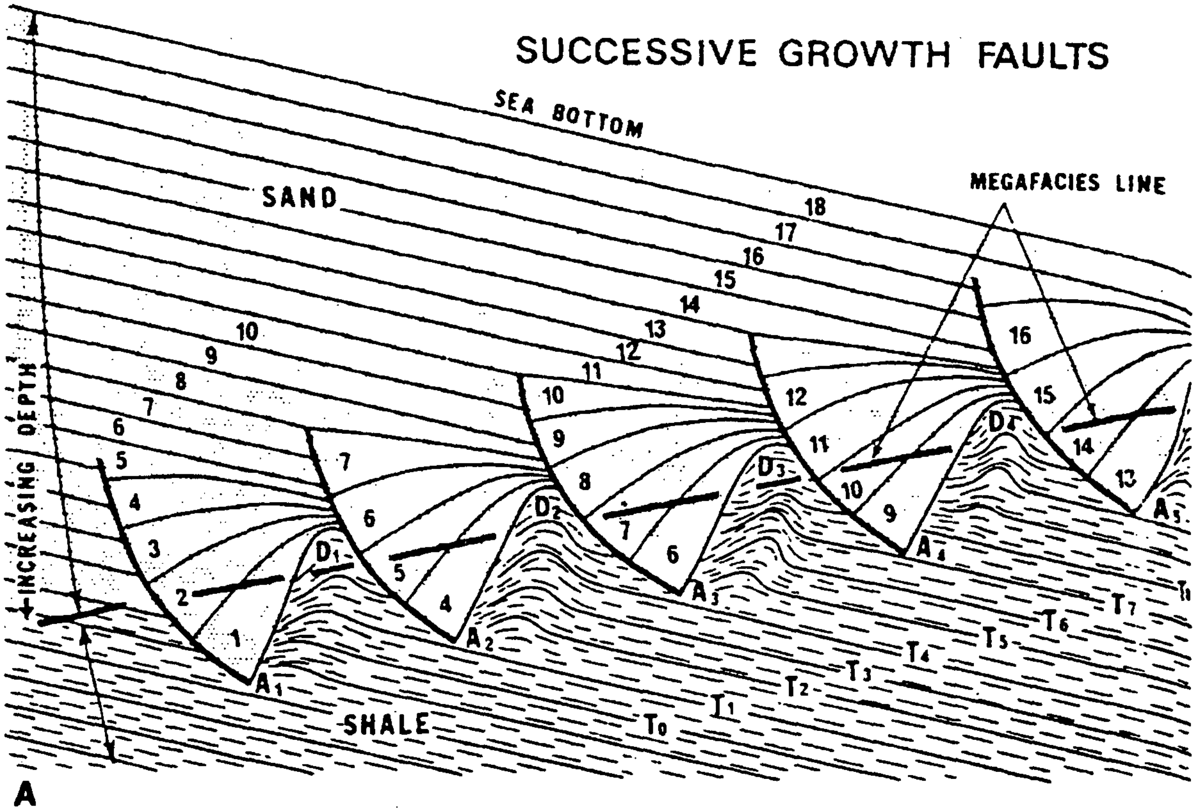


Figure BV.14 Conceptual model for the sequential formation of growth faults during seaward sediment progradation. (H.O. Woodbury, I.B. Murray, Jr., P.J. Pickford, and W.H. Akers, AAPG Bulletin v. 57-12, AAPG © 1973, reprinted by permission of the American Association of Petroleum Geologists).

the absence of regional tectonics, or subside and fall if subjected to extension. Other clues include the following: diapirs having a wide remnant of at their base, forming a salt pedestal that narrows upward where contraction pinched off the upper diapir stem (Figures BV.19 and 20); evidence for diapir rise through flat-lying strata in the adjacent depocenter, indicating that diapir growth occurred after the depocenters had grounded and therefore could be driven by lateral contraction only; compressional or wrench structures affecting the overburden updip, downdip, or alongstrike of the diapirs where contraction was recorded by reverse faulting, folding, or wrenching of the brittle sediments between the diapirs.

6 POSSIBLE CRITERIA FOR DISTINGUISHING FAULTS RELATED TO LARGE-SCALE GRAVITY GLIDING/SPREADING AND BASEMENT-INVOLVED, TECTONIC FAULTS

Faults caused solely by large-scale gravity deformation differ from basement-involved, tectonic faults because they involve only the few uppermost km (< 7 km) of the crust. Because fault spacing partly depends on the thickness of the faulted interval (here, the brittle overburden overlying the salt or shale layer), the spacing of normal faults associated with salt or shale should be less than that of tectonic faults (Figure BV.21, top). However, this clues should be used with caution. The large amount of extension along long-lived, mature growth faults such as faults along the Texas coast or the West African margin (Figure BV.21, bottom), can increase the initial fault spacing by as much as ten km.

Salt/shale related fault blocks also tend to experience block rotation far greater than do tectonic faults, easily accommodated at depth by flow of the source layer. Growth faults are generally listric, their dip decreasing with depth. Moreover, because such faults affect only the suprasalt sediments, block deformation due to faulting, such as rollover folding of the hangingwall or uplift of the footwall are also accommodated at depth by flow of the source layer, rather than by deformation of the basement. This results in more localized rollover folds and footwall uplifts than those associated with basement-involved normal faults.

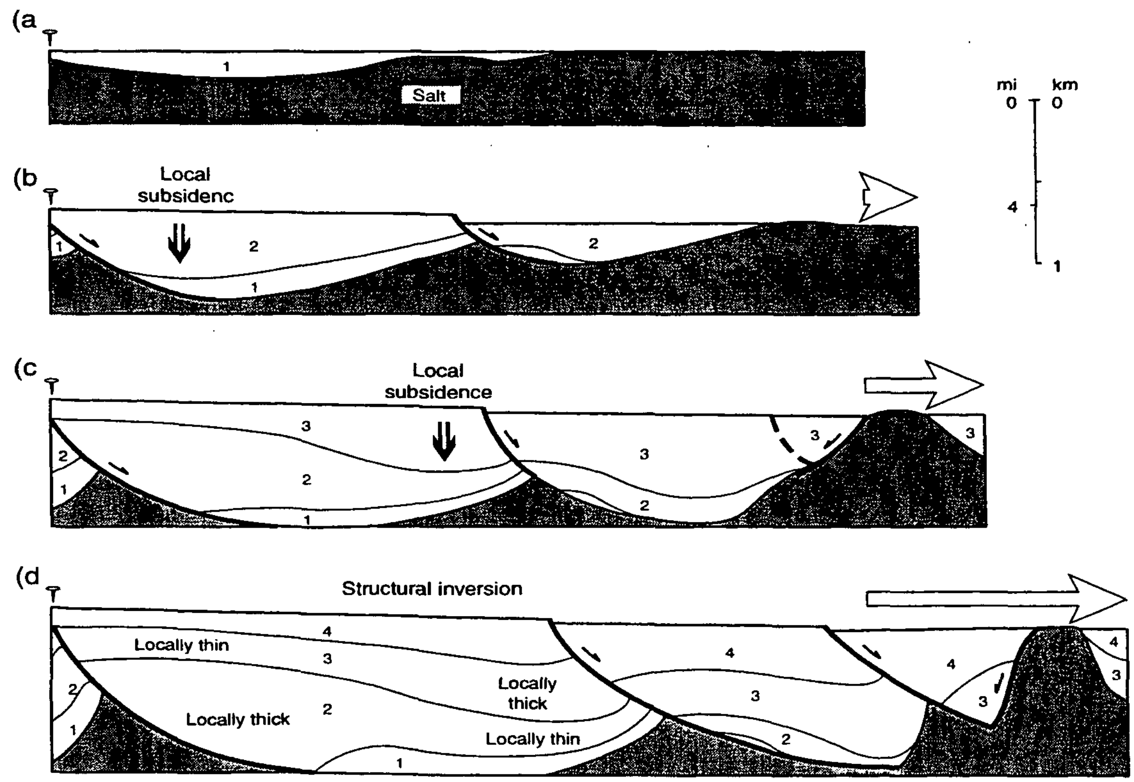


Figure BV.15 Formation of large growth faults above thick salt as a result of thin-skinned extension. Reprinted from Jackson and Talbot (1991), A Glossary of Salt Tectonics, Geological Circular 91-4, with permission from the Bureau of Economic Geology as modified from the Geological Society of America Bulletin, D.M. Worral, and S. Snelson. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1989 Geological Society of America.

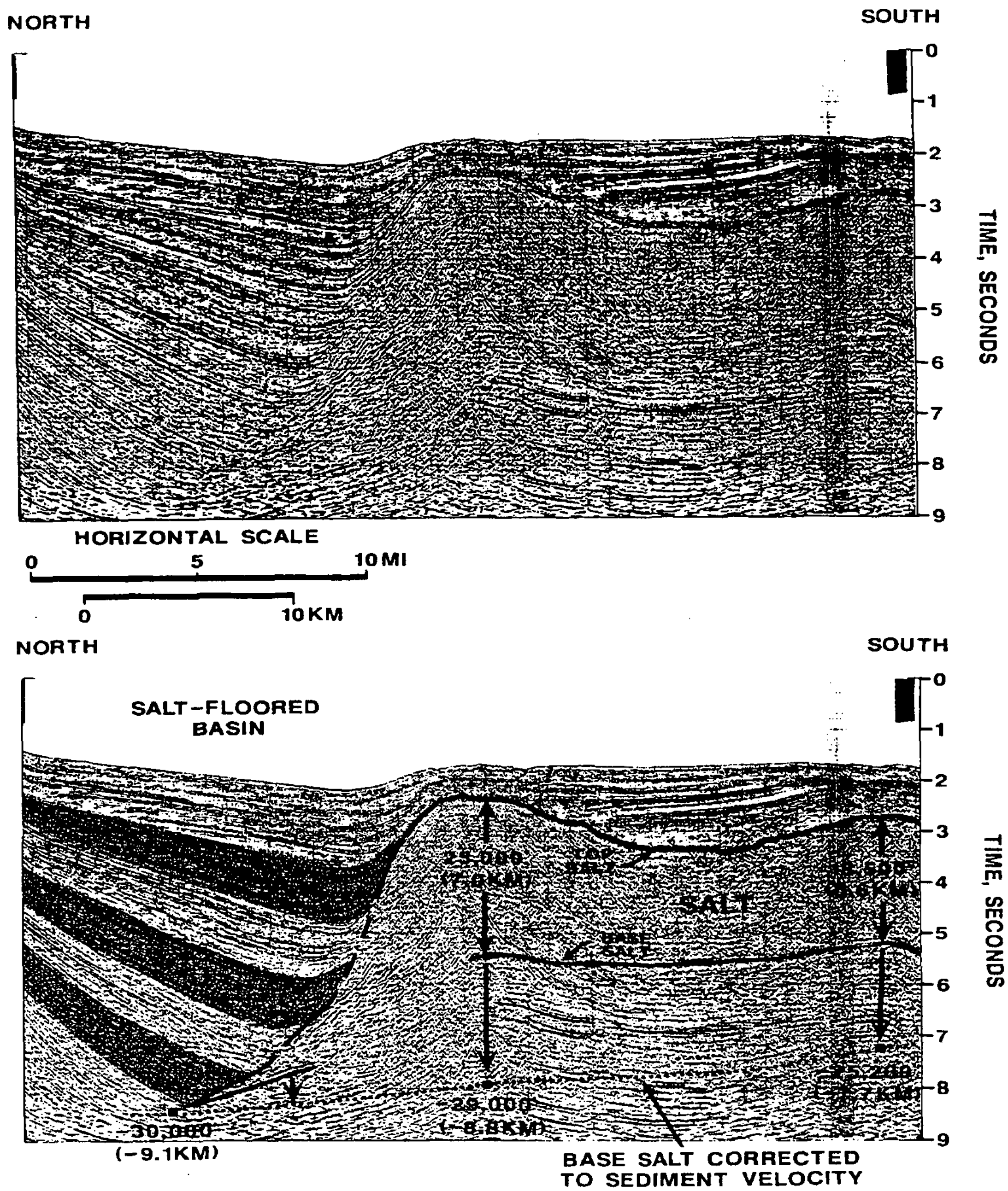


Figure BV.16 Thick, allochthonous salt mass and adjacent Plio-Pleistocene depocenter, Louisiana slope, northern Gulf of Mexico. The salt and the adjacent depocenter are up to 7 km thick and exhibit a velocity pull-up of nearly 2.5 seconds. The base salt has been corrected (dotted line) to the same time horizon to better visualize the relative geometry. From the Geological Society of America Bulletin, D.M. Worral, and S. Snelson. Reproduced with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright © 1989 Geological Society of America.

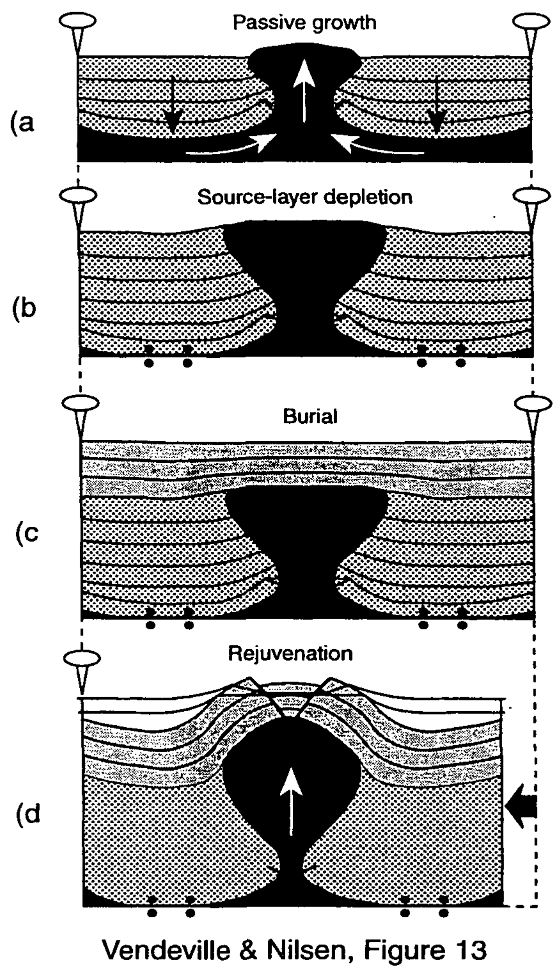


Figure BV.17 Apparent active growth of a diapir rejuvenated by lateral shortening after source-layer depletion and burial under a thick roof. From Vendeville and Nilsen, Episodic Growth of Salth diapirs Driven by Horizontal shortening, GCSSEPM Foundation 16th Annual Research Conference, Salt, Sediment, and Hydrocarbons, (1995). Reprinted with permission of the Gulf Coast Section Society of Economic Paleontologists and Mineralogists Foundation.

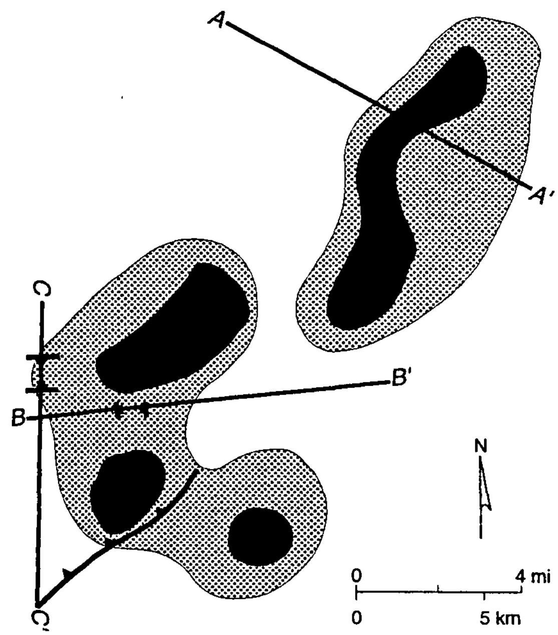


Figure BV.18 Map of salt diapirs rejuvenated by late shortening in the Nordkapp Basin, Barents Sea, illustrating the location of seismic sections shown in Figure 19. Black: diapirs close to the Plio-Pleistocene erosion surface; gray: base of the Cretaceous diapir roofs uplifted above regional datum. (K.T. Nilsen, B.C. Vendeville, and J-T. Johansen, AAPG Memoir Series No. 65, AAPG © 1995, reprinted by permission of the American Association of Petroleum Geologists).

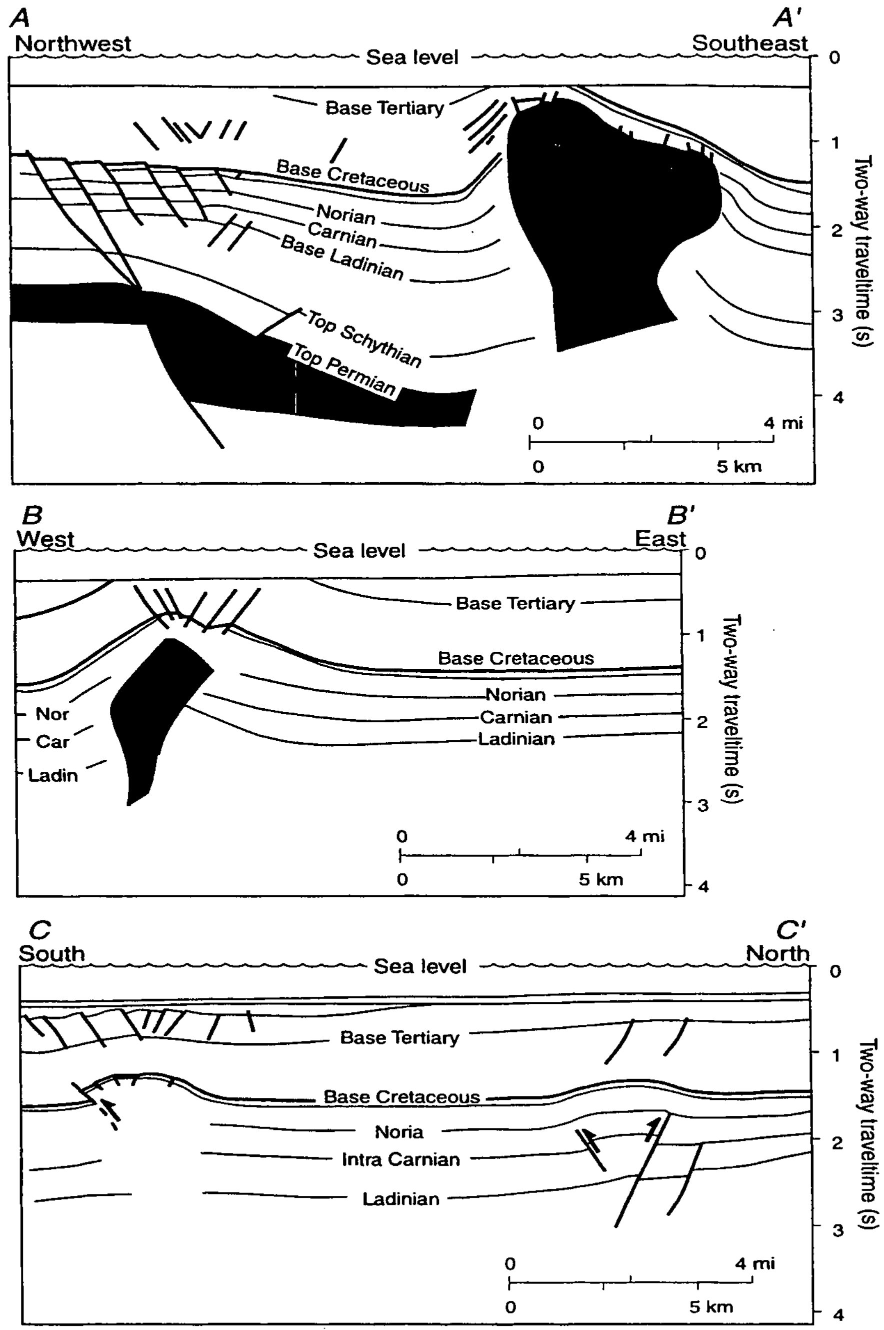
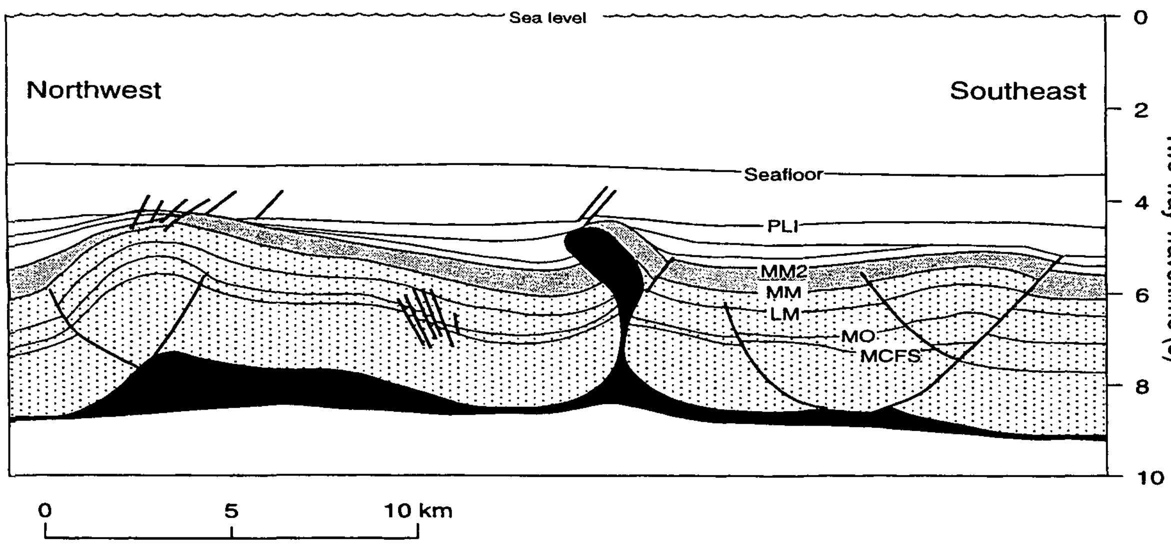


Figure BV.19 Line drawing from seismic sections located in Figure 18. In sections A-A' and B-B', which intersect preexisting diapirs, shortening was accommodated by squeezing and rejuvenating the diapirs. No folds or reverse faults formed. In section C-C', which does not intersect any preexisting diapir, shortening was accommodated by two anticlines cored by reverse faults. (K.T. Nilsen, B.C. Vendeville, and J-T. Johansen, AAPG Memoir Series No. 65, AAPG © 1995, reprinted by permission of the American Association of Petroleum Geologists).



Vendeville & Nilsen, Figure 4

Figure BV.20 Line drawing from seismic section, Atwater Valley, Lower continental slope of the northern Gulf of Mexico. Gravity gliding-spreading caused late lateral shortening that squeezed and rejuvenated the diapir and faulted and folded the adjacent sediments. MCFS = Middle Cretaceous flooding surface; MO = Middle Oligocene; LM = Lower Miocene; MM1 = Early Middle Miocene; MM2 = Middle Middle Miocene; PLI = Pliocene. Modified from Wu (1993). Permission to use this copyrighted material is granted by the author.

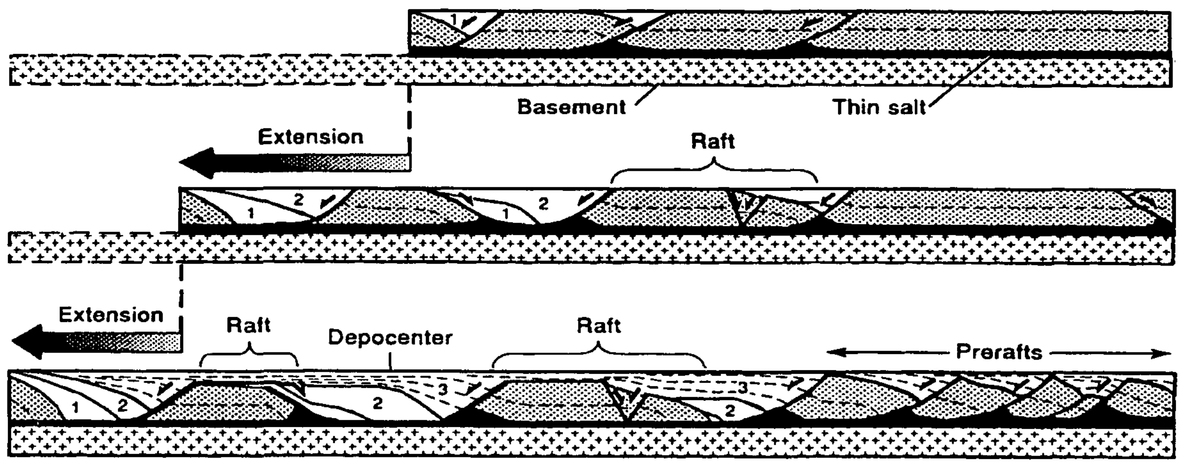


Figure BV.21 Extreme thin-skinned extension above salt. Reprinted from Jackson and Talbot (1991), A Glossary of Salt Tectonics, Geological Circular 91-4, with permission from the Bureau of Economic Geology.

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APPENDIX B

IDENTIFYING FAULTS AND DETERMING THEIR ORIGINS
SUMMARY OF WORKSHOP, JULY 1-2, 1996

IDENTIFYING FAULTS AND DETERMINING THEIR ORIGINS

SUMMARY OF WORKSHOP, JULY 1-2, 1996

by

K. Kelson, K. Hanson, and M. Angell

On July 1 and 2, 1996, Geomatrix Consultants and William Lettis & Associates convened a workshop designed to help identify diagnostic characteristics of nontectonic faults, and develop criteria with which to differentiate these features from tectonic faults. The workshop was sponsored by the U.S. Nuclear Regulatory Commission (NRC) under a research grant to Geomatrix Consultants and William Lettis & Associates. The grant is designed to assist the NRC staff in the review of licensing applications and other issues related to nuclear facilities. In the past, determinations concerning the origins of faults have been a concern in nuclear materials facilities investigations throughout the United States, including: (1) glacial features in the northeastern U.S., (2) karst and collapse features in the eastern U.S., and (3) subsidence-related features in the Gulf Coast region. The recent identification of seismogenic blind thrusts in the western U.S. also highlights the need to develop criteria to differentiate between primary and secondary coseismic deformation. The development of criteria to differentiate between tectonic and nontectonic faults in the geologic record will help the NRC address similar issues in the future.

The workshop was held at the Embassy Suites Hotel in South San Francisco, California, and was attended by 51 researchers from academia, government, and private industry. A list of participants is attached to this memorandum. The workshop was facilitated by Kathryn Hanson and Michael Angell of Geomatrix Consultants, and Bill Lettis and Keith Kelson of William Lettis & Associates. An informal pre-workshop field trip to the San Andreas fault took place on June 30, 1996. The two-day workshop consisted of two sessions: Characterization of Nontectonic Phenomena (on July 1), and Criteria Development (on July 2). During both sessions, workshop participants were divided into small working groups of 6 to 12 people, followed by summary discussions among all participants. A group discussion considering the application of criteria followed both of these sessions. A meeting agenda is attached to this memorandum.

The major issues and ideas from the workshop will be synthesized by Geomatrix Consultants and William Lettis & Associates to further develop criteria for identifying faults and determining their origins. Anticipated products from the synthesis of workshop conclusions, and additional research, will include a report containing guidelines for identifying faults, applying diagnostic criteria, and assessing uncertainties. The report likely will include either flowcharts, tables, or logic trees that will help a user apply the criteria and assess uncertainties. The following text provides a summary of the primary results and conclusions derived from the field trip and workshop.

PRE-WORKSHOP FIELD TRIP: THE SAN ANDREAS FAULT IN PORTOLA VALLEY

Approximately 20 workshop attendees participated in a field trip to the Blue Oaks site in Portola Valley, located about 40 km southeast of San Francisco. This site, which has been studied extensively for a proposed residential development, contains features that are related to surface rupture along the San Andreas fault, as well as features related to nontectonic ridge-top spreading. The site provides a good opportunity to compare and contrast tectonic and nontectonic features, and thus was an appropriate prelude to the workshop. The field trip was led by Bob Wright of Harlan Tait Associates and Bill Cotton of William Cotton and Associates.

The Blue Oaks site is traversed by two strands of the San Andreas fault, one of which experienced ground rupture during the 1906 earthquake. The field visit included a reconnaissance of the 1906 rupture trace, which is characterized by numerous features that are common to strike-slip faults, including linear topographic scarps, linear closed depressions, and offset drainages. Discussion centered on the relative importance of various investigative techniques for identifying active fault traces, such as trenching, shallow boreholes, geophysics, and detailed geomorphic mapping. It was concluded that a critical part of delineating active fault strands is basic geologic mapping, supplemented by trenching and other activities at selected sites.

The field trip also included a reconnaissance of a linear depression located at the top of an adjacent ridge several hundred meters southwest of the fault strands. The ridge-top feature parallels the San Andreas fault and in many ways is similar to linear features along the fault. Results of detailed geologic mapping and trenching across the linear depression suggest that the feature is not directly related to tectonic surface faulting, and likely is a result of ridge-top spreading (perhaps induced by strong ground shaking). Participants generally concluded that the height and linearity of the ridge-top scarp are similar to scarps produced by tectonic surface rupture, and that field mapping and trenching may be necessary to differentiate similar features at other sites from fault-related features.

The field trip concluded with a brief visit to a trench across the 1906 trace of the San Andreas fault in the village of Los Trancos Woods, approximately 1 km southeast of the Blue Oaks site. Although the trench walls had not yet been cleaned and logged, a shear plane coincident with a southwest-facing topographic scarp was visible in the trench. It is likely that this is the 1906 trace of the fault, although many participants agreed that the trench exposure alone was insufficient to definitively identify the shear plane as a tectonic fault rather than as a landslide-related feature.

WORKSHOP INTRODUCTION - (MONDAY, JULY 1, 1996)

Hanson and Lettis welcomed participants to the workshop and introduced members of the project team, including expert panel members and NRC staff. Lettis described the objectives of the workshop: (1) to explicitly characterize nontectonic features that could be interpreted as tectonic features, (2) to develop criteria that differentiate between features produced by nontectonic and tectonic processes, and (3) to address the uncertainties in the criteria and their application. For the purposes of the workshop, the term **tectonic fault** was defined as a fault produced by crustal processes acting at seismogenic depth ($\geq$5 km), and a **nontectonic fault** was defined as a fault produced by shallow crustal or surficial processes at depths of less than 5 km. In addition, the term **seismogenic fault** was defined as a fault that can produce a significant earthquake ($\geq M_W$ 5), and **nonseismogenic fault** was defined as a fault that cannot produce a significant earthquake. These working definitions were presented to provide a framework for the workshop sessions, and to minimize the amount of time spent on defining terms. These definitions are informal definitions that have not been adopted by the NRC.

SESSION I: CHARACTERIZATION OF NONTECTONIC PHENOMENA

The first session focused on identifying characteristics of nontectonic faulting and the most useful techniques for their identification. Participants were divided into seven groups, each addressing the identification of features related to different nontectonic phenomena (as described below). After these group discussions, participants reconvened for short summaries and a discussion of each phenomena. The results and conclusions from each group are summarized below.

1. Large-scale Gravity- and Salt-related Phenomena

In general, large-scale gravity- and salt-related phenomena consist of: (1) gravity glide (translational) features, (2) salt or shale diapiric features, and (3) salt dissolution features (also covered by the group on Subsidence and Collapse Phenomena, see below). These features generally are associated with:

- Presence of subsurface evaporite strata
- Extensional tectonic settings
- Shallow topographic slope (1° to 5°, usually 3°)
- Large lateral extent (commonly as much as 150 km wide)
- Listric normal faults at upslope boundary and in glide mass (growth faults)
- Normal faults that are arcuate in map pattern
- Faults that do not extend below soluble strata
- Numerous individual glide blocks within regional gliding mass
- Regional basal detachment fault
- Salt diapirs and "buckle" folds in downslope areas that accommodate contraction
- Drape folds, missing stratigraphy (stratigraphic gaps), and salt diapirs where salt strata are thin and salt has flowed over subsurface relief produced by basement faulting
- Presence of hydrocarbon resources
- Linear salt-cored anticlines bordered by normal faults that dip toward anticline crest
- Potentially high rates of strain (millimeters/year to meters/year)
- Long duration of deformation (as much as hundreds of millions of years)
- Low levels of seismicity and earthquake magnitudes less than M5 because of inability of salt to accumulate elastic strain

The majority of these features have been identified through extensive shallow and deep seismic reflection profiling and drilling performed primarily by the hydrocarbon-exploration industry. Regional and local geologic field mapping also are important for identifying salt dissolution features. Regional mapping and drill hole data compilation at a scale comparable to the large, regional glide structures and basal detachments may be required to associate observed features with salt-related deformation. Level-line surveying and geodetic methods may also help identify rates of strain that are high relative to regional crustal strain.

2. Landslide Phenomena

Landslides are associated with many features that, individually, are very similar to those produced by tectonic surface rupture. Collectively, however, landslide-related features form patterns that may enable differentiation from tectonic features. Key landslide-related characteristics include:

- Association with topographic relief to provide gravitational potential energy (i.e., mountainous terrain, often within tectonically active regions)
- Listric basal plane of detachment (slide plane)
- Generally shallow depth/thickness (estimated maximum ≈ 200 m)
- Listric normal fault at head scarp, dextral fault at right margin, sinistral fault at left margin, and reverse fault at toe (all or none of which may be preserved)
- Lateral discontinuity (although many large slides may be at kilometer scale)
- Hummocky geomorphology
- Common association with springs
- Slide-boundary faults may be indistinguishable from tectonic faults in trench exposures
- Commonly high strain rates (millimeters/year to meters/day)
- Nonseismogenic, but often induced by seismic shaking
- Single-couple focal mechanisms (opposed to double-couple seismogenic mechanisms)

Landslide features commonly are identified through analysis of aerial photography and local geologic field mapping. Investigative techniques also include small- and large-diameter boreholes to assess material characteristics, depth to groundwater, and presence of a basal slide plane. Regional geologic mapping at a scale comparable to large, regional slide structures and basal detachments may be required to identify landslides and to differentiate them from tectonic features. Level-line surveying may also help identify rates of strain that are high relative to regional crustal strain or local fault slip.

3. Subsidence and Collapse Phenomena

Features produced by subsidence and collapse can be grouped into five process-oriented categories. All of the features produced by these processes may take advantage of pre-existing fault planes, and may be characterized by continuous rather than episodic movement. Most of these features are not linear over great distances, although structurally controlled dissolution may produce linear features that are continuous for tens of kilometers. Virtually all of the faults generated by subsidence and collapse are characterized by normal displacement; observations of strike-slip and/or reverse displacement may indicate tectonic faulting. Lastly, all of these features may occur regardless of tectonic environment, and thus displacements due to subsidence or collapse may differ from those related to regional tectonic stresses. Other features specific to the individual processes are given below:

a. Fluid (e.g., groundwater, hydrocarbons) withdrawal features

- Topographic scarps (heights as much as 1 m and lengths as much as 16 km)
- Extension cracks (lengths as much as 2 km)
- Slow strain rate (growth via creep)
- Temporal association with episodes of historic fluid withdrawal
- Development correlates with timing and magnitude of fluid-level changes
- Presence of unconsolidated, compactible sediments
- Association with seismicity ($\leq$ M5) in hydrocarbon fields, possible association with events larger than M5

b. Substratum dissolution features (evaporite, carbonate)

- Presence of soluble material in subsurface
- Local scarps in concentric pattern, usually associated with normal displacements
- Linear zones of faulting along margins of soluble formation
- Graben-like faulting extends only to depth of dissolution
- Caprock residue deposits derived from dissolution of salt
- Evidence of continuous movement (creep)
- Long history of movement (as much as hundreds of millions of years)
- Absence of associated seismicity

c. Ice contact (glacial) features

- Normal faults with quasi-circular map pattern ("kettles"); rare reverse faults
- Evidence of glaciation (e.g., outwash deposits, glacial striae, eskers, etc.)
- Faulting does not extend below base of glacial outwash
- Absence of associated seismicity

d. Hydrocompaction features

- Circular topographic depressions
- Normal faults and tension cracks in circular map pattern
- Faults extend only to shallow depth
- Associated with areas where water has been applied: usually from ponding or surface saturation, but possibly from rise in groundwater table
- Absence of associated seismicity

e. Mining-related features

- Evidence of historic mining
- Circular depressions bordered by normal faults and tension cracks

- Linear depressions may form above collapsed tunnels or adits
- Faults restricted to formations above mined interval
- Seismicity occurs as "rock bursts", generally < M5

f. Volcanic collapse features

- Presence of volcanic caldera deposits (e.g., ignimbrites)
- Circular faults bordering caldera ("ring" faults)
- Circular topographic depression (as much as tens of kilometers in diameter)
- Ring faults typically concentric with, but within, topographic depression
- Possible association with seismicity

Techniques that may enable identification of these features include geologic and geomorphic mapping, borehole logging and correlation, gravity and magnetic surveys, seismic reflection profiling, geodetic surveying to document fault creep, analysis of fluid-level changes, trenching of ice-contact features, analysis of aerial photography, grain size analysis of unconsolidated sediments, historical research on mining activities, and analysis of heat flow for volcanic features.

4. Loading and Unloading Phenomena

Features produced by loading and unloading of the Earth's crust are related to several mechanisms: addition or loss of glacial ice, sediment, or water; changes in local stress fields from excavations; and changes in far-field stresses. Much of the discussion during this session focused on the scale and characteristics of faulting observed in glaciated and post-glacial isostatic recovery regions. Large-displacement surface faulting events in such regions, which are well documented in Fennoscandia and Scotland, are characterized by:

- Reactivation of pre-existing zones of brittle deformation
- Variable dimensions - in Fennoscandia 100's km long (3 belts approximately 100 km apart); in Scotland - km to 10 km long (a few km apart), greater topographic relief (1,000 m) in Scotland may be a factor
- Single event large displacement scarps - maximum 15 m displacement on the Parvie fault in northern Sweden
- High-angle dip-slip (reverse) displacement
- Orientation and sense of displacement on faults in Fennoscandia are consistent with contemporary stress regime.
- Temporal clustering of activity - early post-glacial approximately 9 ka
- Large displacement faults in Fennoscandia are inferred to be the result of M>7.5 earthquakes; spatial and temporal association with other deformation (contorted lacustrine layers; liquefaction, and landslides) suggest strong ground deformation
- Little or no association with historical seismicity

The absence of comparable large-displacement faults in Canada likely is a result, in part, to the scale and inaccessibility of the region in which they may occur, the lack of detailed mapping or reconnaissance throughout much of this region, and the difficulties in distinguishing from aerial reconnaissance or photo-geologic interpretations of post-glacial surface ruptures from fault-line erosion that occurred during and after glaciation. The recent surface rupture that occurred during the 1989 Ms 6.3 Ungava earthquake shares many of the same characteristics as the early post-glacial surface faults: a reactivated pre-existing ductile shear zone, no evidence for multiple surface faulting events, and no significant prior associated microseismicity. Unlike the Fennoscandian faults, the north-south contraction produced by the Ungava earthquake appears to deviate from the orientation of maximum horizontal compression axis inferred from the regional stress data. However, the majority of these stress data lie south of the Great Lakes/St. Lawrence River region, and the more local data indicate the axis of maximum horizontal compression in northern Canada is oriented approximately north-south.

Small-scale faults (pop-up structures) observed in post-glacial recovery areas and areas that experienced loading/unloading due to pre-glacial lakes have characteristics that include:

- Short length (1-2 km)
- Low height (1-2 m)
- Generally produced by a single event; initial rupture is episodic but, according to many quarry managers, some continue to grow slowly (creep)
- Low magnitude (<M5) earthquakes
- Triangular-surface shape (small symmetric bedding kinks)
- Deformation decreases downwards into voids
- Sometimes aligned along pre-existing faults; (e.g., pop-up structures in Ohio are associated with reverse faults at depth)
- Shallow extent (<1 km); faults become listric at depth, underlain by shallow detachment fault
- Other glacio-tectonic structures that may be confused with tectonic faults include ice-push deformation. Large blocks (km-scale) of bedrock as well as sediment may be involved. Drilling or other subsurface investigations may provide a means to demonstrate the limited subsurface extent of such features.

Seismicity and surface or near-surface deformation related to human-engineering activities, such as reservoir impoundments, mining or large scale quarrying, were noted but not discussed in detail during this session. Also, the localization of post-Cretaceous faulting and Quaternary deformation along the northeastern Atlantic coastal margin at the hinge zone between sediment loading in offshore basins and uplifted regions characterized by regional denudation was discussed.

5. Volcanic-related Phenomena

Features produced by volcanic processes that may be misinterpreted as tectonic surface rupture generally are restricted to volcanic rift zones and are related to dike intrusion. These zones commonly include or are associated with:

- Presence of volcanic deposits
- Linear depressions filled with basalt

- Topographic scarps as much as 10 m high, but with variable height along strike
- Monoclinal folds along strike of scarps
- Scarps within zones generally 10 km long, but can be up to 100 km long
- Overall subdued topographic relief, scarps commonly buried by volcanic deposits
- Common tensional fissures and cracks parallel to scarps, usually more fissures than scarps
- Rift graben up to 2 km wide
- Monoclinal folding along margins of graben
- Fault rupture in short segments
- Little or no scarp-derived colluvium because of the lack of surficial deposits and the resistance of basalt to erosion
- No net displacement across graben
- Seismicity generally $\leq$ M4.5, and occurs as swarms related to dike intrusion
- Focal mechanisms suggesting normal displacement

Identifying the presence of volcanic rocks is critical to evaluating volcanic phenomena. This is most easily accomplished by regional and local mapping, and analysis of aerial photography. Swarms of microseismicity can be detected by local detailed seismic networks if the dike-intrusion process is still active. Seismic reflection investigations generally are not successful in these types of volcanic terranes, although studies of heat flow and gravity may be helpful in delineating regional volcanic sources. Trenching is difficult in areas of outcropping volcanic rocks and no overlying surficial deposits.

6. Secondary Coseismic Deformation

Coseismic deformation commonly includes features that are not directly related to surface fault rupture. These features are viewed herein as tectonic, but not seismogenic. However, some faults may experience secondary deformation during earthquakes on other faults, in addition to producing moderate to large earthquakes and the accompanying primary coseismic deformation. Secondary deformation features may occur in contractional, extensional, and strike-slip tectonic settings, as described below.

a. Contractional settings

- Flexural slip faulting (movement along bedding planes within fold deformation)
- Bending moment faulting (related to extension in crest of fold)
- Hanging wall collapse (normal faulting in hanging wall of reverse fault)
- Conjugate strike-slip faults at reverse fault terminations (cf. El Asnam earthquake)
- Tear faults (accommodating differential movement between thrust blocks)
- Axial surface faults
- Chordal faults that span a frontal thrust (cf. Meckering earthquake)

b. Extensional settings

- Antithetic normal faults
- Synthetic faults (parallel to primary rupture)

- Small antithetic thrust faults in hanging wall of normal fault
- Bending moment faulting (related to extension in crest of drape fold)
- Fault swarms, commonly in transfer zones between en echelon normal faults

c. Strike-slip settings

- Splay faults at fault terminations (e.g., "horsetail" splays)
- Reidel shears
- Contractional and extensional secondary faults in restraining and releasing bends or stepovers

Identifying and characterizing secondary deformation requires geologic and geomorphic mapping at a scale that is appropriate to identify and characterize primary deformation as well. Unfortunately, geomorphic evidence of secondary deformation commonly is ephemeral, and most secondary features are poorly preserved in the geologic record. Characterization of secondary features also requires placing them in regional context, which may require kinematic analysis to assess if their formation is consistent with the regional style of primary deformation. The role and importance of secondary deformation must be understood in the context of the regional strain field.

7. Strong Ground Motion Phenomena

Strong ground motion may cause a suite of features that are similar to those produced by primary deformation, including ground cracks and landslides (see above). As with secondary deformation, the effects of strong ground motion must be understood in the context of the regional strain field. In addition, strong ground motion may produce many liquefaction-related features, such as:

- Lateral spreads
- Sand blows
- Clastic dikes
- Ground subsidence

Although these features are not a result of primary surface rupture, they nevertheless represent potential surface rupture hazards and are indicative of strong vibratory ground motion. Techniques used to investigate these features include detailed geologic and geomorphic mapping, trenching to expose structural and stratigraphic relationships, laboratory analyses of extruded sand, geochronology, and investigations of liquefaction susceptibility.

SESSION II: DEVELOPMENT OF CRITERIA TO DIFFERENTIATE TECTONIC AND NONTECTONIC PHENOMENA

The second session of the workshop took place on July 2 and focused on identifying the diagnostic characteristics of the nontectonic phenomena discussed in the first session, and using these to develop criteria to differentiate between tectonic and nontectonic phenomena. In order to emphasize the importance of and techniques of incorporating uncertainty into the assessment of seismic hazard, Kevin Coppersmith presented a case history in which uncertainties in the seismogenic potential of faults, structures, and tectonics features were addressed for seismic hazard analyses in the eastern United States. As part of the seismic hazard analyses for the Electric Power Research Institute (EPRI), an explicit characterization was made of the seismogenic potential (termed the "probability of activity") of each tectonic feature believed to have some potential to generate M>5 earthquakes. The EPRI assessments were made by six expert teams and were structured to provide a quantitative expression of uncertainty suitable for direct incorporation into a probabilistic seismic hazard analysis for ground motions. The basic steps in the methodology were:

- Identification of faults, structures, and tectonic features that have some potential for generating M>5 earthquakes;
- Identification of criteria that can be used to assess the seismogenic potential of tectonic features (e.g., association with seismicity, evidence for geologically recent slip, evidence for brittle reactivation consistent with contemporary tectonic regime, etc.);
- Evaluation of the relative usefulness of the various criteria in their ability to assess activity/seismogenic potential;
- Feature-by-feature assessment of the degree to which a specific feature displays evidence for the diagnostic criteria;
- Assessment of the probability of activity/seismogenic potential of each feature.

The process of identification of diagnostic criteria, evaluation of the relative value of the criteria, and assessment of each tectonic feature relative to these criteria is similar to the process is being followed in the present NRC project. The EPRI study shows that this assessment can be made in a quantitative manner that incorporates uncertainties for seismic hazard analyses, and that this approach has been used in a regulatory setting for decisions regarding critical facilities.

Following presentation of the EPRI study, participants were divided into five groups, each of which addressed specific nontectonic phenomena (as described below). After these group discussions, participants reconvened for short summaries of criteria for each phenomena. The results and conclusions from each group are summarized below.

1. Large-scale Gravity, Landslide, and Sackungen Phenomena

This group observed that the identification of a basal slide plane is a highly diagnostic criterion for differentiating between tectonic and landslide-related features. Because a slide plane has either reverse, normal or strike-slip movement depending on its location adjacent to the slide mass, the identification of a fault as a slide plane is dependent on documenting its geometry and continuity in the subsurface beneath the slide mass. For example, a reverse fault at the base of a slope that decreases in dip with increasing depth and may become parallel to the overlying slope should be suspected as part of a basal slide plane. Conversely, if a reverse fault at the base of a slope extends into the subsurface at increasingly greater dips with depth, then it likely is a tectonic reverse fault rather than a slide plane. Other criteria that were believed to be moderately diagnostic include:

- **Geomorphology:** The presence of a headscarp, toe bulge and hummocky topography provides evidence of mass movement;
- **Slide headscarp or toe length and linearity:** The lengths of slide headscarps or toes typically are shorter than surface-rupture features associated with tectonic faults;
- **Ratio between scarp height and length:** Ratio generally larger ($>10^{-2}$) for

landslide-related features than for surface-rupture features ($<10^{-4}$);

- **Slide plane depth:** Landslides typically are less than 200 m thick, although thickness may be influenced by subsequent erosion or reactivation of parts of a slide mass;

- **Displacement per event:** Slide headscarp or toescarp from a single movement event may be larger than that from a surface rupture event (i.e., vertical separations associated with faulting events are commonly less than 6 m);

- **Map pattern and sense of slip of boundary faults:** the presence of normal, reverse, dextral, and sinistral faults surrounding an area suggest landslide origin;

- **Sedimentology of slide mass:** Landslide deposits commonly are jumbled and disarticulated, and overlie intact coherent materials below the slide plane;

- **Evidence of vertical creep:** There are no documented examples of dip-slip creep along tectonic faults, but many landslides are characterized by active downslope creep;

- **Seismicity:** Landslides generally are aseismic, although some extremely large landslides may be associated with large earthquakes;

- **Focal mechanism:** Earthquakes are associated with double-couple focal mechanisms, landslide are associated with single-couple mechanisms.

Differentiation between tectonic and sackungen features may be made by several criteria that are similar to those for landslide-related features. None of the criteria is diagnostic, although use of a collection of criteria probably will provide sufficient evidence for differentiation. Criteria that are moderately diagnostic include:

- **Map-view continuity:** Sackungen generally are discontinuous, whereas tectonic fault scarps typically are more continuous;
- **Length:** Sackungen generally are on the scale of a few kilometers (< 10 km), but tectonic faults typically are longer than 10 km;

- **Ratio between scarp height and length:** Ratio generally larger ($>10^{-4}$) for sackung features than for surface-rupture features ($<10^{-4}$);

- **Number of scarps:** Sackungen commonly have numerous scarps, whereas tectonic faults may have only one scarp associated with recent active trace;

- **Relief:** Sackungen require relief for formation, faults may be present across areas with little or no local relief;

- **Displacement history:** Sackungen may form by continuous or episodic movement, but fault scarps are formed by episodic, sudden events;

- **Sense of slip:** Sackungen typically are associated with normal displacement, which may be diagnostic if faults in region dominated by strike slip or reverse displacement;

- **Seismicity:** Sackungen are aseismic, although some may form during large earthquakes.

As during the first session, this group emphasized that many of these criteria are best tested via local and regional geologic and geomorphic mapping. Drilling and geophysical investigations may provide data on the presence and geometry of a basal slide plane. Trenching may also provide data on the sense of slip of landslide boundary faults, and thus help ascertain the kinematic history of landslide or fault movement. Trenching across possible sackungen features also may provide information on the deformational history and sense of fault slip at a site. Analysis of seismicity data, if present, may address the type and depth of faulting, and enable construction of focal mechanisms.

2. Subsidence and Collapse Phenomena

The first workshop session suggested that features produced by subsidence and collapse can be grouped into five process-oriented categories (see above). For each of these categories, several criteria were identified, as given below.

a. Fluid (groundwater, hydrocarbons) withdrawal features

- Spatial association with areas of historic fluid withdrawal;
- Temporal association with timing of fluid withdrawal;
- Association with pre-existing faults in sedimentary basin;
- Presence of compactible sediments;
- Presence of extensional faults, absence of reverse and strike-slip faults;
- Faults are usually concentric but may be linear;
- Presence of tension fractures throughout sedimentary basin;
- Evidence of continuous deformation (creep);
- Presence in a tectonically quiescent area.

b. Substratum (evaporite, carbonate) dissolution features

- Presence of soluble substratum;
- High-angle normal faults in linear, en echelon or arcuate, concentric patterns;
- Circular or oblong geomorphic depressions at large and small scales;
- Strata dipping concentrically inward toward geomorphic depression;
- Faults that die out in the shallow subsurface and do not extend beneath soluble strata (generally less than 3 to 5 km);
- Evidence of continuous movement (creep) along faults;
- Possible reactivation of pre-existing faults;
- Proximity to known solution features;

- Large variability in amount of net separation and high degree of local closure;
- Seismicity related to collapse occurs only within or above soluble strata.

c. Ice contact (glacial) features

- Association with glacial landforms and deposits;
- Irregular shapes and distributions of depressions;
- Presence of faults that die out in the shallow subsurface and do not extend beneath glacial strata;
- High variability in strike and dip along faults and shear surfaces;
- Commonly associated with normal faults, but reverse faults also present;
- Timing of deformation coincident with melting of ice;
- May occur in aseismic area.

d. Hydrocompaction features

- Presence of compactible soils, usually unconsolidated Holocene loess or alluvium;
- Presence of shallow groundwater table or source of water at the ground surface;
- Commonly occur in arid to semi-arid climates;
- Presence of high-angle normal faults in arcuate, concentric patterns;
- Presence of circular or oblong geomorphic depressions at large and small scales;
- Presence of strata dipping concentrically inward toward geomorphic depression;
- Presence of faults that die out in the shallow subsurface and do not extend beneath compactible strata;
- May occur in aseismic area.

e. Mining-related features

- Evidence of historic mining;
- Circular or linear depressions bordered by normal faults and tension cracks;
- Faults restricted to formations above mined interval;
- Scarps and depressions generally tens to hundreds of meters long;
- Potentially large displacements relative to tectonic strain rates;
- Seismicity occurs as "rock bursts", generally < M5

f. Volcanic collapse features

- Spatial and temporal association with magmatic processes;
- Presence of dikes or other volcanic deposits;
- No net vertical displacement across caldera or collapse zone;
- Presence of ring faults with inward-dipping geometries;
- Predominance of normal faults, few or no strike-slip or reverse faults;
- Association with high heat flow and geothermal resources;
- Presence of circular topographic depression (as much as tens of kilometers in diameter).

The group believed that it was possible to differentiate features related to subsidence or collapse from those related to tectonic surface rupture. In particular, the group felt that ice contact, hydrocompaction, and mining-related features generally are more easily differentiated from tectonic features than the other subsidence/collapse features. Techniques that may enable identification of these features include geologic and geomorphic mapping, borehole logging and correlation, gravity and magnetic surveys, seismic reflection profiling, geodetic surveying to document fault creep, analysis of fluid-level changes, trenching of ice-contact features, analysis of aerial photography, grain size analysis of unconsolidated sediments, and historical research on mining activities.

3. Loading and Unloading Phenomena

Diagnostic characteristics were identified by this group for glacio-tectonic faulting, "pop-ups", and ice-shove features. There also was discussion of loading/unloading phenomena in unglaciated intraplate regions, and methods to identify and characterize Quaternary deformation related to these phenomena. The group suggested that, because none of the characteristics may be diagnostic alone, differentiation between tectonic and nontectonic phenomena may require use of a suite of criteria. These characteristics are listed below.

a. Glacio-tectonic faults in post-glacial isostatic recovery zones

- Presence within glaciated region;
- Commonly within intraplate regions and compressional stress regime;
- Occurrence within areas having little or no seismicity;
- Rarely associated with historic tectonic surface ruptures;
- Occurrence of faulting during or immediately after deglaciation;
- Commonly high-angle reverse faulting;
- Commonly associated with reactivation of existing faults;
- Fault scarps usually are produced by single events (none identified to date with multiple events);
- Fault scarps generally are high for their length (i.e., a high ratio between scarp height and length);

- Faults commonly displace glacial and late glacial features.

b. "Pop-up" features

- Typically symmetric bedding kinks;
- Usually smaller than 1 km in length;
- Displacement along faults decreases down dip;
- Usually underlain by a shallow detachment fault.

c. "Ice-shove" features

- May involve bedrock and/or surficial deposits;
- Limited down-dip extent.

The group acknowledged that the characteristics of large-displacement coseismic surface ruptures that have been documented in regions of post-glacial isostatic recovery are not unique. There is much uncertainty in precluding the potential for reactivation of faults in glaciated stable continental regions. The relatively low rate of seismicity, long recurrence intervals, and limited stratigraphic record (generally only the Holocene) limit the ability to identify and characterize active faults in glaciated regions. Documentation of evidence for the presence or absence of strong-ground motions may provide additional information to assess faults in such regions.

In regions of low tectonic activity, regional mapping and geomorphic evaluations of geomorphic systems (fluvial terraces, drainage system patterns, etc.) may provide a means to characterize the spatial and temporal patterns of Quaternary deformation. Deformation may include warping or folding of glacial features or deposits, and regional scale flexure due to sediment loading and/or denudation/erosion. Geologic mapping of bedrock is critical for identifying pre-existing structures, and subsurface investigations (drilling and trenching) may provide data to evaluate the down-dip geometry and extent of surface faults, and the timing and history of displacement.

As summarized by David Ferrill, a technique referred to as slip-tendency analysis has been developed recently that allows for the assessment of slip potential from mapped and suspected faults in a known or inferred state of stress. This technique can be used to assess seismic risk from known or suspected faults, test for compatibility of geologic structures, focus exploration for high-risk and earthquake-prone blind faults, and help interpret faults from focal mechanism solutions.

4. Volcanic and Salt Diapiric Phenomena

This group focused on developing criteria to differentiate between tectonic features and those produced by upward-mobile materials, such as magma and salt. The most diagnostic criteria for the identification of the nontectonic features is the presence of either volcanic deposits or subsurface salt. The list below also includes moderately diagnostic criteria.

a. Volcanic features

- Presence of volcanic deposits (highly diagnostic);
- Presence of elevated heat flow (moderately diagnostic);
- Presence of rift zones up to 2 km wide and tens of km long;
- Topographic scarps, graben, and abundant tension fissures within rift zones;

- A lack of net tectonic displacement across the rift zone.

b. Salt-related features

- Presence of salt diapir or salt ridge at the surface (highly diagnostic);
- Presence of salt layer, ridge, or diapir in subsurface (highly diagnostic);
- Presence of concentric or radial faults (highly diagnostic);
- Presence of low gravity anomaly;
- Occurrence of a "bright" seismic reflector at the top of rising salt;
- Presence of thickened and thinned strata adjacent to linear salt ridge;
- Normal faults and rollover anticlines along linear salt ridges;
- Normal faults that terminate at shallow depth within salt diapir;
- Contemporary seismicity restricted to depths of salt and suprasalt strata;
- Lack of moderate- and large-magnitude contemporary seismicity ($M \leq 5$).

Many of these criteria require data that can be obtained by local and regional geologic mapping. Drill hole and seismic reflection data may provide subsurface information critical to identification of subsurface salt and adjacent truncated strata. Collection of gravity and heat flow data may also be useful in assessing density or thermal anomalies that are produced by salt or magma. These investigative techniques have various levels of feasibility and cost; the group suggested that geologic mapping to assess the characteristics of deformation and their relations to regional features may be the most appropriate investigative technique.

5. Secondary Deformation and Strong Ground Motion Phenomena

"Secondary" tectonic faults were informally defined by this group as those faults that form as a "passive" mechanical response to slip on another, usually larger, fault. The group, therefore, suggested that characterization of the structural geologic context and faulting mechanism may be the most useful approach to differentiating between primary and secondary tectonic features. Assessments should take into account the scale, rate, timing, and patterns of deformation, and assess whether all of the characteristics of a certain area or feature could have been formed through primary or secondary deformation. The style and sense of primary deformation should be consistent with the regional stress field, whereas secondary deformation may reflect the influence of local, *in situ* stresses.

Ground deformation within the Potrero Canyon area produced by the 1994 Northridge earthquake was presented as an example of coseismic surface rupture of ambiguous origin. In this case, topographic scarps were produced along subparallel normal and reverse faults at the margins of Potrero Canyon. Based on the probable levels of ground motion in this area from the earthquake, the depth to groundwater, map pattern of the surficial features, and characteristics of alluvium on which the scarps formed, and the kinematic interpretation of the observed ground deformation, the features are consistent with lateral spreading produced by strong ground motions during the earthquake. This example illustrates how a wide variety of data may be required to assess whether a feature or collection of features are related to primary or secondary deformation.

NRC PERSPECTIVE

Dick McMullen provided a brief overview of the significance of developing criteria to differentiate between tectonic and nontectonic phenomena. Although presently there are no pending applications related to the licensing of nuclear power plants, the results of this study will provide a basis for review of future investigations of seismic hazards to nuclear materials facilities. In addition, because research is ongoing, it is likely that features will be discovered that pertain to existing nuclear power plants and will need to be addressed. In the past, similar determinations concerning the origins of faults have occurred throughout the United States, including: (1) glacial features in the northeastern U.S., (2) karst and collapse features in the eastern U.S., and (3) subsidence-related features in the Gulf Coast region. The recent identification of seismogenic blind thrusts in the western U.S. also highlights the need to consider the existence of similar structures in the central and eastern U.S., and to develop criteria to differentiate between primary and secondary coseismic deformation.

WORKSHOP SUMMARY AND OPEN DISCUSSION

Lettis summarized the main points of the two workshop sessions, and called for an open discussion of major points. Several themes that were common to many of the groups' conclusions are listed below.

- **Use of several criteria:** Because it is rare when a single criterion adequately identifies the origin of a feature, there is a need to use combinations of several criteria and employ sound professional judgment. Direct evidence of fault origin is an exception rather than a rule, and thus many determinations rely on the preponderance of evidence and the relative validity of each criterion used in formulating a judgment.

- **Geologic context:** Many features are or are not consistent with certain geologic settings, thus making the assessment of regional tectonic framework an important part of determining fault origins. Both the scale of deformation and the sense of slip associated with a feature need to be consistent with the regional geologic framework. If not, the process may likely be related to a non-tectonic process.

- **Scale:** The areal extent of a feature is an excellent indicator of the scale of the processes by which it formed. The extent and continuity of a feature yield information on process, and thus may enable differentiation between various origins. The rate or magnitude of strain associated with a feature may also be a critical characteristic.

- **Geometry:** The geometry of a feature in three dimensions is an important characteristic that also provides information on the processes of formation. The map pattern of deformation commonly provides data to interpret the kinematics associated with the formation of a feature, and thus its origin.

- **Techniques:** For most features, if not all, the session groups identified that local field mapping of a feature is an important investigative technique. Mapping provides data on the geologic context, scale, and geometry of a feature, which are critical to assessing fault origin, as noted above.

Many groups noted that trenching may not provide appropriate data to differentiate tectonic and nontectonic features. However, criteria that utilize the sense of movement can benefit greatly from trench data. For example, the presence of subparallel dextral and sinistral faults is an important criteria to identify landslides. Trenching provides time-dependent data, such as the timing, rate, and episodicity of deformation, which are important criteria, and data on the near-surface geometry of faults and fault-related features.

- **Uncertainties:** For any given site or problem, the sources and levels of uncertainty may vary, and should be documented. Two different types of uncertainty were identified: (1) the uncertainty in the level at which a criteria is diagnostic in differentiating fault origin, and (2) the uncertainty in whether a criteria can be used for a given feature. The amount of uncertainty in determining a fault's origin is a combination of both of these types of uncertainty.

- An open discussion that focused on the techniques of identifying faults followed this summary. The advantages and disadvantages of several techniques in assessing faults were discussed, as listed below.

- **Geodetics and radar interferometry:** These emerging techniques will probably play a greater role in future research, particularly in assessments of blind thrusts and active surficial deformation. These techniques provide positive evidence of deformation, and help assess the pattern and rates of deformation.

- **Geophysical techniques:** Seismic reflection data have played a critical role in characterizing salt-related features and the three-dimensional geometry and structural associations of faults.

- **Quantitative geomorphology:** Analyses of local and regional topography, construction of stream-channel and terrace profiles, and patterns of incision may provide important data to identify and characterize tectonic deformation. Coupled with rapidly developing techniques in geochronology (e.g., cosmogenic and luminescence dating), quantitative geomorphology may yield excellent data on the spatial and temporal patterns of surficial processes. Remote sensing applications are also an excellent reconnaissance-level tool to assess the presence or absence of fault-related features and areas of possible tectonic deformation. Many studies require the construction of detailed topographic maps and drainage-basin analyses, which can be produced more accurately and efficiently by manipulation of digital topographic data.

- **Geotechnical exploration methods:** Many techniques commonly used in geotechnical investigations provide data that are applicable to fault studies. For example, cone penetrometer testing (CPT) yields detailed shallow stratigraphic data in many geologic environments, and is rapid and cost-effective. Flat plate dilatometers measure actual stress fields at a site, which may yield important information on the kinematics of active deformation.

APPENDIX C

1997 SPRING AMERICAN GEOPHYSICAL UNION MEETING

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1997 SPRING AMERICAN GEOPHYSICAL UNION MEETING

SPECIAL SESSION T22B - CRITERIA FOR DIFFERENTIATING TECTONIC FROM NONTECTONIC FAULTS

SUMMARY OF ORAL PRESENTATIONS

A special session of the 1997 Spring Meeting of the American Geophysical Union (AGU) was convened by Geomatrix Consultants and William Lettis & Associates to provide an opportunity for researchers, particularly those from central and eastern United States, to present the results of studies that address the issues related to differentiating tectonic from nontectonic faults. Michael Angell and William Lettis presided over the session that was held on May 27,1997. Case studies illustrating the multi-disciplinary techniques and approaches that have been used to infer the origin of faulting and surface deformation in a variety of geologic settings were presented in a half-day session. As summarized below the talks presented at the symposium discussed criteria to differentiate tectonic faults from surface faulting that may result from karst-related subsidence, large-scale gravitational spreading, landsliding, soft-sediment deformation, and other nontectonic mechanisms. Related studies that describe criteria to identify a seismic liquefaction origin for clastic dikes and the use of such features to identify and characterize paleo-earthquakes in the central United States also were presented.

A common theme expressed by almost all of the presenters was the need to apply a variety of criteria and to provide an integrated model that is consistent with available stratigraphic, structural, kinematic, geomorphic observations and one that works from the mechanics perspective as well. In some instances, the mechanism or origin of a feature remains uncertain, even after focussed research efforts. A summary of the oral presentations is provided below.

SESSION T22B-01: Techniques for Identifying Faults and Determining Their Origins -- NRC-Sponsored Research - Lettis, W.R., K.I. Kelson, J.N. Baldwin, K. L. Hanson and M. Angell

Lettis and others (1997, T22B-1) provided an overview of the NRC project, emphasizing the classification of faults with respect to a tectonic/nontectonic origin and seismogenic capability. That the distinction among these characteristics is critical to the manner in which the feature is treated in a regulatory environment also was stressed. Most important to these issues is the concept that the underlying mechanisms of fault formation are reflected in the attendant sedimentological and structural characteristics (processes) of the feature and that these can and should be investigated either directly in the field or indirectly through geophysical investigations. Such investigations provide data that form the basis for developing the criteria necessary to establish origin and seismogenic capability. A brief review of the diagnostic criteria that are useful in distinguishing among faults of various origin underscored the viability of this approach.

SESSION T22B-02: Geologically-Recent Faulting and Folding of Alluvial Sediments near Pearisburg, Giles County, Virginia: Tectonic Origin or Karst Subsidence in Origin? - R.D. Law, E. S. Robinson, J.S. Cyrnak, S. Sayer, R.T. Williams, J. Callis, and M. Pope

[NRC-sponsored] Detailed field investigations of a single outcrop locality of young deformation in a passive margin tectonic environment (Stable Cratonic Region of North America) were successful at defining the structural characteristics and sedimentary features of the outcrop. Temporal characteristics were well-defined by the studies (e.g., synchroneity of faulting and folding and, to a lesser extent, recency of activity). Despite the detailed nature of the studies, the origin (and driving mechanism) of the exposed structure has not been determined with certainty sufficient to support conclusion of the investigations. This lack of certainty primarily is due to the limited areal extent of the studies and the lack of bedrock exposure in the region. The presentation served to highlight the need for subregional studies and defining the local, [subsurface] geologic context of the observed deformation in addition to conducting the excellent detailed geologic and geophysical studies of a single exposure as described in the presentation.

SESSION T22B-03: Discrimination of Tectonic and Nontectonic Faults: A Case Study Involving 11 Integrated Techniques - R. Jacobi, and J. Fountain

Jacobi and Fountain (1997) began by emphasizing the difficulty in obtaining consensus among investigators and reviewers regarding not only conclusions, but also techniques and data format when working on a volatile public issue in a regulatory environment. This difficulty was directly addressed by their investigation through the use of Quality Assurance and Quality Control measures implemented by the use of Geographic Information Systems (GIS) -based data management and analysis. Explicit treatment of uncertainty concerning the data and interpretations also was emphasized. The feature being investigated has very subtle expression (geologically, seismically and on geophysical data) but the investigators were able to form substantial conclusions regarding the origin and geometry of the target feature through innovative quantitative analysis of regional geological and geophysical characteristics. Finally, the approach presented by the authors lends itself directly to explicit treatment of uncertainty in the regulatory environment, which is a critical step towards establishing the viable and conclusive nature of such investigations.

SESSION T22B-04: Evaluation of the Capability of Inferred Faults at the Rocky Flats Environmental Technology Site, Colorado - M. Angell, K. Hanson, T. Crampton, K. Coppersmith, T. Wood, W. Peregoy.

In this presentation, M. Angell provided a summary of a detailed field program to evaluate the capability of one of a series of northeast-trending faults within the Cretaceous Laramie Formation that were inferred from lithologic and geophysical data from deep boreholes at the Rocky Flats Environmental Technology Site, Colorado. The objectives of the study were two-fold: to evaluate the style and geometry of the inferred faults and to assess evidence for recency and activity. Review of both surface and subsurface data in addition to mapping and trenching investigations, provided definitive evidence of no deformation that post-dated the formation of the Rocky Flats Alluvium basal unconformity (estimated to be a minimum of 900 ± 300 ka). Analysis of the structural and stratigraphic characteristics of four zones of deformation observed within the Cretaceous section in a trench excavated across one of the inferred faults strongly indicated a nontectonic origin involving syndepositional or early post-depositional deformation due to sediment loading during deposition of a crevasse-splay unit in a proximal deltaic environment. The talk highlighted the importance of developing a stratigraphic and kinematic model that was consistent with all trench observations and regional stratigraphic studies of the Laramie Formation and stressed that this was critical to demonstrating the nontectonic nature of deformation in bedrock exposed in the trench.

SESSION T22B-05: Geologic and Geotechnical Verification for a Seismic Liquefaction Origin of Clastic Dikes in Indiana and Illinois - S. F. Obermeier.

In this overview of recent paleoliquefaction studies in the southern halves of Indiana and Illinois, Obermeier discussed the evidence for seven paleo-earthquakes that occurred in the Holocene and at least one in the latest Pleistocene. This presentation highlighted the approaches used to identify paleoliquefaction features, differentiate paleoliquefaction features that originate as the results of strong ground motion from nontectonic mechanisms, such as flooding or landsliding, and evaluate the location and magnitude of causative earthquakes. Criteria used to demonstrate a seismic origin for the dikes observed in the study areas, rather than a nontectonic mechanism, such as artesian conditions and landsliding, included: (1) similar geometric characteristics of individual clastic dikes to dikes of known seismic origin, mainly in the meizoseismal zone of the 1811-1812 New Madrid earthquakes; (2) the pattern and location of dikes in plan view on a scale of tens to thousands of meters conforms with a seismic origin; and (3) the size of dikes on regional scale identifies a "core" region of widest dikes, which conforms with severity of effects expected in a meizoseismal zone. Geotechnical analyses, which entailed testing of the properties of source sand at many liquefaction sites, were conducted to verify the seismic origin of these features and to evaluate accelerations from the epicenter to distal sites. These studies provided the basis for evaluating magnitudes and locations for several of the causative earthquakes.

SESSION T22B-06: Liquefaction Evidence for Holocene Paleo-earthquakes in Central and Southwestern Illinois - W. E. McNulty, S. F. Obermeier.

Using the approaches and criteria outlined in the preceding talk, McNulty presented the results of recent field investigations that have uncovered evidence of two large paleo-earthquakes in Illinois. One paleo-earthquake, centered 35 km NE of Springfield, Illinois, occurred between 5,900 and 7,400 yr B. P based on radiocarbon samples from the fine grained cap and possible host sediments of the emplaced dikes and sills. Another paleo-earthquake centered about 65 km ESE of St. Louis, Missouri occurred between 6,500 and 7,000 yr B.P. Both earthquakes likely were in excess of M6, but geotechnical testing of the liquefaction sites will be required to resolve the magnitudes of each.

SESSION T22B-07: Probable Gravitational (Nontectonic) Origin for Two Conspicuous Ridge-top Scarps in the Southern Coast Mountains, British Columbia - S.C. Thompson, J.J. Clague, and S. G. Evans.

Talk cancelled.

SESSION T22B-08: Differentiation of Landsliding From Seismogenic Faulting: Criteria From the Southern Rocky Mountains and Columbia Plateau -

M. M. West.

Using case studies from the Frontal fault, Gore Range, Colorado; the Roubideau Creek fault, Uncomphagre Plateau, Colorado; and the Smyrna Bench graben on Saddle Mountain anticline, Yakima fold belt, Washington, West emphasized the difficulties of differentiating between surface deformation related to the landslide process from that produced by seismogenic faults, particularly in micro- (exploratory trenches) to intermediate-scale mapping (10-100 km2). He stressed also the difficulties in areas where landsliding may mask or overprint seismogenic structures, making the assessment of the cause of recent deformation problematic. In each of the case studies, West used a variety of geologic criteria to evaluate the origin of the surface deformation, including: scale, continuity, geomorphology, stress field, mechanics (mode of failure), shear strength and pore pressure, geometry of primary and secondary deformation, stratigraphic and structural relations.

SESSION T22B-09: Quaternary Deformation Along the Criner Fault, Oklahoma: A Case Study for Evaluating Tectonic Versus Landslide Faulting -

K. L. Hanson, F. H. Swan, J.R. Wesling, and K.I. Kelson.

In this presentation, K. Hanson summarized the arguments favoring a nontectonic origin for a shear in Quaternary sediments observed during reconnaissance mapping investigations along the Criner fault in southern Oklahoma. The Criner fault, which forms the southeastern part of the 310-km long Meers-Duncan-Criner fault system, exhibits geomorphic surface expression similar to that of the Meers fault to the north that has clear evidence of Holocene and late Pleistocene reactivation. A tectonic origin for the surface deformation was initially inferred based on the location of the fault - along the projected trace of a bedrock fault that exhibits geomorphic evidence of possible Quaternary reactivation to the north and a strong photolineament to the south. Detailed mapping of the entire zone of deformation, however, showed that the variable orientations and spatial kinematic relationships of the associated faults are consistent with bounding structures typical of landslides and was not consistent with the strike of the Criner fault. Additional reconnaissance along the strong photolineament also showed that the lineament coincided with a sharp monoclinal flexure that clearly was not faulted. In this example, description of the deformation at the outcrop scale was not sufficient to evaluate the tectonic significance of the fault in Quaternary sediments. Understanding the structural relationships of bedrock deformation associated with the Quaternary fault and more regional mapping was critical. Although the study did not resolve the timing of most recent movement along the Criner fault, it concluded that the evidence for Quaternary deformation at this locality likely was not evidence for Quaternary reactivation of the fault.

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11. ABSTRACT (200 words or less)

Satisfying regulatory criteria for siting nuclear power plants requires the ability to distinguish between tectonic and nontectonic faulting. Nontectonic faults can produce ground deformation but are not capable of producing significant earthquakes. These faults often have characteristics similar to those of tectonic faults, but they differ in terms of their causative forces and potential hazard. Tectonic faults, which may or may not be seismogenic, include primary structures capable of producing earthquakes and secondary structures that are produced by earthquakes but are not themselves capable of generating an earthquake. An understanding of the geologic, geomorphic, and tectonic processes that result in surface deformation is essential for developing criteria to identify and evaluate the seismogenic potential of faults. In this report, we (1) summarize the characteristics of faults resulting from tectonic and nontectonic mechanisms and (2) develop criteria to identify and differentiate tectonic and nontectonic faults. We find very few diagnostic criteria to differentiate tectonic from nontectonic faults. Determining the geologic context of a fault provides the best method for differentiating the origin of a fault. Observations and measurements of scale, geometry, and timing of fault movement are the most important attributes to understand in order to confidently assess the origin and, thus, potential hazard of a fault.

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