Continental Tectonics

Geophysics Study Committee, Geophysics Research Board, Assembly of Mathematical and Physical Sciences, National Research Council

ISBN: 0-309-54048-8, 256 pages, 8.5 x 11, (1980)

STUDIES IN GEOPHYSICS

Continental Tectonics

Geophysics Study Committee
Geophysics Research Board
Assembly of Mathematical and Physical Sciences
National Research Council

NATIONAL ACADEMY OF SCIENCES
Washington, D.C. 1980

NOTICE: The project that is the subject of this report was approved by the Governing Board of the National Research Council, whose members are drawn from the Councils of the National Academy of Sciences, the National Academy of Engineering, and the Institute of Medicine. The members of the Committee responsible for this report were chosen for their special competences and with regard for appropriate balance.

This report has been reviewed by a group other than the authors according to procedures approved by a Report Review Committee consisting of members of the National Academy of Sciences, the National Academy of Engineering, and the Institute of Medicine.

The Geophysics Study Committee is pleased to acknowledge the support of the National Science Foundation, the U.S. Geological Survey, the Department of Energy, the National Oceanic and Atmospheric Administration, the Defense Advanced Research Projects Agency, and the National Aeronautics and Space Administration for the conduct of this study.

Library of Congress Cataloging in Publication Data

National Research Council. Geophysics Study Committee. Continental tectonics.

(Studies in geophysics)

Based on papers presented at the American Geophysical Union meeting in Miami in Apr. 1978.

Includes bibliographies.

1. Plate tectonics. 2. Continents. I. Title. II. Series.

QE511.4.N37 1980 551.′36 79-26942

ISBN 0-309-02928-7

Available from Office of Publications National Academy of Sciences 2101 Constitution Avenue, N.W. Washington, D.C. 20418

Printed in the United States of America

Geophysics Research Board

ROSEMARY VIDALE, Los Alamos Scientific Laboratory
A. LINCOLN WASHBURN, University of Washington
CHARLES A. WHITTEN, Silver Spring, Maryland
WARREN S. WOOSTER, University of Washington

Ex Officio

D. ALLAN BROMLEY, Yale University
C. GORDON LITTLE, National Oceanic and Atmospheric Administration
ARTHUR E. MAXWELL, Woods Hole Oceanographic Institution
ELSKE V. P. SMITH, University of Maryland

Assembly of Mathematical and Physical Sciences Liaison Representatives

PRESTON CLOUD, U.S. Geological Survey; University of California, Santa Barbara
ROBERT G. FLEAGLE, University of Washington
CARL H. SAVIT, Western Geophysical Company of America

Geophysics Study Committee

CHARLES L. DRAKE, Dartmouth College, *Chairman*
LOUIS J. BATTAN, University of Arizona, *Vice Chairman*
JOHN D. BREDEHOEFT, U.S. Geological Survey
ALLAN V. COX, Stanford University
FRANCIS S. JOHNSON, University of Texas at Dallas
HUGH ODISHAW, University of Arizona
CHARLES B. OFFICER, Dartmouth College
TJEERD H. VAN ANDEL, Stanford University

NRC Staff

THOMAS M. USSELMAN
PEMBROKE J. HART
DONALD C. SHAPERO

Liaison Representatives

JAMES R. BALSLEY, U.S. Geological Survey
EUGENE W. BIERLY, National Science Foundation
EDWARD A. FLINN, National Aeronautics and Space Administration

Panel on Continental Tectonics

B. CLARK BURCHFIEL, Massachusetts Institute of Technology, *Co-chairman*
JACK E. OLIVER, Cornell University, *Co-chairman*
LEON T. SILVER, California Institute of Technology, *Co-chairman*
LAWRENCE W. BRAILE, Purdue University
KEVIN C. BURKE, State University of New York, Albany
GREGORY, A. DAVIS, University of Southern California
GORDON P. EATON, U.S. Geological Survey
WARREN HAMILTON, U.S. Geological Survey
GILBERT N. HANSON, State University of New York, Stony Brook
WILLIAM, J. HINZE, Purdue University
ROBERT W. KAY, Cornell University
SUZANNE MAHLBURG KAY, Cornell University
G. RANDY KELLER, University of Texas, El Paso
EDWARD G. LIDLAK, University of Pittsburgh
PETER W. LIPMAN, U.S. Geological Survey
WILLIAM R. MUEHLBERGER, University of Texas, Austin
MARTIN A. SCHUEPBACH, Exxon Production Research Company
PETER R. VAIL, Exxon Production Research Company
ISIDORE ZIETZ, U.S. Geological Survey

With a contribution from

T. K. HUANG, Chinese Academy of Geological Sciences

Studies in Geophysics*

ENERGY AND CLIMATE Roger R. Revelle, *panel chairman*, July 1977, 158 pp.

CLIMATE, CLIMATIC CHANGE, AND WATER SUPPLY James R. Wallis, *panel chairman*, August 1977, 132 pp.

ESTUARIES, GEOPHYSICS, AND THE ENVIRONMENT Charles B. Officer, *panel chairman*, August 1977, 127 pp.

THE UPPER ATMOSPHERE AND MAGNETOSPHERE Francis S. Johnson, *panel chairman*, October 1977, 169 pp.

GEOPHYSICAL PREDICTIONS Helmut E. Landsberg, *panel chairman*, May 1978, 215 pp.

IMPACT OF TECHNOLOGY ON GEOPHYSICS Homer E. Newell, *panel chairman,* September 1979, 121 pp.

CONTINENTAL TECTONICS B. Clark Burchfiel, Jack E. Oliver, and Leon T. Silver, *panel co-chairmen*, February 1980, 197 pp.

* Published to date.

PREFACE

In 1974 the Geophysics Research Board completed a plan, subsequently approved by the Committee on Science and Public Policy of the National Academy of Sciences, for a series of studies to be carried out on various subjects related to geophysics. The Geophysics Study Committee was established to provide guidance in the conduct of the studies.

One purpose of the studies is to provide assessments from the scientific community to aid policymakers in decisions on societal problems that involve geophysics. An important part of such an assessment is an evaluation of the adequacy of present geophysical knowledge and the appropriateness of present research programs to provide information required for those decisions.

This study on *Continental Tectonics* was motivated by a combination of scientific and societal problems. Among societal problems related to the solid earth that have taken on urgency in recent years are those involving natural hazards—including questions of siting of dams, power plants, and other facilities and the isolation of toxic and radioactive wastes. The thesis of this report is that more reliable scientific input to the decisions concerning these problems can be provided only with a much improved basic understanding of continental tectonics. This basic understanding has been stressed in a broader context by the U.S. Geodynamics Committee, which is placing greater emphasis on the continents in its program planning for geodynamics in the 1980's.

The study was developed through meetings of the panel and presentation of papers in preliminary form at the American Geophysical Union meeting in Miami in April 1978. They provide examples of our current basic geophysical knowledge of the architecture and processes on the continents. They also pose many of the fundamental questions and uncertainties that require additional research. The essays allude to several practical applications for which an improved understanding of the continents is needed. In completing their papers, the authors had the benefit of discussion at this symposium as well as the comments of several scientific referees. Responsibility for the individual essays rests with the corresponding authors.

The overview of the study summarizes the highlights of the essays and formulates conclusions and recommendations. In preparing it, the panel chairmen had the benefit of meetings and discussion that took place at the symposium and the comments of the panel of authors and selected referees. Responsibility for its content rests with the Geophysics Study Committee and the chairmen of the panel.

CONTENTS

Continental Tectonics

OVERVIEW AND RECOMMENDATIONS

Erosional processes driven by the oceans and atmosphere would eventually reduce the continents to sea level if the earth's surface were static. But the earth's surface is not static; tectonic processes, driven by the earth's internal heat, have continually produced and elevated the continental crust. These processes are manifested by earthquakes, volcanoes, uplifts, and depressions.

The diverse and intricately distributed rocks of the continents yield most of our essential mineral resources. Knowledge of how, when, and why mineral and energy resources are concentrated in the continental crust should enhance the efficiency of exploration for and management of such resources. The complicated architecture of continental crust is the product of dynamic processes that may constitute geological hazards. Earthquakes and volcanic eruptions are obvious hazards. More subtle hazards (e.g., contamination of the environment) may have long time constants (up to many thousands of years); they represent prime con-terns in selecting stable sites for nuclear power plants and other large structures, in the disposal of radioactive waste, or in ensuring the integrity. of supplies of underground water. As we increase our demands on the earth and seek to increase our ability to predict or mitigate the instabilities. we must improve our knowledge of the basic tectonic framework and processes involved in the formation and modification of continental crust.

Tectonics is a branch of the earth sciences dealing with the origin, evolution.

structure, and internal relations of regional features of the earth's crust. Tectonics is closely related to geodynamics, which is that branch of the earth sciences that deals with the forces and processes of the earth's interior. Only during the past few decades has it been possible to study geodynamic forces and processes as a global system.

The revolution in the earth sciences that began in the 1960's has provided a major new concept—plate tectonics—that has been successful in explaining dynamic relations among major features of the earth. This concept, developed largely from data gathered from oceanic areas during the past 20 years, has had greater success in explaining the tectonics of the oceanic crust than of the continental crust. This comparative success was facilitated by the relative youth of the oceanic crust—approximately 200 million years (m.y.). The near-surface record—albeit incomplete—of the first 95 percent of earth history (approximately 4000 m.y.) resides solely in the continents.

Studies of continental geology suggest that much of the continental crust was formed by complex transformation of oceanic crust into continental crust. Plate tectonics offers a framework within which many of the processes of more recent continental evolution can be understood; but as we deal with increasingly ancient portions of the continents, it is increasingly difficult to test the applicability of the plate-tectonic concepts. Continental tectonics, however, must be explainable as part of a global geodynamic system, whose component processes may have changed through geological time.

"Scum of the earth" is used as a derogatory term, but it effectively describes the continents. They are like rafts in the global dynamic system, built of low-density materials physically and chemically extracted from the earth's interior during its long history, and floating on a higher-density interior. Continents have accumulated by the aggregation of fragments of earlier continental crust or by the additions of transformed oceanic crust, have refragmented, and have drifted apart once again. They are exposed to the atmosphere, weathered and eroded to form soils and sediments, and uplifted or depressed. Persistence of continents is long, and the processes acting on them are complex. Although the oceanic areas lend themselves well to a rather simple thermal model, a comparable model has not yet been established for the continents.

While we have appreciable knowledge about the surface of continents, we know little about their nature at depth—particularly the crystalline basement rocks. Major efforts and innovative techniques will be required to explore the continents and their margins in three dimensions. Because evidence indicates that differences between continents and oceans may extend to depths of hundreds of kilometers, we cannot limit our exploration to the crust alone. The resulting fundamental knowledge from the scientific exploration of the continents may well result in a comprehensive synthesis that will rival and complement the concept of plate tectonics. Although the challenge is scientific, we can anticipate practical consequences because the improved understanding should enlarge and strengthen the framework in which we view resources, resource exploration, waste storage, and geological hazard prediction.

The time scale for policy decisions that depend on a geophysical, geological, and geochemical knowledge base is far shorter than the time scale for developing that knowledge base. A decision may be required in a few weeks, while the information required to make it intelligently might require a decade to produce. The resultant cost of pursuing an objective from a base of ignorance is likely to be far greater than the cost of proceeding from a solid knowledge base. We have entered a period of an increasing number of important policy decisions that have a significant dependency on understanding continental tectonics. It is important that we proceed to develop this knowledge base of continental tectonics promptly and, to the extent possible, well in advance of the need to use it in policy decisions.

Four decades ago our societal demands on the earth were smaller than they are now. Two decades ago we did not have the framework to consider the proper questions about the continents. A decade ago we could ask the questions but many of the tools were unavailable. *There is now a convergence of societal need for an understanding of continental structure and processes, and a developing scientific theory and available technology that will permit us to begin to understand the basic framework and processes involved with the structure and origin of the continents.*

THE PLATE-TECTONIC FRAMEWORK

The general concepts of plate tectonics are simple. The rigid outer shell of the earth is broken into a small number of large plates moving relative to each other. Oceanic plates are generated at and move away from ocean ridges where the plates are separating. The plates plunge into the mantle at subduction zones where plates converge. Along transform faults such as the San Andreas Fault system in California, relative horizontal motion occurs between adjacent plates. These three types of plate motion (separation, convergence, and lateral sliding) are marked by structural deformation and seismicity that trace a network of activity on the earth's surface. Most of the earthquakes, volcanoes, and deformation are concentrated along this network of plate boundaries. The outer part of the earth behaves in a more rigid manner than does the interior. Nevertheless, the outer portion is subjected to various forces that can result in nonrigid deformation and motion. The rigid behavior extends from the surface to depths ranging from 5-100 km beneath oceans and perhaps 20-250 km beneath continents. The rigid portion of the earth, the lithosphere, which includes the crust and the upper portion of the upper mantle, overlies a more ductile portion of the mantle, the asthenosphere, capable of adjusting for the effects of varying surface loads. The lithospheric plates ride as passengers on the convecting asthenosphere.

A large variety of contemporaneous activity is associated with plate boundaries, such as volcanism, deformation, seismicity, high heat flow, sedimentation, and ore deposition, Although plate motions have not yet been routinely measured, geodetic techniques may soon permit such measurements. However, the results of plate motions over long time periods have led to estimates of average rates and to workable models of mountain building, birth and destruction of oceans, distribution of faunal provinces, paleoclimate, evolution of the crust and mantle, and a great variety of other phenomena that reach into every field of earth and related sciences. The understanding of plate tectonics is dependent on a multidisciplinary approach to develop broad concepts further, to establish their limitations, and to identify geological phenomena that require supplementary or complementary models.

The development of plate tectonics is mainly a result of the study of oceanic areas. The generation of oceanic lithosphere is a rapid geological process; we know of no crust flooring the modem oceans older than about 200 million years. This relative youth together with a limited number of geological processes results in a system that lends itself well to simple physical modeling. The continents do not appear amenable to simple modeling, but there are clear correlations with the plate-tectonic model. For example, modern plate-boundary activity develops a characteristic suite of geological features, and similar features can be recognized in the earlier geological record. Thus, from a scientific point of view, the existence of a geological framework makes a major effort to study the continents timely. *The revolution in the earth sciences that resulted from the study of the ocean crust should spread to the continents.*

THE CONTINENTS AND PLATE TECTONICS

The concept of plate tectonics has had a unifying effect on the earth sciences. Field geologists, laboratory geochemists, structural geologists, geophysicists, geodesists, stratigraphers, and other specialists can view their ideas and results in a common framework. In joining together they have produced results of greater significance than could the simple sum of their individual efforts.

The papers in this book represent a sampling of the kind of information that has been accumulated to date about the continental crust, the tools that are available to improve our knowledge, and some of the challenges for the near future. These papers not only illustrate how plate tectonics can be applied to the study of the continents but also identify the uncertainties in the application of the plate-tectonic model to older parts of the continents and in deducing the nature of the continents at depth. The papers represent a broad range of disciplines. Advances come from many specialties; excellent communication among the disciplines must continue.

Continents clearly are dynamic environments and are the accumulated products of long-term dynamic processes. Studies of comparative planetology, made possible through recent space programs, are teaching us that planetary evolution does not involve simple, constant, and predictable processes. The processes may be expected to vary with planetary character and to change with time; only the physical laws are invariant. Because the continents contain most of the geological record, it is to them that we must turn to determine the variations in geological processes through time.

Reliable scientific answers to societal problems related to the earth will depend on an improved understanding of continental tectonics. The study of continents should proceed from an understanding of modem plate-boundary processes to efforts to apply this understanding to the geological record of continents. By such a procedure, we can determine the extent to which modem plate-tectonic processes can explain the development of continents, where in space and when in time tectonic processes in an evolving earth were different from modem analogs, and which features of continents cannot readily be related to plate-boundary processes.

Geological and geochemical examinations of rocks that form the underpinnings of continents lead us to the conclusion that many of these originated n oceanic settings through plate-tectonic processes or earlier variants of ?? Rocks continue to be incorporated into continents by such processes as isl?? or continental collisions and by accretion of oceanic and volcanic rocks at nocollisional convergent boundaries. Superposition of new plate-boundary systems often reworks newly formed parts as well as older parts of continents to produce the complex, multiply deformed, and faulted rocks that make up the bulk of present continental crust. After continental crust has been developed and becomes part of a continental interior, it is still subject to important intraplate activity, such as basin and arch development, and modification of the lower crust and lithosphere by intraplate igneous activity. These processes may not be simply related to the current plate-tectonics model. The visible geological record demonstrates that no continental interior is safe against catastrophic rifting and a renewed cycle of continental drift.

The task of understanding the continents requires the determination of evolutionary patterns of the dynamic systems that have led to the formation and modification of the continents. This means (1) understanding plate-boundary systems as we know them today, so that they can be related to contemporaneous geological features and events; (2) projecting this understanding as far as is feasible into the past; (3) determining the nature and causes of systems that have no modern analog; and (4) understanding intracontinental processes that are not explained by plate tectonics in its present form.

The dynamic processes that lead to the formation and modification of continental lithosphere involve additions to continents from oceanic and mantle sources, recycling of oceanic and perhaps some continental lithosphere, reworking and remobilization of continental lithosphere by dynamic systems, and the vertical transfer of material from the deeper mantle to the lithosphere or vice versa. All these processes lead to lateral and vertical inhomogeneities within the outer shell of the earth.

The character of the inhomogeneities is mostly beyond our direct observation. We see only the surficial expression of processes and their consequences. No matter how "deeply" we are permitted to see into continental crust as a result of uplift and erosion, there is always more continental crust and lithosphere beneath us. Only by using geophysical techniques and direct sampling by drilling can we determine the nature of the continents at depth. The goal will be one of learning about, and understanding, the evolution of the continents in time and space, at the surface, and at depth. No single continental terrane contains complete geological and geophysical information on plate-boundary and non-plate-boundary systems, both modern and ancient. The older the rock, the longer it has been subjected to processes that can destroy or bury it; as a result, accessible exposures of older rocks are correspondingly scarce. Thus, as we search for the older geological record, international cooperation becomes especially important because the relatively rare, accessible occurrences of ancient rocks are widely distributed.

CURRENT CAPABILITIES

Recent progress in four broad categories of technology (instrumentation, transportation, communication, and computation) has had a tremendous impact on advances within the various earth-science disciplines. The importance of technology for the earth sciences is reviewed in detail in an earlier report of the Geophysics Study Committee, *Impact of Technology on Geophysics*. This report also reviews many examples of the recent innovations in the four broad categories of technology and their application within the earth sciences. Among the most important are solid-state electronics and microminiaturization, which have profoundly influenced design of field and laboratory instruments, transportation and communication systems, and computers. All four categories of technology are crucial to the use of earth-oriented remote-sensing satellite systems, which can provide valuable geological, geophysical, geodetic, and geochemical data. Computers permit a large expansion in utilization of diverse kinds of earth-science data, in automation of instruments, and in modeling.

Many uses of technology in the earth sciences reflect adaptation from developments outside the earth sciences—especially materials and information sciences.

There are also transfers of discipline-specific technologies among the earth-science disciplines. There have been important transfers of technologies from academic and government laboratories to industry and from industry to both academia and government. The adaptation of petroleum exploration techniques to study the deep structure of the continents is one example; the transfer of stable isotope and geochronology techniques from academia to industry is another.

Future advances within the earth sciences that can contribute to a multidisciplinary study of the continents depend on modem technological capabilities. Availability of up-to-date scientific equipment is a basic requirement for the achievement of an improved understanding of the continents and a more effective application of that knowledge.

The current generation of U.S. earth scientists is not only stimulated by recent conceptual advances, but is also trained to utilize the most sophisticated techniques in exploring the parameters of space, time, temperature, pressure, and chemistry by which the continents must be characterized. There now exists in universities, government laboratories, and industry an increasing pool of diverse and able scientists capable of integrating observation and concept into an improved understanding of the continents.

CONTINENTAL TECTONICS IN PROPOSED GEOSCIENCE PROGRAMS

National and international plans and programs in earth science are calling for increased scientific attention to the continents. Almost all of them reflect increased societal concerns related to resources and natural hazards and concomitant demands on a basic understanding of the behavior of continents.

The U.S. Geodynamics Committee (USGC) has prepared a report, *Geodynamics in the 1980's*, that emphasizes crustal dynamics, particularly the dynamics of the continents as a framework for understanding resource systems and natural hazards. The USGC prepared a more specific report, *Continental Scientific Drilling Program*, that is particularly relevant to understanding continental tectonics; the report recommends a national effort that focuses primarily on obtaining greater scientific return from the large amount of ongoing drilling for specific mission purposes in federal agencies, as well as a limited amount of drilling for broader scientific purposes. The scientific topics in the drilling program are (1) basement structures in deep continental basins, (2) thermal regimes of the crust, (3) processes of mineral resource concentration, and (4) the understanding of earthquakes and faulting mechanisms.

There are several reports of other committees in the National Research Council that are related to continental tectonics, for example, *Continental Margins* (Ocean Sciences Board), which deals with a critical region for understanding many continental tectonic processes, and *Geodesy—Trends and Prospects* (Committee on Geodesy), which deals with the measurement of horizontal plate motions, vertical displacements, and gravity. Several other relevant reports of the Committee on Geodesy, Committee on Seismology, the U.S. National Committee for Rock Mechanics, and the U.S. National Committee for Geochemistry are included in the bibliography.

In response to societal demands, federal agencies are increasing their emphasis on the structure, behavior, and evolution of the continents. The U.S. Geological Survey (USGS) held a workshop and prepared the report, *Dynamics of the Continental Crust*, to help coordinate existing activities within the uses in continental tectonics and geodynamics to assist in providing assessments of the nation's mineral and energy resources and natural hazards. This workshop was largely motivated by the symposium on Continental Tectonics organized by the Geophysics Study Committee at the American Geophysical Union meeting in April 1978. The National Aeronautics and Space Administration's (NASA'S) increasing emphasis on the solid earth is indicated in various reports, especially *Application of Space Technology to Crustal Dynamics and Earthquake Research*. The National Oceanic and Atmospheric Administration (NOAA) is increasing emphasis on crustal movements and gravity. The Department of Energy (DOE) has a developing program in geosciences and a diversity of programs that yield important data regarding the continents. Various components of the Department of Defense (DOD) are actively concerned with continental structures and behavior. The National Science Foundation (NSF) has long supported a wide range of research devoted to basic and applied problems in earth science. In recent years, the NSF has indicated the need for increased attention to the continents, including the offshore continental margins. The Nuclear Regulatory, Commission is sponsoring research dealing with basic questions of earthquakes, faulting patterns, and other aspects of earth sciences related to the disposal of radioactive wastes and the siting of nuclear power plants. The broad interest of federal agencies in geodynamics led to the formation by NOAA, NASA, USGS, NSF, and DOE of an Interagency Coordinating Committee for Application of Space Technology to Geodynamics.

The challenge of understanding continental evolution has also led the International Union of Geodesy and Geophysics and the International Union of Geological Sciences to design an international, interdisciplinary program in the solid-earth sciences. This program is currently under development, but it seems clear that the central focus will be on the outer shell of the earth and that the dynamics of the continental crust will be an important element of the program.

RECOMMENDATION

To remove a major gap in man's understanding of his environment, and to provide an adequate scientific basis for geological hazard and waste-disposal evaluation and for exploration, assessment, and appropriate utilization of earth resources, a broad multidisciplinary effort based on modern technology should be directed toward the exploration and understanding of the dynamics, structure, evolution, and genesis of the continents. This effort should be a major component in programs for geodynamics in the 1980's.

The following are four main targets of this effort:

• *Three-dimensional structure of the continents*.

Our knowledge of the structure and composition of the continents as a function of depth is particularly limited. Moreover, geological mapping, the basic tool in evaluating the surficial portions of the crust, is still incomplete for both the United States and the rest of the world. A complete aeromagnetic survey of the United States is necessary to explore the shallower depths (up to 10-20 km) of the crust. Satellite magnetometers show promise of being able to map the depth to the Curie temperature. There should also be a continuation of detailed gravity surveys, reflection and refraction seismic profiling, and other geophysical investigations to explore the three-dimensional variations within the continental crust and mantle. Where feasible, these data should be corroborated with drill-hole data. Additional information on the structure and composition of the lithosphere and asthenosphere should also be sought from petrological and geo chemical studies of magmas and xenoliths. Data from all these sources will allow a greater understanding of the continents and their processes.

• *Timing of the evolution of continental crust and associated mantle.*

The times of initiation, the rates, and the durations of major tectonic processes are essential temporal parameters. They need to be known for the preparation of effective genetic models of earlier continental development for comparison with our observations and models of contemporary plate tectonics. Understanding of the changing tectonic forces in relation to the driving energy with geological time can then be achieved.

• *Nature and origin of the stress fields within the continents*.

The stress field within an area is related to its potential seismic risk and its physical state. Little is known about the stress fields near *plate margins*, where the majority of earthquakes and deformations occur. Even less is known about the intraplate stress fields and the less-frequent (although equally destructive) intraplate earthquakes. Relating observed strains to the existing stress field should result in methods and models for determining paleostress fields from paleostrain measurements. Knowledge of the nature and origin of the stress fields and their relationship to geological structures of all ages would enhance earthquake predictive capabilities.

• *Thermal processes and thermal structure of the continents and underlying mantle*.

The importance of thermal processes and thermal structure has long been recognized as fundamental in connection with the evolution of the earth's crust and upper mantle. The driving energy for plate tectonics is heat; ore deposits are in many instances produced as a consequence of thermal processes, and many are indirect results of these; maturation of petroleum and upgrading of coal are functions of temperature; geothermal areas reflect concentrations of heat at depth. Nevertheless, the thermal characteristics of the crust and upper mantle are poorly understood. Heat flow near the surface can be measured directly. All other determinations of thermal structure depend on indirect geophysical, geological, and geochemical techniques. Through use of geochemical tools it is possible to examine the thermal structure in the past in selected areas and to compare these data with present thermal conditions to gain insight into the evolution of thermal processes.

Integrated multidisciplinary studies will be necessary for effective investigations of continental tectonics. Integration can and should be achieved through cooperative programs and through proper communication. These investigations should be aimed at a basic understanding of the processes of continental formation and modification. Armed with this understanding, the earth-science community will be much better equipped to aid in the assessment and solution of current and unforeseen problems related to the solid earth.

BIBLIOGRAPHY

Orientations in Geochemistry, U.S. National Committee for Geochemistry, National Research Council. National Academy of Sciences, Washington, D.C., 122 pp., 1973.

U.S. Program for the Geodynamics Project: Scope and Objectives, U.S. Geodynamics Committee. Geophysics Research Board, National Research Council. National Academy of Sciences. Washington, D.C., 235 pp., 1973.

The Geology of Continental Margins (C. A. Burk and C. L. Drake, eds.), Springer-Verlag, New York, 1009 pp., 1974.

Geodynamics: Progress and Prospects (C. L. Drake, ed.), American Geophysical Union, Washington, D.C., 238 pp., 1975.

Predicting Earthquakes. Committee on Seismology, National Research Council. National Academy of Sciences, Washington, D.C., 62 pp., 1976.

The Earth's Crust: Its Nature and Physical Properties (J. G. Heacock, ed.). Geophysical Monograph 20, American Geophysical Union. Washington, D.C., 754 pp., 1977.

Trends and Opportunities in Seismology, Committee on Seismology, National Research Council. National Academy of Sciences, Washington. D.C., 158 pp., 1977.

Geodesy: Trends and Prospects, Committee on Geodesy, National Research Council. National Academy of Sciences, Washington. D.C., 86 pp., 1978.

Geophysical Predictions, Geophysics Study Committee, Geophysics Research Board. National Research Council. National Academy of Sciences, Washington, D.C., 215 pp., 1978.

Limitations of Rock Mechanics in Energy-Resource Recovery and Development. U.S. National Committee for Rock Mechanics, National Research Council. National Academy of Sciences. Washington, D.C., 67 pp., 1978.

Applications of a Dedicated Gravitational Satellite Mission. Committee on Geodesy, National Research Council. National Academy of Sciences, Washington, D.C., 53 pp., 1979.

Application of Space Technology to Crustal Dynamics and Earthquake Research, NASA Technical Paper 1464, National Aeronautics and Space Administration, Washington, D.C., 257 pp., 1979.

Continental Margins. Panel on Continental Margins, Ocean Sciences Board, National Research Council. National Academy of Sciences, Washington, D.C., 302 pp., 1979.

Continental Scientific Drilling Program, U.S. Geodynamics Committee, Geophysics Research Board, National Research Council. National Academy of Sciences, Washington, D.C., 192 pp., 1979.

Dynamics of the Continental Crust: Proposals for a U.S. Geological Survey Program in the 1980's. U.S. Geological Survey Open File Report, 1979.

Impact of Technology on Geophysics*. Geophysics Study Committee, Geophysics Research Board, National Research Council. National Academy of Sciences, Washington, D.C., 131 pp., 1979.

Earthquake Research: An Aid to the Safer Siting of Critical Facilities. Committee on Seismology. National Research Council. National Academy of Sciences, Washington, D.C. (in press).

Geodynamics in the 1980's, U.S. Geodynamics Committee, Geophysics Research Board, National Research Council. National Academy of Sciences. Washington, D.C. (in press).

I

SUMMARY

1

Plate Tectonics and the Continents: A Review

B. Clark Burchfiel
Massachusetts Institute of Technology

INTRODUCTION

The plate-tectonics model has had marked success in explaining the first-order features of the earth; it has been particularly successful in explaining the nature and age of the crust of the ocean basins. The model accounts for the distribution of earthquakes and volcanoes along plate margins and the nature of deformations occurring along the margins of the continents. The model, however, has had less success in accounting for geological activity within plates, especially the continental portions, and our confidence in the usefulness of the model diminishes as the age of continental crust increases.

The model has been applied in an attempt to explain the tectonic features of the continents and to test the limitations of the model in time and space. The accompanying papers in this volume illustrate this application and the limitations and methods used to study the continental crust. The characteristics of plate boundaries, along which the major deformation seems to occur, and the nature of tectonic events that are not associated with plate boundaries are summarized in this paper. Chapter 2 treats the problems of understanding the pre-Mesozoic [older than ~200 million years (m.y.)] evolution of the continents, which represents the majority of continental crust.

PLATE BOUNDARIES

The study of modem plate boundaries allows us the opportunity to examine the geological and geophysical activity at the boundary and contemporaneous activity adjacent to the boundary. Studies should be directed toward establishing relationships between plate-boundary activity and the wide range of geological and geophysical events, so that plate boundaries can be treated as dynamic systems.

We know from modem and recent plate-boundary activity that conditions at the boundary may change rapidly, on the order of a few million years, and in order to relate geological events to plate-boundary activity contemporaneity of events must be established. Changes in plate-boundary activity may lead to partial destruction, deformation, or overprinting of the geological features of an older system by a younger system. Interrelations between the various parts of each system become lost or blurred.

Such changes are the reason why the details of an ancient plate-boundary system become hard to decipher. It is only through the study of modem plate-boundary systems that we can begin to comprehend fully the geological, geophysical, and geochemical processes and their interrelations within these dynamic systems. The end product of such studies would be to understand the processes and geological results of these systems from the earth's surface into the underlying mantle. We are a long way from this level of understanding, but continued investigations into modern plate-boundary systems must continue as they form the fundamental basis for plate-tectonic analyses.

Even though there is great diversity in modern plate-boundary systems, they do not represent the complete range of potential plate-boundary systems. Therefore, geological investigation of ancient plate-boundary systems may demand interpretations different from those of modern systems.

Plate-boundary systems can be divided into divergent, transform, and convergent types. Relative motion across plate boundaries is often not orthogonal, particularly at convergent boundaries; thus more complex combinations of these types may also exist.

Divergent Boundary Systems

Divergent motion within continental lithosphere causes extension of the lithosphere, producing elongate fault-controlled depressions called rifts. If divergence continues, oceanic crust may develop and the continental lithosphere becomes part of two diverging plates. In many cases rifting ceases before oceanic crust is developed, and the rift and its related features remain within continents. Where divergent motion continues and oceanic crust is formed, the rift structures become buried and form the basement for passive continental margins. Burke (see Chapter 4) has outlined many of the features associated with continental rifts and has emphasized that to evaluate these divergent systems properly we must establish contemporaneity of events within the system, the evolution of rift systems, and the temporal structural relations of rifts to adjacent oceanic crust.

A modern example of a rift system that has led to the development of continental separation, formation of oceanic crust, and a divergent plate boundary is the East African, Red Sea, Gulf of Aden rift system. Studies in this region have produced good, yet still incomplete, understanding of the evolution of a divergent plate-boundary system. The East African rifts are in the rift stage of development, whereas the Cull of Aden has entered the drifting stage. The Red Sea appears to be in a transitional stage.

From studies of the East African-Gulf of Aden area (see Chapter 4) and from geophysical studies of continental margins (see Chapter 5), an evolution of a divergent plate-boundary system can be sketched. During the early rifting stage, the system consists of a broad linear belt of active normal faults that produce elevated and depressed parts of the crust and characteristic patterns of sedimentary rock types that commonly include great thickness of salt. This early stage is also accompanied by volcanism, both alkaline and tholeiitic, and significant geothermal activity. The lithosphere is thinned over a broad linear zone, and the asthenosphere rises beneath the rifted zone. During continued rifting, this system enters the drift stage, in which oceanic crust forms between the separated continental lithosphere. Volcanism, structures, and associated sedimentary basins may extend for hundreds of kilometers away from the site that eventually develops into a divergent plate boundary. Modification of continental lithosphere must take place beneath the entire affected region. The divergent system thus affects a broad region of continental lithosphere, whereas divergent activity is more restricted within oceanic lithosphere. The relationship between upper crustal brittle faulting and deeper lithospheric structural and chemical behavior is at present poorly known and is certainly an important area for future studies.

As the rifted continental margins drift away from the spreading ridge they become less active and finally form a passive continental margin, where the transition from continental to oceanic lithosphere takes place within a single plate. Early subsidence along the passive margin leads to the formation of continental shelf sedimentary rocks, which overlap early-stage rift deposits. Further subsidence can be explained by thermal decay away from a spreading ridge, and the formation of a constructional passive margin consisting of shelf slope and rise morphology underlain by characteristic sedimentary-rock assemblages.

The continued subsidence of some continental margins indicates that they are not truly "passive," but important dynamic processes are still active at the interface between continental and oceanic lithosphere. Sediment loading has been proposed as a mechanism for continued subsidence, but it is insufficient to provide the proper magnitude of subsidence. Bott (1971) has argued that continued subsidence along some passive margins may be the result of lateral flow in the lithosphere across the ocean-continent interface. Such ductile flow could produce its own characteristic structures and modification of a passive margin that are superposed on those of the earlier rift stage. By the time a passive margin has completed its subsidence due to thermal decay, it is outside the influence of the divergent boundary system. Its further development is in an intraplate setting, and the processes that affect it properly should be discussed under the topic of non-plate-boundary systems. But since passive-margin development is clearly part of the evolution of divergent systems it is considered here. Investigations of modem passive margins at all stages of development are still in their infancy, and a great deal is yet to be discovered about their dynamics.

Ancient passive margins are well known from the late Precambrian to the Recent. The marginal parts of most deformed belts, now present as part of the internal struc ture of continents, contain rock assemblages and structures that indicate that they developed as part of a divergent dynamic system. Similar passive-margin sedimentary sequences are known as eroded remnants from some middle Precambrian deformed belts, which suggests that deep erosion could remove all sedimentary evidence for former passive margins. Some structures, igneous activity, and lithospheric modification by a divergent dynamic system could, however, still be preserved. Some middle Precambrian deformed belts show no sedimentary record of a former passive margin, and the question of whether one was ever present may have to rely on studies of the geological and geophysical characteristics of the crust beneath passive-margin sedimentary sequences. Archean passive margins are unknown, which, along with much evidence, has suggested to many scientists that during most of Archean time continental lithosphere was thin and ductile (see Chapter 15). If plate-tectonic processes were active in the Archean, they may have operated at rates and scales on lithosphere and mantle with characteristics unlike those of later Precambrian and Phanerozoic time.

Some ancient passive margins are eroded where deeper parts of the lithosphere are exposed. In the southern Alps, an entire section of the continental crust and several kilometers of its tipper mantle are exposed. Geological evidence indicates that this area was the site of a divergent plate boundary that evolved into a passive margin during Mesozoic time. Study of ancient passive margins permits the direct examination of geological features produced at greater depths than those of modem passive margins.

During the early rifting stage, several rifts may develop that do not progress to the formation of ocean crust. During the drifting phase they become inactive but retain the geological features characteristic of rifting, such as extensive normal faulting, igneous activity, coarse clastic and evaporitic deposition, followed by local overlap by passive-margin sediments. Studies of these failed rifts are important because they yield evidence for the evolution of adjacent passive margins and oceanic terranes, which for ancient examples may be completely destroyed. Failed rifts, commonly referred to as aulacogens. are sometimes reactivated by other dynamic systems. forming either local mountain belts. subsiding basins (see Chapter 4), or the locus of modern intraplate seismicity. and faulting (see Chapter 7). Many failed rifts contain important gas and oil reserves as well as being related locally to other types of ore deposits.

Transform Boundary Systems

Transform boundaries, where two lithospheric plates slide past each other horizontally, that affect continental lithosphere often form wide and complex belts of deformation. The San Andreas Fault of California has been regarded as marking the transform plate boundary between the Pacific and North American plates. It is, however, only one of numerous northwest trending faults of similar right-slip displacement. Some of these faults have large displacements and may have served as the plate boundary prior to the development of the present San Andreas Fault. It is more likely that the relative motion between the two plates is a much broader zone, stretching from the marine California borderlands inland to at least western Nevada, a zone perhaps 500 km wide. Thus, this transform boundary, like most others within the continental crust, is a broad zone of plate interaction and may be called more properly a transform boundary system.

Although transform boundary systems contain a wide variety of associated geological features, the San Andreas system can serve as an example. Associated with the numerous strike-slip faults are areas affected by extension, such as pull-apart basins, or compression, such as the folds and thrusts within the Transverse Ranges (Crowell, 1974). Large-scale right-lateral structural bending and faulting in southwestern Nevada and adjacent California are probably part of the San Andreas transform system (Figure 1.1). Blake *et al.* (1978) have reviewed the Neogene history of coastal California and have shown that the development of the oil-rich Neogene basins is related to evolution of the San Andreas system. In addition, reconstructions of Mesozoic and early Cenozoic paleogeography are complicated by the scrambling of paleogeographic belts and associated rotations of these terranes by strike-slip faulting of late Cenozoic age (see Chapter 3) related to this system.

Travel-time data from teleseismic studies of the upper mantle in southern California by Hadley and Kanamori (1977) suggest that the San Andreas Fault does not offset a ridge like body of high-velocity mantle extending from 40 to perhaps 100 km. The San Andreas Fault apparently is confined to the upper 40 km of the continental lithosphere. The ridge of high-velocity upper mantle extends 150 km east of the San Andreas before ending. Hadley and Kanamori suggest that the upper lithosphere beneath southern California is decoupled along a surface of horizontal shear, perhaps marked by low velocity, and that the plate boundary at depth is further east than at the surface (Figure 1.2). These interpretations are significant because they raise the question of correlation between tipper lithosphere and deeper plate and mantle dynamics. The decoupling at one or more layers in the lithosphere has been suggested for other areas and other types of plate-boundary, systems (see Chapters 8 and 9; Armbruster *et al.*, 1978). This problem of lithospheric decoupling and its effects on correlating shallow and deep plate motions and dynamics is clearly an important question that demands further attention.

The San Andreas Fault and other faults of this system have been the focus of studies concerned with earthquake prediction and associated geological hazards. A wide range of studies have been directed toward understanding the dynamics of strain build-up, fault initiation. and movement. Because the San Andreas system has been active from about 30 m.y. ago to the Recent. the opportu nity exists to understand the details of this transform system through time and its imprint on continental geology.

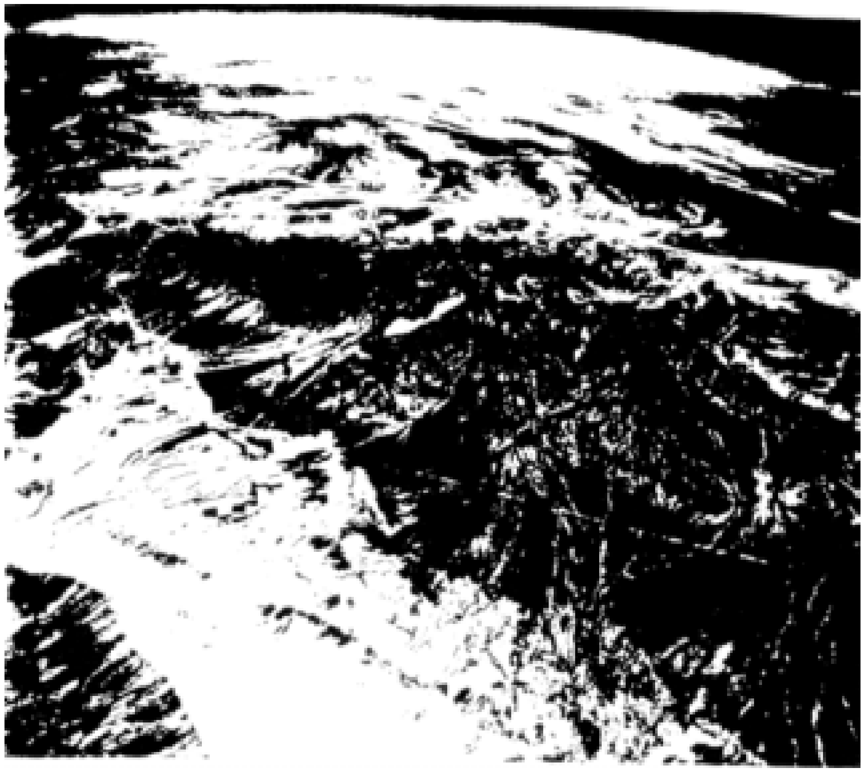


Figure 1.1
Looking northwest across southern Nevada and across the Las Vegas Valley shear zone. The shear zone is marked by right-lateral bending of all pre-Miocene rocks and marks the displacement near the eastern side of the San Andreas transform system. The bending and faulting along the shear zone account for a minimum of 70-km differential motion of late Miocene Age.

Many large faults of proven strike-slip displacement are known in the geological record, such as the Great Glen Fault in Scotland. Other large linear faults or zones of faults, such as the Kings Keweah Suture in California (Saleeby, 1977) and broad Proterozoic shear zones of the midcontinent region of the United States (Warner, 1978) have been interpreted as ancient transforms. The fact is that the importance of transform boundary systems in the evolution of continents is essentially unknown. Where ancient aseismic displacement may have occurred over a broad zone, it is possible that we might not recognize it. A deformed belt such as the Precambrian Limpopo Belt of southern Africa is an intracratonic orogenic belt resulting from horizontal shear between two undeformed crustal blocks. The deformation described by Coward *et al*. (1976) occurs across about 100 km of terrane and took place at amphibolite and locally granulite metamorphic grade. Structural evidence indicates that the deformation was mainly by horizontal crustal shear. Can this deformational style be the deeper expression of transform motion? An answer cannot be given with our present state of knowledge of transform systems within continental lithosphere.

Convergent Boundary Systems

Convergent plate boundaries, where one lithospheric plate passes beneath another, are the sites of subduction or recycling back into the mantle of the lithosphere created at spreading ridges. The stable configuration at a convergent boundary is where oceanic lithosphere is subducted because it is more dense than the underlying asthenosphere. Where oceanic lithosphere is subducted beneath oceanic lithosphere a volcanic island arc chain is present. Island arcs or continents may form part of a subducting plate and may eventually be drawn into a subduction zone. Because they are formed from material less dense than oceanic crust, they are more difficult to sub-duct. The result is arc-arc, arc-continent, or continent-continent collisions.

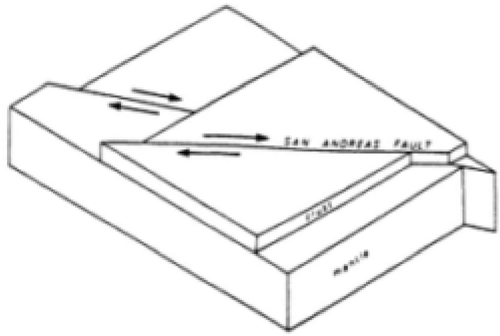


Figure 1.2 Interpretive diagram of the proposed divergence of the crust and mantle boundaries along the San Andreas Fault in southern California (from Hadley and Kanamori. 1977, reprinted from the Bulletin of the Geological Society of America, with permission).

Study of modern subduction boundaries demonstrates that geological, geophysical, and geochemical events and features may extend for several thousand kilometers from the boundary. From studies of events and features at modem convergent boundary systems, it is clear that ancient analogs exist within continents, and it appears that most material that forms continental lithosphere has been generated or reworked in convergent boundary systems. It is within such boundary systems that rocks formed within an oceanic setting are added to continents.

Noncollisional Convergent Systems

In modern noncollisional convergent systems geological features related to the system occur mainly within the overriding plate and only to a minor degree within the lower plate. In fact, those features in the lower plate that are part of the system—extension faults caused by flexing of the lithosphere seaward of the subduction zone—are mostly lost during subduction; this is not the case during collisional events where the subducted plate is not oceanic lithosphere. It is the association of geological features of the overriding plate that is commonly preserved in the geological record and leaves a record of plate convergence.

Hamilton (Chapter 3) has reviewed briefly some of the elements of convergent systems. The outermost elements of the system are the trench and outer nonvolcanic ridge formed from deformed packages of sedimentary rocks containing occasional slivers of oceanic lithosphere, an outer or forearc basin of undeformed sedimentary rocks, and volcanic arc consisting of a characteristic suite of volcanic rocks. Behind the volcanic are, three dynamic settings are possible: (I) where the overriding plate is in extension, leading to the development of a marginal sea floored by young oceanic crust or more distributed extension leading to extended continental crust such as the Basin and Range province (see Chapters 8 and 9); (2) where the overriding plate is in compression, leading to the development of a wide belt of compressional deformation behind the arc such as in the modern Andes (see Chapter 6); and (3) where the overriding plate is neither in extension nor compression and little or no back-arc structural activity takes place.

The dynamics of these modern systems are being pieced together from studies in all branches of earth science. Geophysical research (see Chapter 10) has begun to define the geometry, thickness, and behavior of the subducted slab; the state of the lithosphere and asthenosphere above the subducted slab; the nature of the thermal regime both above and within the subducted slab; and the general stress field within the overriding plate. Geochemical research has focused on problems related to the origin of the magmas and what data the volcanic are rocks can contribute to our understanding about the nature of the source material (see Chapter 13), what modifications take place during assent of the magmas. and their final emplacement (see Chapter 12). Geologists have begun to establish the paleogeographic environments of the sedimentary and volcanic sequences, the deformational and metamorphic history of the different parts of the system, and the range in evolutionary patterns for convergent systems. This is by no means a complete list of important research conducted in modem convergent systems but only a sampling to exemplify the multidisciplinary approach that must be undertaken to understand the dynamics of such complex systems.

Within continents, ancient noncollisional systems can be identified as far back as the early Proterozoic or the Archean, although the earliest examples are significantly different from modern systems, and some workers have suggested that they have formed from dynamic systems that should be considered unrelated to plate tectonics. Mesozoic examples for each type of system are common and well established, Paleozoic examples are also common but many are less well established, and Proterozoic examples are present but with a few exceptions are poorly understood. The problem becomes even more acute in the Archean where some elements of plate-boundary. systems can be recognized. but the assemblage of geological features present in modern systems is lacking. Whether these Precambrian belts were formed by plate-boundary systems as we know them today, by modifications of modern systems, or by systems unique to the Precambrian is still an important question (see Chapters 2 and 15) and one that certainly needs considerable attention because much of the continental lithosphere consists of Precambrian belts of activity.

Noncollisional convergent systems can affect continental growth and evolution sometimes for more than 1000 km from the plate boundary. The Mesozoic and

Cenozoic development of the southern part of the western United States can serve as an example. Subduction of oceanic lithosphere beneath the continental margin of the western United States began about 240 m.y. ago and has continued, at least locally, to the present. During much of Mesozoic time it appears that the convergent boundary system generated compression within the overriding plate and produced structural and magmatic effects more than 1000 km eastward from the surface trace of the plate boundary (see Chapters 3 and 6). The sequence of terranes from west to east across the western United States consists of a western terrane composed of accreted oceanic sedimentary and volcanic rocks, a central terrane composed of an Andean-type magmatic arc, an eastern terrane composed of belts of folds and east-directed thrusts where older continental basement was reactivated, and locally a foreland belt of faulted Precambrian crystalline rocks formed by older plate-boundary systems. The relationship between all these terranes is poorly understood, but correlation with modern systems such as the Andes will allow a better understanding of the dynamics of such systems. Study of the Cordilleran belt will permit us to study relations between these terranes at crustal levels not exposed in the modern Andes. These two types of studies need to proceed together

During the later part of the early Cenozoic time, convergence continued and magmatic effects took place across the western United States, but only minor deformation occurred, with the exception of the Rio Grande Riff that began about 29 m.y. ago (see Chapter 14). From about 20 m.y. ago to the present, extensional tectonics have disrupted the continental crust, forming the Basin and Range province. Extension began in a back-arc setting and evolved complexly so that today part of the extensional terrane is being overprinted by the San Andreas transform system (see Chapters 8 and 9).

This late Cenozoic development of the Basin and Range extension has been recognized by several generations of earth scientists, and many explanations for the extension involve dynamics that would more properly be discussed under the category of divergent or transform plate-boundary systems. Recent studies have indicated that the extension is an integral part of a convergent system. Recent studies in the Basin and Range province exemplify how new and exciting discoveries await us in unexplored territory and how the melding of data from all earth-science disciplines can yield completely new insights into the development and evolution of the area. Field studies about 10 years ago led to the discovery that many subhorizontal faults, long considered Mesozoic thrust faults, were Cenozoic in age. A few years later it was determined that some areas affected by high-grade metamorphism and deformation, again considered Mesozoic in age, were Cenozoic in age and locally could be related to the subhorizontal low-angle faults. Tying together all the available geophysical, geological, and geochemical data known for the Basin and Range province, Davis (see Chapter 8) and Eaton (see Chapter 9) have developed models suggesting that the crustal rocks above 15 km extend by lystic normal faulting and are detached along a variably thick transition zone from the underlying crust that is extended by ductile flow and dike intrusion. Erosional levels within the Basin and Range province have exposed perhaps all three crustal levels. The ideas expressed in these models are new and unproved but demonstrate that the potential still exists for major new discoveries in regions that have been studied for several generations.

Pre-Mesozoic examples of noncollisional plate-boundary systems are poorly known largely because the ultimate end of continental-margin evolution is collision by either island arcs or other continental fragments. These events commonly overprint the terranes generated prior to collision. Because a Paleozoic ocean basin was present in the site of the modem Pacific, evidence for Paleozoic noncollisional convergent systems is present around the Pacific margin, such as in Australia, South China (see Chapter 16), Canada, and parts of South America and Antarctica. Noncollisional systems are probably present within some Precambrian deformed belts, but because collisional events have overprinted these systems, they have become difficult to recognize. Only through careful unraveling of the overprinting events and recognition of the characteristic features of the noncollisional systems can these earlier systems be reconstructed.

Collisional Convergent Boundary Systems

Island-arc terranes are formed by noncollisional convergent boundary systems, but, because oceanic crust ultimately is subducted, the end product of all island-arc development is collisional with a continent. The dynamics of island-arc systems can be studied from modem examples, but the features of these systems become disrupted during their inevitable collision. During arc-continent collision, new crust is added to the continent, but in many cases collided arcs are reworked by later superposed plate-boundary systems before they become a part of typical continental lithosphere. Continent-continent collisions simply rearrange continental material, but the processes leading to collision may add some new continental material.

Several *arc-continent collision systems* are at present active and in different stages of evolution. In most arc-continent collisions, the features of noncollision systems are present in the island arc, but where continental lithosphere is present in the subducting plate and enters the subduction zone, considerable deformation may take place within the subducting plate, extending the effects of the convergent boundary system. The following modem convergent systems show an increase in the extent of disruption of the subducted continental lithosphere: (1) northwest Australia, (2) Taiwan, and (3) northern New Guinea. In each case, the continental passive-margin sedimentary rocks, and in the latter two cases even the underlying continental crust, have been cut by thrust faults locally as much as 200-300 km from the site of the collision boundary. Because island arcs are backed up by oceanic lithosphere, following collision with a continent, the new ocean-continent interface is potentially subject to the development of a new plate-boundary system. If the geometry of the new convergent system subducts oceanic lithosphere beneath the collided arc, rocks of the collided system are overprinted by a noncollisional continental-margin arc system and become deformed and metamor-phosed to form the polyphase terranes common within continental crust. The transitions from arc collisions are not easy to interpret, and some of the volcanic rock associations developed following collision in New Guinea are difficult to relate to plate-tectonic processes (Johnson *et al*., 1978).

Ancient mountain belts abound with evidence of arc-continent collisions and overprinting by younger plate-boundary systems (see Chapters 3, 6, and 16). Because of the common overprinting of collided arcs, the reconstruction of the plate-boundary systems related to the arc, its collisional history, and overprinting episodes are complex indeed. Study of older and more deeply eroded collided are terranes gives us the opportunity to examine the details of the structure, geochemistry, metamorphism, distribution of rock types, and evolution of arcs and collisional margins that lie at depth below modem systems where the details remain obscure. Studies of these arc terranes will clearly extend back into Archean rocks of continents where are systems have been identified. There are, however, significant differences between modem collisional systems and Archean ones (see Chapters 2 and 15). These differences must be understood, because they relate to the evolution of the earth, its lithosphere, and its continental regions.

Continent-continent collisions occur where oceanic lithosphere between continents is completely subducted in one or more subduction zones. Remnants of the oceanic lithosphere with or without associated island-arc terranes may be preserved within the collisional boundary.

Active systems of continent-continent collision are present in the Alpine-Himalayan mountain belt. Studies of this modem collisional system have shown that once collision has taken place, but convergence between the two plates continues, geological activity within the system broadens throughout both plates and becomes very diffuse. In fact, the geological activity may become developed in so many local structures that it is difficult, if not impossible, to define a single plate boundary.

Nowhere is this type of activity better exemplified than in Asia. Molnar and Tapponnier (1975) have traced the collisional history between India and Asia from collision about 50 m.y. ago to the present. Since collision, India and Asia have converged by more than 2000 km. This postcollisional convergence has caused deformation within the Asian plate for a distance of 3000 km from the collisional boundary (Figure 1.3), In addition, the Indian plate was disrupted by south-directed thrust faults. Molnar and Tapponnier have suggested that the faults within the former Asian plate form a pattern that is explainable by indentation mechanics (Figure 1.4). India is displacing large parts of Asia laterally as it continues northward. Many of the active faults follow structural lines formed by older collisional events within Asia (see Chapter 16). Thus, it appears that such features as the extensional opening of the Baikal Rift are part of a convergent system whose initial collisional boundary was 3000 km away. Within this plate-boundary system complex patterns of volcanism, crustal deformation, and lithospheric modification are taking place. Once such a system becomes inactive, it is obvious that detailed geological, geophysical, and geochemical studies will be necessary to establish contemporaneous relations between widely diverse geological features and events.

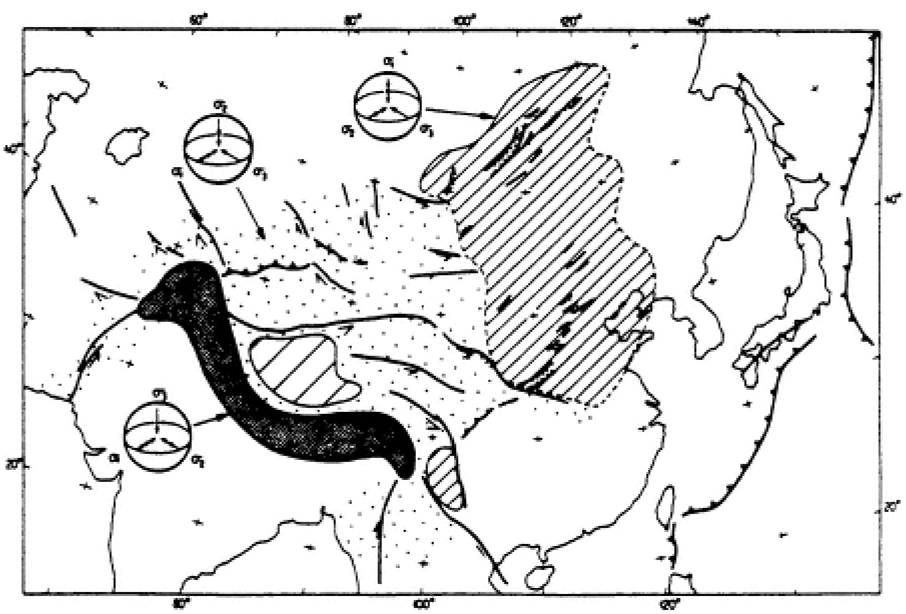


Figure 1.3
Distribution o f tectonic styles in Asia along the Indian-Asian collisional system (from Tapponnier and Molnar, 1976). Dark area, region of thrusting and crustal thickening; dotted area, regions of strike-slip faulting; lined area, region of normal faulting and crustal thinning. Corresponding stress states are also indicated.

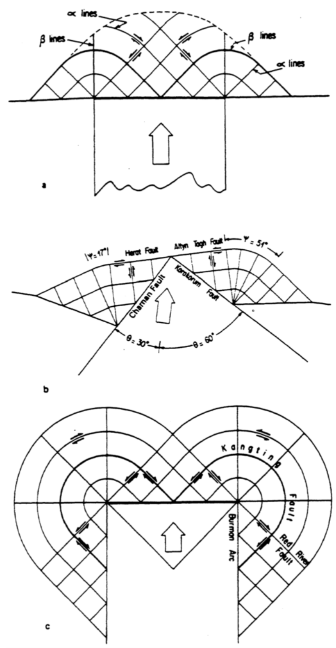


Figure 1.4
Plane indentation of semi-infinite rigid-plastic media by different rigid dies (from Tapponnier and Molnar, 1976). Arrows indicate sense of shear along slip lines. Principal stresses σ_1, and σ_3 bisect small quadrilaterals delineated by slip lines. The names of corresponding major tectonic features are indicated. (a) Flat rigid die; (b) rigid wedge, similar to the situation that occurs at the Pamirs (western end of the Himalayas); (c) flat triangular indenter and hallowed-out medium, similar to situation that arises at the Himalaya-Burman syntaxis (eastern end of the Himalayas).

Ancient examples of continent-continent collisions are common in the geological record. Huang (Chapter 16) has described the tectonic evolution of China in such a way that it is obvious that China has been built by a complex series of arc and continent collision events that began in the late Precambrian, continued through the Paleozoic and Mesozoic, to the most recent collision of India during the Cenozoic. The ancient sutures can be identified in many places, but the far-reaching effects of convergent systems can produce a complex pattern of extensional, translational, and compressional features, which is only now being recognized. Burke (Chapter 4) has suggested that the Keweenawan rift system, which forms a major crustal structure of the central United States (see Chapter 11), may be related to a collisional system within the Grenville (1000-m.y.-old) deformed belt (see Chapter 15).

Compound convergent systems may contain characteristics of two types. Such plate boundaries are characterized by relative motions that are oblique or change rapidly because of continual shifts of rotational poles. Along the western part of the Java subduction system, the relative motion is oblique to the boundary and under-thrusting takes place in the trench, whereas strike-slip faulting takes place in the overriding plate and is super-posed on the volcanic arc terrane. Further north, along the same plate boundary, the motion becomes strike-slip, and extension occurs behind the boundary in the Andaman Sea.

Similar types of compound boundaries can be recognized in ancient systems within continents. Within the Cordilleran belt of Canada and Alaska, both collisional and noncollisional convergence during late Mesozoic time was associated with major strike-slip faults that were formed within the continent several hundred kilometers from the plate boundary (Davis *et al*., 1978). The evolution of the Pyrenees has been interpreted as a combination of transform and both extensional and convergent systems at different times (Mattauer and Seguret, 1971).

Complex combinations of boundary systems can also result from collisions that take place along initially irregular boundaries. Dewey and Burke (1974) hypothesized that continent-continent collisions along irregular boundaries could result in fragmentation and complex relative motions of small fragments. McKenzie (1972), from studies of active seismicity in the eastern Mediterranean region, showed that during such fragmentation complex relative motions take place. It can be argued that the modem seismicity gives us only an instantaneous picture of the boundaries and relative motions of small fragments, and, given longer periods of geological time, the entire belt of deformation could be regarded as a continuous or at best a semicontinuous system of deformation. Analysis of the motions of small fragments suggests that their motions may not be a reflection of motion at depth. Within the alpine belt, analysis of fragment motion suggests that the continental fragments are probably decoupled at levels within the crust and/or within the lithosphere and their motion is not directly reflected by deep lithospheric or asthenospheric flow (Burchfiel, in press). Such studies as those of Hadley and Kanamori (1977) along the San Andreas Fault and Davis (Chapter 8) and Eaton (Chapter 9) for the Basin and Range province add evidence that such detachments are possible. Some geophysical evidence suggests that for these zones decoupling may be related to low-velocity. crustal layers. All these data and interpretations suggest that lower lithosphere now found beneath such continental fragments is exotic and may have an evolution different from the crust that overlies it. These types of complexities are only now being recognized, and the extent of such activity must be understood because it could modify greatly the models we develop for the evolution of continental crust.

NON-PLATE-BOUNDARY SYSTEMS

There are many structural and magmatic features within continents that have no apparent relation to plate-boundary activity. Some of the features regarded above as related to plate-boundary systems are so far removed from the boundary that they may be independent of such systems and should be included in the category of non-plate-boundary systems. Included in non-plate-boundary systems, but not discussed in this chapter, are features of the interiors of oceanic plates.

Intraplate Magmatic Activity

Intracontinental magmatic activity that is not related to plate-boundary systems takes the form of small plutons, plugs, laccoliths, dikes, diatremes, kimberlites, and some volcanic rocks. They have a complete range of compositions from ultramafic to alkalic. Sources of material for these rocks are highly varied. Some basaltic igneous rocks and radiogenic isotopic data suggest that they are developed from a deep mantle source, possibly below the asthenosphere, that was homogenized about 2.0 b.y. ago (see Chapter 13). Other basaltic rocks such as those of the Absaroka volcanic rocks, Snake River plain, and at Yellowstone, are derived from subcontinental mantle that has been attached to the overlying continental crust since the last major thermal event to affect these rocks (see Chapter 14). The data indicate the continent and upper mantle in this area have been together since 2.8 b.y. ago.

All these types of igneous activity require perturbations in the mantle to cause local melting. What these perturbations may be is uncertain and clearly represents problems for considerable future study. Whatever the causes, these rocks record events that contribute to the evolution of already formed continental lithosphere. Many of these igneous rocks contain fragments, xenoliths (see Chapter 12), that give us a sampling of the rock types present within the lithosphere and asthenosphere. If their source can be properly placed, and their age determined, they yield additional valuable evidence concerning the evolution of continental lithosphere. These xenoliths, however, must be interpreted cautiously as they come from perturbed areas within the earth and may not be representative of the mantle or lower crust (Irving. 1976).

Intraplate Vertical Movements

The development of basins and arches within continental areas appears to be unrelated to plate-boundary activity. Basins contain sedimentary accumulations that record the history of subsidence. The arches, on the other hand, are uplifted areas, and much of the record has been removed by erosion, and the history of their movement is more difficult to decipher. Vertical movements of the crust are the dominant processes in the formation of basins and arches, but the nature and cause of these movements is virtually unknown. Intracratonic basins are superposed on already formed continental crust of varied age and structure. The Michigan Basin overlies an older late Precambrian rift terrane, whereas the Illinois Basin overlies an older Proterozoic folded belt that was affected by later Precambrian rhyolitic magmatism. Some basins are accompanied by early igneous activity and normal faulting, whereas others are not. The variety of basin settings has made it difficult to arrive at an explanation for their formation. A number of models have been proposed that rely on thermal, eustatic, tensional, and compressional mechanisms (see Chapter 7). None of these, however, has offered a completely satisfactory explanation. Basins are an important feature of continental tectonics because many contain important accumulations of hydrocarbons.

Basin and arch development within continents indicates late-stage modification of already formed continental lithosphere. What types of modifications take place at depth are unknown but must be understood before a complete evolution of the lithosphere can be deciphered.

Recent studies of unconformities and the fluctuation of sea level within basins has suggested a similarity between basins within a single continent and on several continents (Sloss, 1972). Studies on continental margins show similar histories, indicating that these features may be a worldwide phenomenon. If this proves to be the case, at least part of the sedimentological controls could be related to plate-tectonic activity.

Broad epirogenic movements, as with their more restricted counterparts, basins and arches, may also lead to modification of the continental lithosphere. Broad vertical movements, such as the elevation of the Colorado Plateau and adjacent Rocky Mountain area in the Cenozoic, must involve changes within the mantle. What these changes are and how they affect the continental crust and lithosphere are essentially unknown.

Intraplate Deformation

Even though continental lithosphere away from active plate-boundary systems is considered rigid, it is not en tirely inactive. Recent studies have shown diffuse seismic activity to be present throughout the eastern part of the United States (see Chapter 7). Some of these earthquakes have been large, with magnitudes greater than 7. Although intraplate continental seismicity is generally at a very low level, recent studies have shown that the seismicity occurs in several linear belts. It can be demonstrated that some of these linear belts are controlled by older trends (see Chapter 11) related to Precambrian, Paleozoic, and Mesozoic structures. Thus, mapping of older structures, both at the surface and their continuation beneath younger sedimentary cover within the continents, may be useful in understanding the trends of modem and potential seismicity.

It is not entirely clear that intraplate continental seismic activity is unrelated to plate tectonics. Studies of earthquake focal mechanism, *in situ* stress measurements, and hydraulic fracturing studies have produced data showing the present stress field within the eastern United States to have a general pattern with the maximum compressive stress trending northeast, but with local complexities. Sbar and Sykes (1973) suggested that the regional stress field is related to the motion of the North American plate and that the earthquake zones are controlled by the presence of unhealed fault zones. If intraplate fault activity is related to plate motion and to the reactivation of older structures, perhaps some intraplate igneous activity is also related to these reactivated fault and structural lines of weakness in the crust. Lipman (Chapter 14) has suggested that the temporal progression of volcanism associated with the Yellowstone volcanic field as well as some older Mesozoic and Cenozoic volcanic trends follow Precambrian structural trends. It is clear that the relations between plate tectonics and intraplate tectonic and igneous activity require continued research to determine how much intraplate activity is related to plate motions and how much is related to independent causes.

CONCLUSIONS

The plate-tectonics model offers an explanation for the present heterogeneity of the external part of the earth. It explains the youth and simplicity of oceanic lithosphere and offers the potential to explain the antiquity, complexity, and evolution of the continental lithosphere. Progress can clearly be made if we (1) develop a three-dimensional understanding of the kinematics, dynamics, and thermal structure of modem plate-boundary systems and at the same time recognize those geological and geophysical features that are unrelated to plate interactions; (2) use this understanding to reconstruct the extent and evolution of ancient systems that form the major elements of the continental crust; and (3) determine the dynamics and evolution of systems that have no modem analogs.

The continents display an internal heterogeneity, that is still poorly understood and that represents an obvious challenge to earth scientists. They are complex but not incomprehensible. It is clear from the discussions above that no single continental terrane contains complete information on plate-boundary and non-plate-boundary systems, both modem and ancient. To understand these systems and continental development requires not only multidisciplinary studies but international cooperation. We are on the threshold of significant advances in our understanding of continents, and our future advances may rival those of plate tectonics as we understand it today.

References

Armbruster, J., L. Seeber, and K. H. Jacob (1978). The northwestern termination of the Himalayan mountain front: active tectonics from microearthquakes, *J. Geophys. Res. 83*, 269-282.

Blake, M. C., Jr., R. H. Campbell, T. W. Dibblee, Jr., D.G. Howell, T. H. Normark, E. A. Silver, and J. C. Vedder (1978). Neogene basin formation in relation to plate-tectonic evolution of San Andreas Fault system, California, *Am. Assoc. Petrol. Geol. Bull. 62*, 344-373.

Bott, N.H.P. (1971). Evolution of young continental margins and the formation of shelf basins, *Tectonophysics 11*, 319-327.

Burchfiel, B.C. (in press). Eastern European Alpine system and the Carpathian Orocline as an example of collision tectonics, *Tectonophysics*.

Coward, M. P., P. R. James, and L. Wright (1976). Northern margin of the Limpopo mobile belt, southern Africa, *Geol. Soc. Am. Bull. 87*, 601-611.

Crowell, J. C. (1974). Origin of late Cenozoic basins in southern California, *Soc. Econ. Paleontol. Mineral. Spec. Publ. 22*, 190-204.

Davis, G. A., J. W. H. Monger, and B. C. Burchfiel (1978). Mesozoic construction of the Cordilleran "Collage," central British Columbia to central California, in *Pacific Coast Paleogeography, Symposium 2*, Pacific Section, Society of Economic Paleontologists and Mineralogists, pp. 1-32.

Dewey, J. F., and K. Burke (1974). Hot spots and continental break-up: Implications for collisional orogeny, *Geology 2*, 57-60.

Hadley, D., and H. Kanamori (1977). Seismic structure of the Transverse Ranges, California, *Geol. Soc. Am. Bull. 88*, 1469-1478.

Irving, A. J. (1976). On the validity of paleogeotherms determined from xenoliths suites in basalts and kimberlites, *Am. Mineral. 61*, 638-642.

Johnson, R. W., D. E. Mackenzie, and I. E. M. Smith (1978). Volcanic rock associations at convergent plate boundaries: re-appraisal of the concept using case histories from Papua, New Guinea, *Geol. Soc. Am. Bull.* 89, 96-106.

Mattauer, M., and M. Seguret (1971). Les relations entre la chaine des Pyrenees et le Golfe de Gascogne, in *Histoire Structurale du Golfe de Gascogne*, Vol. 1. pp. IV 4-1-IV 4-24.

McKenzie, D. P. (1972). Active tectonics of the Mediterranean region, *Geophys. J. Roy. Astron. Soc. 18*. 1-32.

Molnar, P., and P. Tapponnier (1975). Cenozoic tectonics of Asla. Effects of a continental collision, *Science 189*, 419-426.

Saleeby. J. B. (1977). Fracture zone tectonics, continental margin fragmentation. and emplacement of the Kings-Kaweah ophiolite belt. southwestern Sierra Nevada. California. in *North American Ophiolites*. R. C. Coleman and W P Irwin. eds., Oregon Dep. Geol. and Mineral Industries Bull. 95. pp. 141-160.

Sbar. M. L., and L. Sykes (1973). Contemporary compressive stress and seismicity in eastern North America: an example of intraplate tectonics, *Geol. Soc. Am. Bull. 84*. 1861-1882.

Sloss. L. L. (1972). Synchrony of Phanerozoic sedimentary-tectonic events of the North American craton and the Russian platform. *Int. Geol. Congr., 24th session. sect 4*. pp. 24-32.

Tapponnier. P., and P. Molnar—(1976). Slip-line field theory. and large-scale continental tectonics. *Nature 264*. 319-324.

Warner. L. A. (1978). The Colorado lineament: a middle Precambrian wrench fault system. *Geol. Soc. Am Bull. 89*. 161-171.

2

Problems of Pre-Mesozoic Continental Evolution

Leon T. Silver
California Institute of Technology

INTRODUCTION

By virtue of its remarkable ability. to explain currently observed dynamic processes of ocean basins and continental margins, and to account for much of the geological record of the last 200 million years (m.y.), the plate-tectonics model is a rational starting point in interpreting the earlier evolutionary history of the earth. The model's numerous predictions facilitate continued testing and upgrading of its own validity. Since early Mesozoic time, modem continental margins have developed with distinctive physical and chemical characteristics that can serve as bases for comparative studies of older margins, which have been incorporated within the continents in the past. The recent progress of geological studies in North America and the diversity of its geological endowments make the United States a particularly appropriate region for many of these essential investigations.

There is, however, some risk in directing these research programs solely to the paradigms of plate tectonics. Our current knowledge of the earth's evolutionary history includes evidence for many unidirectional changes in earth properties and suggests the need for a cautious approach to the extrapolation of modem geodynamics to earlier geological eras.

Earth energies, forces, processes, material distributions, temperatures, gradients, rates, and other factors all combine to constrain earth dynamics, and all have changed with time. If we are to understand the past fully, other versions of the plate-tectonics model and other models also must be considered, It is vital not to be seduced into allowing modem plate tectonics to assume premature dominance of interpretations of earth history, and therefore the analysis of the early geological and geochemical evolution of the continents deserves a significant fraction of our continental research efforts.

The immense bibliography required to document all the assertions, suggestions, questions, and other provocations contained in this brief chapter would outstrip it in length. I omit them here (with one deferential exception) but suggest that the reader will find most of them contained in the references in the other papers contained in this volume, especially the papers by Hanson (Chapter 13),

This paper is Contribution Number 3271. Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91125.

Muehlberger (Chapter 15), and Kay and Kay (Chapter 12).

TIME CALIBRATION OF THE TECTONIC RECORD

It is the essential dimension of time, recorded in the paleontological time scale, in the various radiometric systems, and in the compelling paleomagnetic record, that revealed for plate tectonics its true complex dynamic character. Extension and testing of plate-tectonic concepts to the earlier geological record will continue to require time calibration. For the earlier Phanerozoic, fossil zonation is still the most precise time-resolving tool available. Its intercalibration with radiometric techniques is required for establishment of dimensions of physical time and for extension to unfossiliferous segments of the geological column. Continuing efforts in this direction deserve high priority. The paleomagnetic pole-reversal sequences and the apparent pole-migration paths have not yet been integrated into a continuous record that can be used for precise age assignments, but significant progress is being made in this direction. The establishment of a comprehensive magnetostratigraphic record in the sedimentary and igneous columns of the Phanerozoic deserves high priority.

For the first 4 billion years (b.y.) of earth history, various radiogenic isotope systems in rocks and minerals, increasingly supplemented by biological, chemical, and remanent magnetic information, will be the principal tools for both general time assignments and precise time resolution. The characterization of long-term evolutionary additions of radiogenic ^{87}Sr, ^{206}Pb, ^{207}Pb, ^{208}Pb, and ^{143}Nd to the major crust and mantle reservoirs provides important genetic and temporal information on the nature and history of the source regions of continental rocks.

Current research in the precise measurement of isotopic systems promises time resolution on the order of ±0.1 percent in the oldest crustal rocks, under favorable conditions and with suitable lithologies. It is essential, however, that integrated petrological and geochemical research into the methods for interpreting the geochronology of high-temperature, high-pressure metamorphisms be continued. Among the most challenging problems in tectonics are the time constraints for orogenic evolution at depth in the lower crust and upper mantle leading to granulite and eclogite facies metamorphism.

CONTINENTAL TECTONICS FROM THE PRE-ARCHEAN THROUGH THE PALEOZOIC

The Interplanetary Connection

The stunning revelations of 20 years of space exploration in our 4.5-b.y.-old solar system, capped by direct lunar sampling and field studies and by sophisticated orbital probing of Mars, Mercury, Venus, and the Jovian systems, have focused scientific attention on the first 500-800 m.y. of history in the terrestrial planets. The phenomenology of initial planetary aggregation and differentiation has been developed with extraordinary results for the earth's moon.

Internal lunar differentiation must have been completed in the first 100 or 200 m.y. Catastrophic surficial modifications from external bombardments are inferred in the interval from the moon's formation to about 3.9 b.y. ago. Comparative studies of the surfaces of the other terrestrial planets also reveal profound effects of early impact histories.

On the earth, the oldest rocks, peculiarly enough, are identified at close to 3.8 b.y. old and represent early differentiated granitic crust. Is this timing simply fortuitous? Are there still older rocks to be found in the continents? Or is this a most significant time convergence implying a critical shared episode in the genesis of the earth-moon system? Is this a time of common cessation of planetesimal infalls? Planetary fission? Planetary near-collision? Some other phenomenon?

In these, the earth's oldest rocks, the record commonly appears to be a montage of multiple episodes of igneous and metamorphic activities. No direct evidence for impact features has been reported; but impact involvement cannot be precluded. Can various constructional elements of the montage be stripped away to unveil cataclysmic impact products? What is certain is that crustal dynamics was well established more than 3.5 b.y. ago but in forms and with rates that are still obscure. Was this *rigid-plate* tectonics as it is modeled today?

The Archean Need not be Arcane

The earliest geological era for which there exists a significant record is the Archean, extending from 3.8 to 2.5 b.y. ago. It is no longer quite so cryptic or mysterious as it was when first tentatively identified. In North America, Greenland, Eurasia, Africa, and Australia, the earliest continental records have been extended back to 3.3-3.8 b.y. Although the rocks approaching 3.8 b.y. of age invariably are strongly metamorphosed and complex, they establish a minimum age for planetary differentiation and for the production of continent-forming crustal materials. Such rocks are found in west Greenland and in Minnesota.

In the Barberton Mountains of South Africa and in the Pilbara region of western Australia, there exist essentially unmetamorphosed sedimentary and volcanic strata that are clearly 3.2 to more than 3.4 b.y. old. In their existence is the earliest record of the initiation of the very-long-term crustal stability, which is the unique distinction of continental structures. In the rocks of the Barberton Mountains, there is evidence for primitive life forms. Clearly, these regions contain the best surviving records of the early surface environments of the earth. At these sites the crust, the atmosphere, and the hydrosphere interacted to produce at their interfaces the premonitory biological, chemical, and physical environments in which biochemical assemblages have evolved into the life forms and the life environments that we know.

Widely recognized features of Archean shields are the numerous great linear arrangements of mafic volcanic rocks called greenstone belts, commonly more than 1000 km in length and hundreds of kilometers in width. Because immensely valuable massive sulfides and other types of ore deposits are associated with them, the details of their local structure and stratigraphy, petrology, and mineralogy have received considerable attention. However, their broad spatial distributions, their isolation in "seas" of ancient granitic gneisses, their spatial and temporal relations to granite crust development are still incompletely understood. Among a number of features that distinguish Archean greenstone belts are abundant Mg-rich, Si-poor basaltic lavas, which are unlike most recent basalts. These ancient rocks, called komatiites, require much higher melting temperatures, probably reflecting the much steeper temperature gradients that prevailed in the crust and upper mantle of the Archean earth. The relationship of these rocks to ancient crust is not understood. Were they part of ancient global oceanic floors subsequently deformed into linearity? Are they part of ancient island arcs produced at convergent plate boundaries? Do their geometry, age ranges, and chemistry compare directly with Phanerozoic systems?

Although Archean continental crustal rocks are present in large areas of the northern United States, their complete distribution and range in ages are still not established. Their lateral boundaries with younger rocks are invariably concealed or complicated by igneous intrusion or intense shearing. Their vertical distribution and lithospheric roots (extension to or below the base of the crust) are unknown. Their relationship to ancient ocean basins is not clear. Were there once supercontinents, subsequently fragmented and disaggregated prior to the Proterozoic? Whether "rigid" crustal plates, as we know them, existed in the Archean has not been established. This is a central question for understanding early continental tectonics.

Proterozoic Perimeters in Space and Time

It is in the Proterozoic era, extending from 2500 to about 600 m.y. ago, that the opportunities for comparison of pre-Phanerozoic tectonics to plate tectonics are most favorable. The boundary in time between the Archean and Proterozoic eras is commonly set arbitrarily at about 2500 m.y. ago. The boundaries in space between Archean and Proterozoic rocks in North America are far more diverse and potentially offer critical information on tectonic processes.

The last major magmatic culmination in the Archean appears to be recorded on all continents in the interval 2500-2650 m.y. ago. In the following half aeon, 2000-2500 m.y. ago, well-documented evidence for significant orogenic and magmatic arc development is surprisingly sparse. Cratonic sedimentary sequences with plateau basalts (diabases) and felsic volcanic rocks are the dominant lithologies of this period and are well preserved on all the continents. There is, therefore, a significant possibility that at approximately 2500 m.y. ago a global transformation in the earth's dynamic systems occurred and was followed by 500 m.y. of comparative continental stability. This possibility is still incompletely seen, much less documented. The search for a more complete orogenic record of this seemingly quiet period needs new momentum on all continents.

A distinctive aspect of the sedimentary record found in the cratonic and circumcratonic basins of the early Proterozoic is the abundance and diversity of sedimentary ore deposits. The banded iron formations, which are the principal world source of iron ores, reached maximum development between 2500 and 1800 m.y. ago. The great uranium-rich conglomerates of the Canadian Huronian formations and the South African gold-uranium ?? glomerates (e.g., Witwatersrand) were formed in ?? interval. In each case, the sources of the metal?? localization mechanisms are in need of better ?? In the United States, iron, uranium, and gold ores are major targets for exploration; research into the early Pro-terozoic history of the continent conceivably may provide a substantial contribution in these important areas of resource development.

Broadly viewed, widespread Proterozoic orogenies in North America appear only in two significant time intervals, 2000-1600 m.y. before the present (B.P.) and 1300-900 m.y.B.P. In Canada, these are recognized as the Hudsonian and Grenvillian orogenies, respectively. Most of the United States and Mexico appears to be underlain by crustal materials that seem to have been initiated in these two great episodes (see Chapter 15 for geographic distributions). Although their total distribution is incompletely known, especially under the Phanerozoic cover of modem continental margins, it commonly has been inferred that the major Proterozoic orogenic belts are sutured against the Archean crustal remnants. Suggested examples are along the Grenville front in Ontario and Quebec, the Nash Fork-Mullen Creek shear zone in Wyoming, and in portions of Minnesota and Michigan. The geological style of each of these "sutured" perimeters appears to be sufficiently distinctive, however, to make all such boundaries worthy of specific study. As the proposed loci of continent-continent collisions, or transforms, they constitute a major class of targets for the study of Precambrian crustal dynamics.

Are there unique elements of the Proterozoic tectonic record that differ from more ancient or more recent tectonic products? Perhaps! At least two possibilities may be considered.

The midcontinent gravity high of the United States (see Chapter 11) is an anomalous geophysical feature with no precise analog recognized anywhere else on earth. It is the consequence of a late Proterozoic continental rifting event in which older Proterozoic and Archean granitic crust was invaded in an arcuate belt by great masses of basaltic magma over a distance of about 2000 km, widths of 50-150 km, and probably to depths of 20-30 km (base of granitic crust). The timing of this catastrophic rifling event is remarkable in that most of it occurred about 1100-1120 m.y. ago, essentially synchronous with a major plutonic magmatic arc culmination recorded throughout the not too distant Grenville province. If the Grenville province was developing at a convergent plate boundary, or was introduced by continent-continent collision at this time, the Keweenawan basaltic riff has no clear equivalent in comparable plate-tectonic settings developed in the last 200 m.y. Does this reflect a fundamental change in crustal or lithospheric strength and rigidity from the Proterozoic to the Mesozoic?

Extensive geochronological studies have suggested a second remarkable Proterozoic phenomenon in North America. In the interval from 1400-1500 m.y. ago, in newly developed mid-Proterozoic crust, a magmatic event of enormous magnitude and with distinctive lithologies occurred without identifiable relation to synchronous orogeny or sedimentation. From Labrador to California (and beyond?) hundreds, perhaps thousands, of anorthosite-syenite and rapakivi granite plutons with surface diameters up to 100 km perforated the continental crust in a belt at least 6000 km long and 1000 km wide. Within that belt these plutons became the most important volumetric constituent in the upper crust. The volume of magma is staggering for an anorogenic setting; the material and energy sources are a first-order problem in crust-mantle evolution; the state of crustal stress that accompanied the emplacement is still speculative. Has this type and magnitude of event been recorded in the Phanerozoic or in the Archean? If so, the record has not yet been read.

In summing up some of the problems of Proterozoic dynamics in such cursory fashion, it is perhaps best to return to a first-order question. Can the nearly 2 b.y. of Proterozoic time essentially be represented by two great orogenic intervals, each approximately 300-400 m.y. long and separated from each other and the preceding Archean and succeeding Phanerozoic orogenic culminations by a quiet interval of comparative durations. In the detailed record, we know of some regionally significant orogenic events in the so-called quiet intervals, but the prominence of these two great orogenic periods, corresponding to the Hudsonian and Grenvillian, is clearly seen not only in North America but on several other continents. What does this indicate for episodicity, even periodicity, in global dynamics during this time interval? Is there possibly an extension of periodicity as well as episodicity beyond the Proterozoic? Were these orogenies shared events on the margins of supercontinents that subsequently were fragmented and redistributed? Such propositions represent some of the outstanding questions for those who would extend post-Paleozoic plate tectonics into the Precambrian.

ARE PALEOZOIC PLATES MORE PALATABLE THAN PRECAMBRIAN?

In a summary. discourse of this nature. casual provocation is easier than thoughtful analysis. Current views of plate tectonics may not be adequate to explain even the Paleozoic orogenies. With the extraordinary. body of Paleozoic faunal data on provinciality, ecology, and climatology, with increasingly detailed stratigraphic and paleomagnetic records, with over 100 years of intense geological study of Paleozoic orogenic provinces on both sides of the Atlantic Ocean, plus an essentially straightforward reconstruction of Pangaea at the end of the Paleozoic. a critical analysis of the effectiveness of the plate-tectonics model in the Paleozoic seems possible. Indeed. an impressive number of rational interpretations of various early Paleozoic plate-tectonic features have been made. They range from old marginal geosynclines and island arcs (northeast North America, Great Britain) to ocean floor (Bay of Islands) to closing oceans (Iapetus) to continental collisions (Appalachian and Caledonide orogenies) to a coherent supercontinent (Pangaea). Given the immense complexity yet coherehcy of most of the diverse data inputs, it would be difficult not to accept what J. Tuzo Wilson (1966) suggested were the consequences of the rifting, spreading, and closing of an ancient ocean in the early to mid-Paleozoic (the Wilson cycle).

The Devonian to Permian history of the same region, however, involved a widespread, perplexing orogenic episode (the Hercynian), which is found superimposed on the earlier orogenies, without the obvious plate-tectonic features that characterized the earlier events. Intense folding, high-grade metamorphism, and granitic mobilization of a different chemistry are some of the consequences of the Hercynian (Allegheny) orogenic interval.

A large number of studies attempt to relate the Hercynian episode to hypothetical subduction and convergence, but almost always without broad acceptance. Many thoughtful students of plate tectonics recognize in this case that there are additional dimensions to the phenomena of continental dynamics that have not yet been elaborated in the existing models.

One of the most exciting rewards of the study of pre-Mesozoic continental evolution will be, obviously, the extension and improvement of current tectonics models. It is those workers who exercise admiration for, and careful restraint in, the use of these models from whom we can expect the next generation of state-of-the-art tectonics concepts. The greatest rewards. perhaps, may be found in our increased ability to approach the continental dynamics record as the key to the global dynamics record and, in turn, utilize terrestrial tectonic evolution as a key to the comparative tectonic history of the inner planets of our solar system.

Reference

Wilson, J. T. (1966). Did the Atlantic close and then reopen? *Nature 211*, 676.

II

PLATE-BOUNDARY TECTONICS

3

Complexities of Modern and Ancient Subduction Systems

Warren Hamilton
U.S. Geological Survey

INTRODUCTION

Until the advent of plate-tectonic concepts in the late 1960's, geologists and geophysicists studying the components of ancient orogenic and magmatic terranes on the continents had no actualistic framework within which to comprehend the origins and interrelationships of those components. We now see that modem analogs for many features of the ancient terranes form primarily in continental-margin arcs and island arcs, above oceanic lithospheric plates sliding beneath continental plates or other oceanic plates. These active terranes can be understood in a plate-tectonic framework, and many relationships explained on a genetic basis. Ancient terranes are being increasingly comprehended in terms of analogy with the active ones.

The first decade of plate-tectonic explanations of continental geology has been enormously fruitful. Increasingly, however, our early explanations appear overly simplistic; we have seen the broad relationships but have missed many of the detailed ones because we have visualized plate interactions as much less complex than has been the actual ease. Correspondingly, recent studies of active continental margins and island arcs show them to undergo rapid and complex changes in response to fast-evolving plate interactions. Too little communication takes place in either direction between the scientists primarily studying the modem marine systems and their dry-land counterparts studying ancient systems. Some purportedly plate-tectonic explanations published recently for ancient systems are mere *ad hoc* conjectures, incompatible with the known features of active systems. Conversely, the active-margin students could infer much about their domains by analogy with deeply eroded equivalent tracts exposed on land.

A vast amount of work is required to increase our understanding of the products and processes of plate tectonics. Much more integration of data already obtained is possible. More data need to be gathered in both field and laboratory, with emphasis on the petrological, structural, and geophysical criteria for identifying, distinguishing, and characterizing the component products of plate interactions, on paleontologic and radiometric criteria for defining their ages, and on paleobiogeographic, paleoclimatic, and paleomagnetic studies to constrain their global wanderings.

The brief summary here of complexities of modem arc systems is adapted primarily from my study (Hamilton, 1979) of the onshore and offshore tectonics of the Indonesian, western Melanesian, and southern Philippine regions. The discussion of ancient arcs in western North America is derived mostly from another report (Hamilton, 1978). The reader is referred to those works for elaboration and documentation of the concepts summarized here and for references to the work, by hundreds of other geo-scientists, from which those concepts are derived.

ACTIVE CONTINENTAL MARGINS

Where an oceanic lithospheric plate is subducted beneath a continental plate, a tectonic and magmatic system like that of modem Sumatra or the Andes is developed. Most early continental plate-tectonic papers (including my own), and unfortunately also many current papers, assumed erroneously that in such settings the oceanic plate tips abruptly down at a trench to an inclined trajectory., along which the plate dips directly into the mantle. Bather, it is now obvious that the usual case is that exemplified by Sumatra (Figure 3.1). The subducting plate dips very gently beneath the continental slope, landward of a trench with gently sloping sides, for a distance of 100 km or more; only beyond this distance does the plate roll toward the steeper dip of the inclined Benioff zone as deduced from mantle earthquakes. The continental slope typically rises gently to an outer-arc ridge, landward of which is an outer-arc basin. Along some active margins, the basin is full of sedimentary strata and is expressed

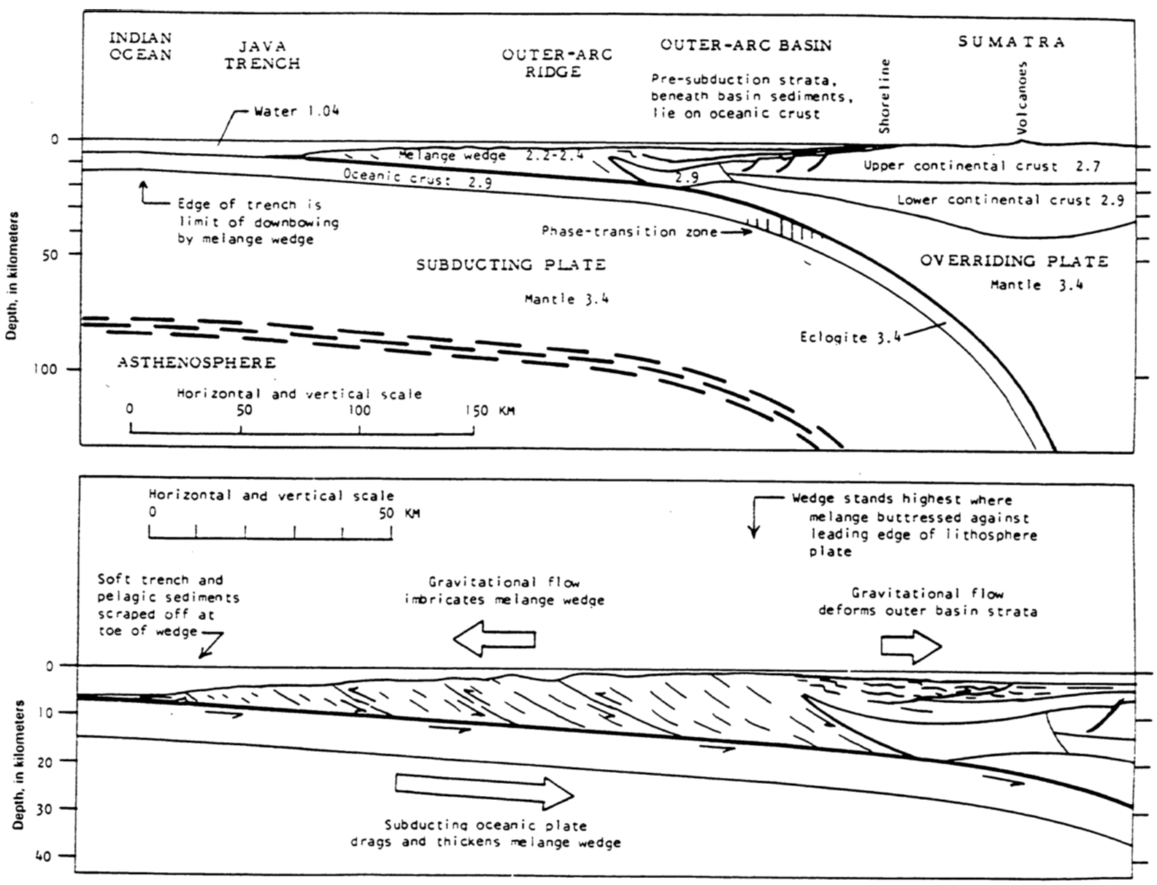


Figure 3.1
Section through the subduction system of the Java Trench, off southern Sumatra. The Cretaceous system of California presumably was similar. The configuration of the mélange wedge and the structure of the outer-arc basin are constrained by reflection profiling, and the shape of the subducting plate is constrained primarily by the location of the reclined Benioff zone of mantle earthquakes. Upper diagram: major components, densities in g/cm^3. Lower diagram: mechanism of deformation of accretionary wedge of mélange and imbricated materials.

bathymetrically as a subhorizontal continental shelf. Seismic-reflection profiles and some drilling show that between the continental-slope seafloor and the top of the subducting oceanic plate there is commonly an accretionary wedge formed largely of sediments scraped off the undersliding plate and imbricated and converted to melange along shear structures that dip moderately landward, sharply discordant to the gentle dip of the oceanic plate beneath. Large accretionary wedges develop where voluminous elastic sediments are conveyed into the subduction system. Such sediments are deposited mostly longitudinally relative to the trench, as turbidites in the trench or as abyssal fans.

Some recent papers have interpreted voluminous complexes of mélange and sheared rocks in ancient subduction settings to be formed primarily by submarine slumping rather than by shearing within the accretionary wedge and its tectonic basement. Such conjecture receives no support from studies of modem trench systems, within which large submarine slides are uncommon.

Available data suggest that an outer-arc basin, like that of Figure 3.1, develops only where the leading edge of the continental plate consists of a strip of oceanic crust, either formed m contact with the continental crust by a previous spreading event or sutured to the continent following previous subduction. The basin is formed by the raising of the leading edge of this strip, not by depression of the center. The raising is presumably a result of the stuffing beneath the strip of accretionary mélange. Where the accretionary wedge is buttressed directly against a leading edge of continental crust, no basin is present.

The active volcanism of the Sumatra system, like that of most subduction systems, is concentrated in a belt above that part of the seismic Benioff zone that is about 125 km deep. The volcanic rocks are of silicic and intermediate compositions and, in part, form calderas developed by eruption from large, shallow magma chambers. It is reasonable to assume that granitic batholiths must now be forming beneath the volcanic belt. The magmatic belt is superimposed on a regional anticline consisting of prevolcanic rocks bulged broadly upward; apparently the crust has been thickened by voluminous magmatic intrusions.

BEHAVIOR OF MODERN ISLAND ARCS

Island arcs do not sit stably on underlying mantle for an era, constantly consuming incoming oceanic lithosphere, although many continental geologists assume them to behave so. Rather. arcs characteristically migrate relative to the plates behind them; they reverse their subduction polarity, or go dead, they collide with each other and with contments; and they are dethroned by strike-slip faults and oroclinal folds. The products of these processes are exceedingly complex aggregates and superpositions of' are materials. Oceanic crust disappears beneath modern arcs at rates reaching at least 13 cm/year, or 1:30 kin, 10^6 year: an arc cannot long face anything except a spreading center without major changes in geography and geometry.

Consider the southern Philippine Islands, an aggregate of the products of six Cenozoic subduction systems that are clearly identifiable and, undoubtedly, others that cannot yet be distinguished. The six recognizable island-arc systems diverge southward from the main southern Philippines and, hence, preserve their individual trenches, ridges, and magmatic arcs. On the west is the extinct Palawan arc, trending northeast from the northern tip of Borneo, active during middle Tertiary time; subduction stopped at about the end of Miocene time, without any collision or other obvious external cause. A fossil trench lies along the northwest side of Palawan; the island is the top of the mélange wedge (formed mostly from quartzose elastic sediments derived from Asia and Borneo and deposited longitudinally in the trench); and an extinct volcanic are lies beneath the sea surface to the southeast. The next eastward subduction system is the Sulu island arc, active with southeastward subduction from at least late Miocene into Pleistocene time and possibly still subducting at a very slow rate. The Sulu are connects northeast Borneo to the Zamboanga Peninsula of Mindanao, and the geology of the onshore continuations of the are suggests a reversal of subduction polarity, from previous northwestward subduction beneath the arc, in Miocene time. Two sectors of the northern extension of the Sulu arc swung together in a Y configuration, intersecting between Zamboanga Peninsula and Negros Island. (Such Y's are evidence for a subduction mechanism by vertical sinking, rather than by inclined injection, of the downgoing plate. Central and eastern Mindanao consists in part of the fused north-trending Sangihe and Halmahera arcs, which collided as the intervening Molucca Sea floor was subducted beneath both of them, and in part of the products of the Cotabato and Philippine subduction systems, which have subduction polarities opposite to those of the precollision arcs. Each of these identified are systems includes mélange wedges, magmatic-arc complexes, and sedimentary basin fills. Further complications have been added by oroclinal folding and strike-slip faulting.

Western Melanesia displays the products of Cenozoic interaction between numerous small lithospheric plates of fast-changing character. formed between the obliquely converging Pacific and Indian-Australian megaplates. The Solomon island are now faces southwest. complicated by a sector in which a spreading ridge intersects its trench. but the arc faced northeast until middle Tertiary time. The northwest extension of the Solomon arc. the New Ireland-Manus arc, is sliding northwest past a trench-trench-transform triple junction at the east end of the New Britain arc: new subduction breaks through on the north side of the projection to define another reversal of subduction polarity, so that subduction is again south-westward. The fast-moving transform. offset along short spreading centers, curves westward from the New Britain triple junction to intersect the north margin of central

New Guinea; west of that intersection, subduction southwest beneath New Guinea is very rapid, whereas east of the intersection subduction is slow. The slow-subduction Mussau arc system trends southward to a trench-trench-trench triple junction against the curving New Ireland-Manus arc, which continues past the junction to another trench-trench-trench junction off the north coast of central New Guinea. Northern New Guinea records the collision during Miocene time of a south-facing island arc (itself likely a composite) with a continental margin, mostly stable since the Jurassic rifting away of another continental mass. The collision progressed eastward with time, and the east, or unblocked, part of the are advanced past the end of the New Guinea continent to form the Papuan Peninsula. Subduction, previously northward beneath the island arc, reversed to southward beneath the continent, as enlarged by the addition of the arc and underlying lithosphere, after the collision; only subsequently did subduction reverse beneath the peninsular extension. Strike-slip faulting, oroclinal folding, and other collisions have been additional complications.

Large aecretionary wedges of mélange and imbricated rocks are developed in front of only those island arcs along which voluminous sediments are available on the ocean floor. Abyssal-fan and longitudinal-trench sedimentation can, however, extend several thousand kilometers from a major continental source. Immature oceanic island arcs, represented above the surface, if at all, only by small volcanic islands, lack such sediment supply; these arcs face steep-sided trenches, at which the subducting oceanic plate tips sharply downward and dips steeply beneath the arc, without any broad mélange wedge and underlying gently dipping oceanic plate. Ophiolitic mélanges—pelagic sediments and marie oceanic crustal rocks, in a matrix of serpentinite disrupted and hydrated from oceanic mantle rocks—are known in many orogenic assemblages that are now parts of continents, where they are associated with magmatic assemblages analogous to modem island arcs. These ophiolitic mélanges may mostly represent the small accretionary wedges formed along oceanic island arcs, and contrast with mélanges, dominated by terrigenous elastic sediments that formed along active continental margins.

Modem arcs provide no support for the various hypotheses of back-arc obduction, or flake tectonics. Where slabs of oceanic crust and mantle have ridden onto other rock assemblages and are exposed in young arc terranes, dips and transport directions are as required for those slabs to be the fronts of overriding plates.

MECHANISM OF SUBDUCTION

The inclined descent of subducted lithospheric plates, down to depths of at least 700 km, is commonly visualized as representing either the injection of a slab down an inclined trajectory. that is fixed with regard to the mantle or the gravitational sliding of a slab down such a trajectory. Neither mechanism of subduction can explain the geometry of many modem systems. The Banda. Antilles, and Scotia arcs have subducting trenches that are U-shaped in plan and Benioff zones, hence subducted plates, that are spoon-shaped. Opposed arcs are in the process of colliding in the Molucca Sea and Solomon Sea regions, over subducted plates that have an inverted V configuration in section.

Such geometry suggests that the subduction process below a depth of 100 or 150 km may consist primarily of the vertical sinking of the downgoing plate and that the apparent dip of a subducting plate is a function of the rate of advance of the plate hingeline over the mantle beneath the subducting plate. This rate would be equal to the convergence rate only where the subducting plate is fixed with regard to the mantle beneath it and where no marginal sea spreading occurs behind the advancing arc. The sinking velocity of the subducted lithosphere must be a function of the density contrast between subducting slab and surrounding mantle and, hence, must increase, at an exponentially decreasing rate, with age of that slab. Given a constant rate of hinge advance, the dip of a subducted slab should be a function of its age.

PRE-CRETACEOUS ISLAND-ARC AND MÉLANGE ASSEMBLAGES OF THE WESTERN UNITED STATES

The Cretaceous and early Tertiary development of the western conterminous United States was dominated by subduction of large plates of oceanic lithosphere directly beneath the continent. Similar activity of Andean type occurred during parts of Triassic and Jurassic time also, but much of the subduction recorded during those periods occurred beneath oceanic island arcs. The arcs consist dominantly of submarine basaltic and andesitic lavas, breccias, and tuffs; their intrusive and volcaniclastic equivalents; carbonate strata; and diverse accretionary-wedge materials including ophiolitic mélanges. These assemblages resemble those that characterize modem oceanic arcs. These arcs were added to the west edge of the continent when intervening oceanic lithosphere was wholly subducted beneath either the continent or arcs advancing toward it. Complexities in these ancient, accreted arc terranes are still poorly understood but are numerous enough to indicate that complex histories, comparable with those of modem Indonesia, Melanesia, and the Philippines, are recorded. We must be dealing with the products of consumption of dozens of large and small lithospheric plates and with complex, jumbled aggregates of are components that were in part pre-assembled far from their final resting place along the continental margin.

After Precambrian rifting and the formation of a new ocean in the direction that is now west, the western margin of North America during early and middle Paleozoic time trended from west-central Idaho to east-central Cali fornia. This margin was stable tectonically, and on it developed an oceanward-thickening wedge of continental shelf strata. In Late Devonian and Early Mississippian time, deep-water strata from the west, with some of their underlying oceanic crust, were shoved onto the shallow-water shelf strata, and both were pushed eastward above the continental basement in the Nevada sector (Figure 3.2). This event may record the collision of an island arc with the continent: but the arc itself did not remain attached to the continent. for another period of stable-margin sedimentation followed, with deep water again to the west. Presumably the polarity of subduction reversed beneath the collided arc, which migrated away from the continent, opening an ocean basin behind it. Another similar collision occurred against Nevada near the middle of Triassic time. Again, deep-water sedimentary rocks and oceanic basement were driven eastward upon shallow-water strata, and both moved eastward over the continental crust; but this time, the causative island arc remained attached to the continent. The Permian and Triassic marine faunas of this island arc differ only moderately from those of mainland North America, so the pre-collision wandering of the are need not have been at vast distances from the continent. The collision was followed by a reversal of subduction polarity, and Upper Triassic arc magmatic rocks were erupted through the old continental crust, above the subduction system dipping eastward from the Pacific edge of the added island arc.

Still greater motions of lithospheric plates are required by juxtapositions of other are and mélange assemblages that yield fossil or paleomagnetic evidence for distant sites of origin. One belt of such assemblages occurs in the tectonic-accretion terrane from the southern Sierra Nevada of California to the Yukon Territory in Canada; it is characterized by shallow-water Upper Permian limestones bearing tropical Asian fossils. Particularly distinctive among these fossils are fusulinids—algal-grazing, calcareous-test protozoa—of the verbeekinid family. These are widespread in correlative deposits of paleotropical Eurasia yet are completely unknown in paleotropical mainland North America. (The equator of the time trended through Texas, in the direction now northeastward, so appropriate habitats were available; but the American fusulinid assemblages are quite different from the Eurasian ones.) These fossils occur in limestone blocks in mélanges, in reef limestones capping atolls, in the scraped-off complexes along the paleomargin of North America, and in part between island-arc belts. Subduction of more than 10,000 km of oceanic lithosphere beneath North America during Late Triassic and Jurassic time, and beneath island arcs that were themselves added to North America in Jurassic time by subduction of intervening oceans, is required. Oceanward of this tract in British Columbia and southern Alaska is still another tract of island-arc and related rocks, characterized by southern-hemisphere paleomagnetic latitudes. The Cretaceous and early Tertiary pattern of relatively simple subduction of oceanic lithosphere beneath the continent, with minimal involvement of island arcs. followed the addition of these various terranes to North America.

CRETACEOUS AND CENOZOIC CALIFORNIA

Subduction Configuration

Both the geology of California and the seafloor-spreading history of the northeast Pacific Ocean require that during Cretaceous and early Tertiary time oceanic lithosphere was being subducted beneath California. The Cretaceous components of most of California fit the model of Figure 3.1, when palinspastic restoration is made to reverse the disruption and offset by late Cenozoic strike-slip faults and other structures. In the west, forming most of the Coast Ranges (Figure 3.2), is the accretionary wedge. the Franciscan complex of mélange and imbricated rocks—mostly of abyssal uppermost Jurassic, Cretaceous, and lower Tertiary elastic sediments but including fragments of oceanic crust and mantle and of seamounts, and masses of high-pressure, low-temperature metamorphic rocks. Shear structures in the chaotic Franciscan complex typically dip steeply to moderately eastward, yet gravity studies indicate that the oceanic lithosphere beneath the wedge dips only gently eastward. Along the east side of the Coast Ranges, coeval outer-arc-basin strata, the Great Valley facies, lie with depositional contact upon oceanic crust and upper mantle, which in turn is in thrust contact with the tectonically underlying Franciscan accretionary wedge. Lower Great Valley strata were deposited at abyssal depth on oceanic crust, a narrow strip of which remained attached to the continent while Cretaceous subduction proceeded. The strata lap eastward across the oceanic crustal strip and in the subsurface butt against what was in Cretaceous time the moderately sloping edge of continental crust. The leading edge of the oceanic strip was rotated upward concurrent with Cretaceous subduction.

Farther inland, the Sierra Nevada batholith, granitic rocks mostly of Cretaceous age, formed synchronously with the formation of the Franciscan accretionary wedge and with the development of the Great Valley outer-arc basin. The batholith represents the subjacent equivalent of a volcanic are such as that of modem Sumatra. Batho-lithic rocks of any one age display a cross-strike increase in the ratio of potassium to silicon. characteristic of active volcanic terranes formed above subducting plates, and the distribution of the rocks defining the gradient indicates moderately steep subduction dips compatible with the Franciscan and Great Valley geometry. in a system like that illustrated in Figure 3.1.

The magmatic record of Cretaceous and early Tertiary. time in the western states indicates that the dip of the subducting slab was moderately steep and relatively constant during Early and early Late Cretaceous time but that the dip was markedly gentler during late Late Cretaceous

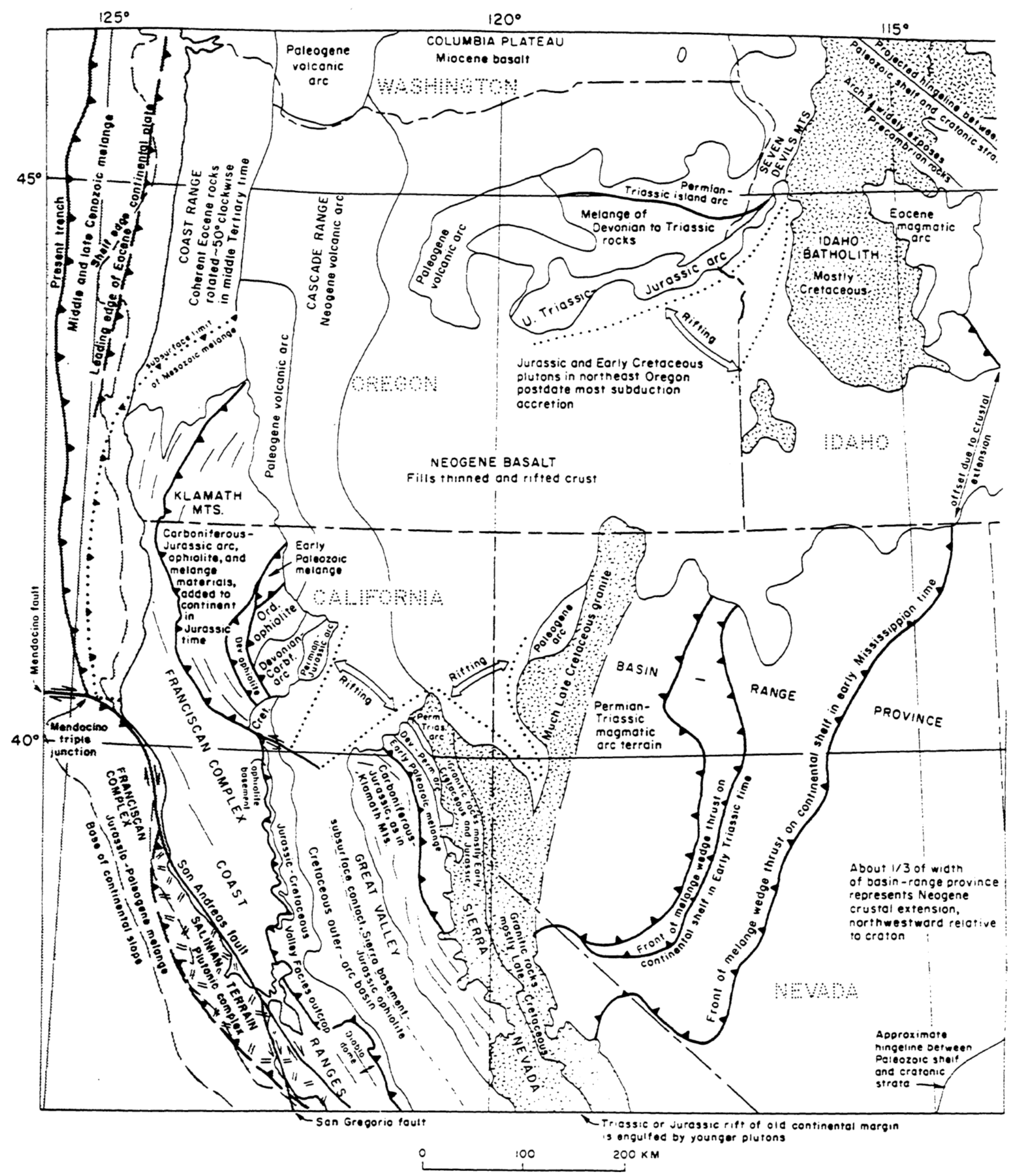


Figure 3.2
Selected tectonic elements of the west-central United States. The present distribution of these elements was affected greatly by disruption during middle and late Cenozoic time.

time, perhaps as the result of an increased convergence rate between the North American and subducting plates. Magmatic-arc patterns changed complexly during early Tertiary time, indicating corresponding variation in subduction dips. In Oligocene time, for example, the sub-ducting slab was apparently segmented: the southern segment decreased in dip northward from moderately gentle to very gentle. there was an abrupt change across a line trending northeast from southern California to northern Colorado. and the northern segment decreased m dip northward from moderately gentle to moderately steep (see Chapter 14). Much of the Tertiary complexity may be due to the subduction of oceanic lithosphere of varying age and density as the East Pacific Rise, offset along transform faults, approached the continental margin and young lithosphere of variable temperatures was being subducted. Conversely, the spreading history and geometry of eastern Pacific plates, since wholly subducted beneath North America, can perhaps be inferred from the magmatic-arc migrations.

In the Klamath Mountains region of northwest California and southwest Oregon, the Franciscan mélange wedge lies directly beneath crustal assemblages that were part of the continental crust in Cretaceous time (Figure 3.2). No outer-arc-basin strata lie to the east. Here, the leading edge of the continental plate was formed of continental crust; the narrow strip of attached oceanic crust is missing, and no outer-arc basin developed.

Late Cenozoic Disruption

During Cretaceous and early Tertiary time, the Pacific margin of California was a relatively simple system of parallel tectonic and magmatic belts, controlled by the vigorous subduction then occurring beneath all of California. The subduction was more rapid than the spreading of the East Pacific Rise offshore to the west, so the rise moved progressively closer to the trench despite the active spreading. The oceanic plate east of the rise was moving northeastward relative to North America, whereas the plate west of the rise was moving northwestward. When the rise itself hit the subduction system about 30 million years (m.y.) ago, the western oceanic plate came into direct contact with a segment of the continental margin, and strike-slip motion resulted. The strike-slip boundary, lengthened with time, and its north limit migrated northwestward, as spreading ridge and trench collided along progressively more of the continental margin. The oceanic lithosphere, which had been subducted shortly before the transition. was young and hot, so the change from subduction to strike-slip regimes occurred when the continental margin was underlain by soft, weak mantle.

The results of these changes and conditions can be seen in southern California and northwest Mexico (Figure 3.3). The Cretaceous magmatic arc (Sierra Nevada. Peninsular Ranges, and Baja California batholiths), outer-arc basin. and Franciscan mélange wedge are all present—but their distribution is scrambled. The simplest and youngest major disruptive structure is the San Andreas Fault, which is now the dominant break along which the western Pacific plate is moving northwest past the continent, carrying Baja California and coastal California along. The San Andreas offset of a well-documented 300 km and the concurrent opening of the Gulf of California have occurred within the past 5 m.y.

Motion patterns during the transition period. between 30 m.y. and 5 m.y. ago, from subduction to strike-slip regimes were much more complex. In the southern Coast Ranges (northwest part of Figure 3.3), strike-slip faults underwent pre-San Andreas offsets of at least 200 km; yet these faults probably do not cut the Transverse Ranges, which trend across them in the south. In the Coast Ranges, the westward progression from magmatic arc to outer-arc basin to Franciscan mélange wedge is present, variably truncated by Cenozoic faults; but then more outer-arc-basin strata were present beyond the Franciscan, in the Santa Ynez and Santa Monica Mountains of the western Transverse Ranges. The complete progression is present also in the Vizcaino Peninsula sector of Baja California. Along southern California south of the Transverse Ranges, however, the sequence is partly repeated: offshore in the San Nicolas Island-Santa Rosa Island sector, outer-arc basin strata are again present, out of place to the west of Franciscan; and Franciscan is present west of these strata, apparently in proper succession. One explanation possible for some of these and other relationships is that a western sliver of outer-arc basin and Franciscan terranes slid northwestward from Baja California, along a right-lateral strike-slip fault, around the Franciscan terrane now enclosed to the east; the leading edge of the sliver impinged on the continent, and the rest rolled clockwise past it, forming the western Transverse Ranges by rotation. The required rotation and northward translation are consistent with the paleomagnetic orientations of lower Miocene basalts in the Santa Monica Mountains and islands to the west.

Other styles of crustal disruption affected the western states farther inland. Distributed crustal extension produced the Basin and Range province (see Chapters 8 and 9) of blocks bounded by downward-flattening normal faults. The amount of total late Cenozoic extension is in dispute; my interpretation, more extreme than most is that between 30 and 50 percent of the present width of the terrane represents extension. The nature of the transition from brittle fracturing and block rotations of the upper crust to hypothetical uniform extension in the deep, hidden lower crust is also in dispute. Studies in progress of deeply eroded basin-range terranes in southeast California and southern Arizona suggest that the level of deformation beneath the horizontal bases of rotated fault blocks is characterized by giant boudins or mullions, having lengths of as much as several tens of kilometers, and thus by deformation that is partly ductile and partly brittle. In

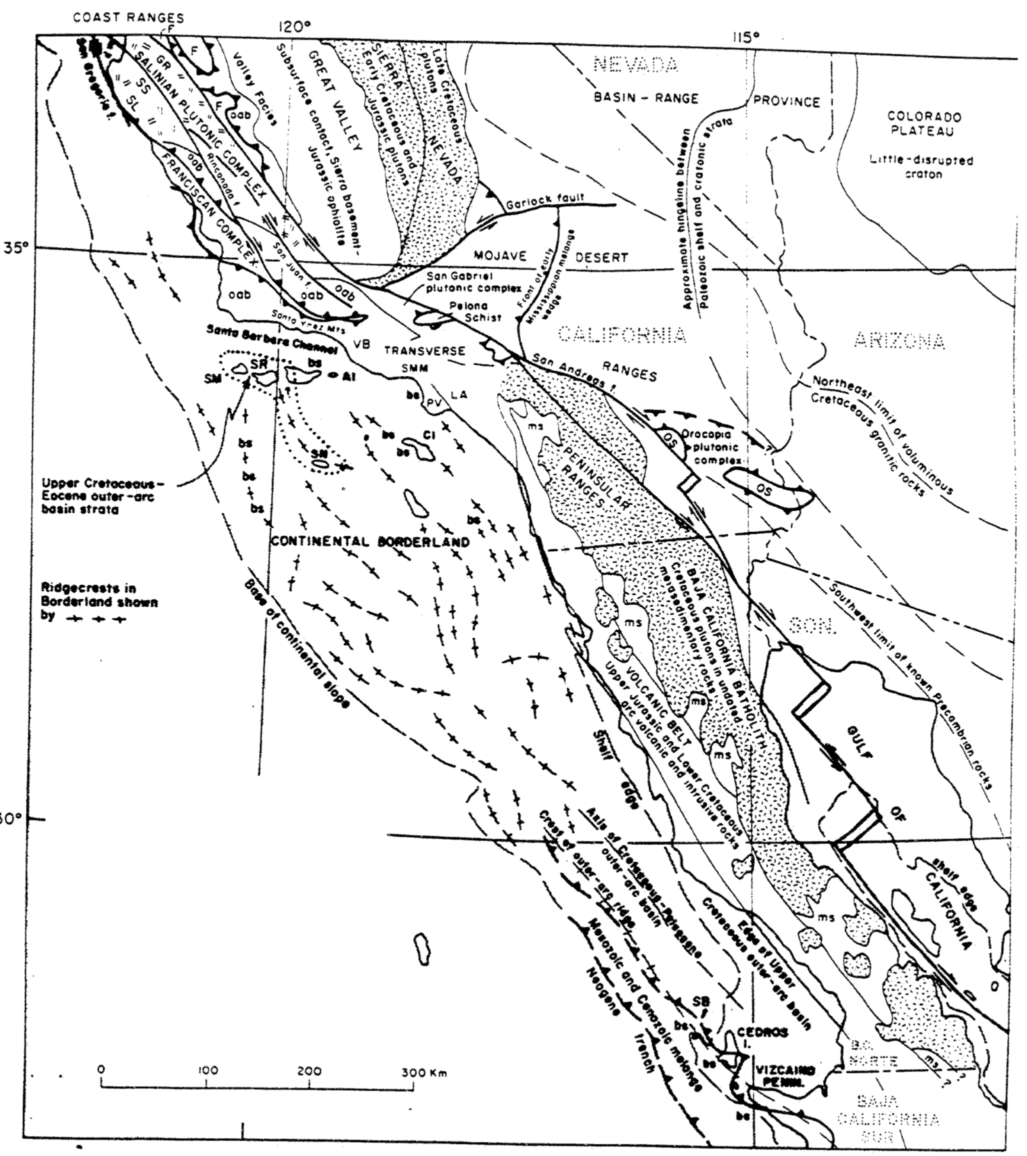


Figure 3.3
Selected tectonic elements of the southwestern United States and northwestern Mexico.

the northwest states, from northwest Nevada and southern Idaho northwestward, continental crust older than middle Tertiary was thinned by extension in some areas and completely rifted in others, the volumes having been filled subsequently mostly by middle and upper Cenozoic basalts (Figure 3.2). Other aspects of Cenozoic disruption are discussed in Chapter 8.

References

Hamilton, W. (1978). Mesozoic tectonics of the western United States. in *Pacific Coast Paleogeography Symposium 2*. Pacific Section, Society of Economic Paleontologists and Mineralogists. pp. 33-70.

Hamilton W. (1979). Tectonics of the Indonesian region. *U.S. Geol. Surv. Prof. Paper 1078*, 34.5 pp.

4

Intracontinental Rifts and Aulacogens

Kevin Burke
State University of New York at Albany

INTRODUCTION

The word rift is used in a variety of ways in geology, most of which have nothing to do with major structures. Gregory (1921) was the first to use and later to popularize the word rift for a large-scale structure when he described the East African rift. He defined a rift valley as a long depression between parallel normal faults. Here a modified version of Gregory's definition is used, and rifts are defined as "elongate depressions overlying places where the entire thickness of the lithosphere has ruptured in extension." This omits reference to parallel faults because many familiar rifts, for example, the Connecticut rift, have a major fault on only one side, and in some rifts it is hard to be sure whether a boundary is a fault or a steep monoclinal flexure. By referring to "the entire thickness of the lithosphere" emphasis is given to rifts as large-scale structures, and small depressions are excluded. The reference to extension is to distinguish rifts from the other major fractures penetrating the whole lithosphere such as transform faults and those at convergent plate boundaries.

Rifts are the commonest major lithospheric fractures because the strength of the lithosphere is least in tension. Since rupture of the lithosphere is involved in rift formation, they are commonest where the lithosphere is thinnest, that is, in the oceans. This review is solely concerned with continental rifts that are now and probably always have been much less abundant than oceanic rifts.

Two great changes in continental-rift studies over the last 10 years are the recognition that intracontinental rifts are numerous and that rifts have developed within continents in a range of structural environments. These are all environments with dominantly extensional tectonics, and active and ancient rift occurrences are associated with a number of different extensional plate-tectonic regimes. The Wilson cycle is the concept that permits application of plate-tectonic principles to the record of historical geology within the continents.

RIFTS IN THE WILSON CYCLE

Our discussion of continental structure and evolution comes 12 years after J. Tuzo Wilson first drew attention to the best way in which to apply the revolutionary recogni tion of the plate structure of the lithosphere to an understanding of continental geology. Wilson (1968) argued, before the American Philosophical Society, that plate tectonics showed that on the surface of the earth, oceans are opening in some places and closing in others. The history of the earth's surface can therefore, he suggested, be considered as a record of the opening and closing of oceans. Since ocean floor disappears by subduction, the record of these cycles has to be sought within the continents, and looking for records of these cycles is the most powerful way of interpreting continental geology. Wilson analyzed the cycles in terms of oceanic evolution from youth (continental rupture) to old age (continental collision) and elaborated his analysis in the second edition of his textbook (Jacobs *et al.*, 1972). Dewey and Burke (1974) later suggested that complex interwoven cycles of ocean opening and closing of the kind recorded within the continents be called Wilson cycles (Figure 4.1).

RIFTS AT CONTINENTAL BREAKUP

Rifts occur at all stages of the Wilson cycle, because extensional tectonics can develop at all stages, but those associated with continental rupture are both particularly well developed and particularly accessible to study. The East African rift system is the best known continental rift system, and progress in its understanding has accelerated since appreciation that its present activity dates only from the beginning of the Neogene when the African plate appears to have come to rest with respect to the underlying mantle convection pattern (Burke and Wilson, 1972). Distinction of the current episode of rifting from a very similar episode in Africa 100-200 million years (m.y.) ago that was associated with the breakup of Gondwana has emerged. Rift studies that define the timing of events accurately have become of particular importance in recent years and seem likely to remain important in the future because the establishment of how rifting relates to other tectonic events will only be possible if the relative timing of the phenomena are accurately known.

Two other rift properties that have been firmly established in the East African rift system are that rift faults commonly follow and reactivate old structures (McConnell, 1974) and that rift igneous rocks are almost wholly mantle derived. Detailed studies of igneous rocks associated with the East African rift system have emphasized the widespread occurrence of alkaline rocks, most of which appear to have last equilibrated with the mantle at depths of 60-100 km, and many of which appear to have suffered complex subsequent histories. Comparable alkaline rocks occur in many rift systems, but so do tholeiitic rocks, suggestive of equilibration at much higher mantle levels.

Studies of sediments in the East African rift system are numerous but unsystematic. They come from such diverse fields as petroleum exploration and fossil hominid research, but there has as yet been little attempt to make tectonic inferences from these sediment studies or to study the rift-filling process as a whole. On the other hand, volcanology and geomorphology have been intensely pursued in East Africa and tectonic inferences made from them.

Teleseismic studies have been among the most fruitful of geophysical investigations in the East African rift system and have shown the peculiar nature of the mantle below the rift system as well as the extensional character of most rift earthquakes. Although local networks have yielded important results, the limited number of modem seismic stations on the African continent has proved somewhat of a barrier to research. Gravity studies are well advanced in East Africa and have proved most susceptible to tectonic analysis in which gravity information has been coupled with other data. Seismic refraction studies, for example, have indicated the presence of high-velocity dike-like objects in some rifts, and this has helped to constrain gravity interpretations (Figure 4.2).

The East African rift system has been, since its discovery, one of the most studied of continental rift systems, and it is likely to continue to yield important insights m the understanding of rifts. Further seismic and rift-sediment studies appear particularly promising.

If the East African system continues to propagate and eventually becomes a system extending across the African continent, then a new ocean can develop and the African plate may split in two. If this happens, many currently active rifts will be left in one or the other of the two new continents. Such rifts will have failed to develop into ocean. Failed rifts of this kind stretching into continents at a variety of angles from Atlantic-type ocean margins are numerous and are rapidly becoming the best known of all fossil rifts (Burke, 1976a, and Figure 4.3). This is because

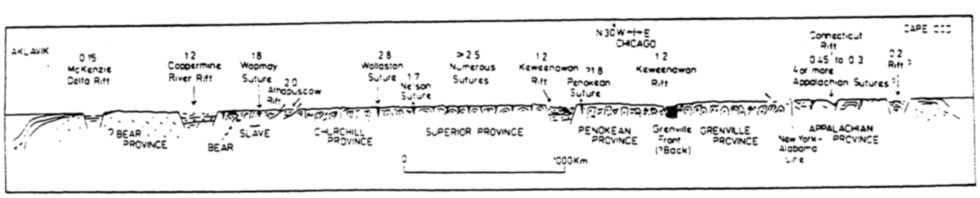


Figure 4.1
Simplified cross section across North America from the Beaufort Sea to George's Bank showing numerous rifts and sutures produced through operation of the Wilson cycle over the last 4×10^4 years. Numbers are approximate ages m years times 10^9.

the sediment fill of the failed rifts is oil-bearing in some places. Oil is produced from three major environments in rifts of this class: nonmarine graben facies at the bottom of the rifts, marine limestones formed during rapid subsidence of the rifts a few tens of millions of years after continental rupture, and younger elastic material prograding off the continent along or across the rift. The last environment can be subdivided into shallow, often deltaic, and deeper water subenvironments. Reflection seismic studies and drill holes associated with petroleum exploration can yield accurate information on how rifts subside over intervals of tens of millions of years. Characterization of individual rifts, or parts of rifts, as behaving thermally like continents, oceans, or in some intermediate state may prove possible.

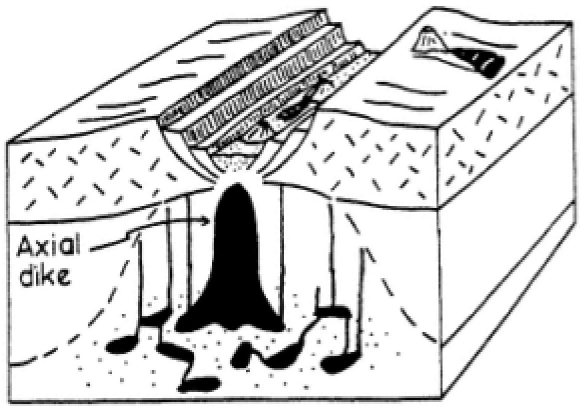


Figure 4.2
Sketch (by A. M. C. Sengor*) of structure of typical, active, intracontinental rift showing thinned lithosphere, an axial dike em-placed at a shallow level, listric faults, volcanic rocks, and sediments.

The widespread occurrence of failed rifts trending at varied angles to Atlantic continental margins may provide some explanation of the difficulty that has faced attempts in many areas to locate precisely the boundary between continent and ocean (Figure 4.4).

Although the general behavior of failed rifts at Atlantic margins seems to follow a regular pattern, there are exceptions and much remains to be learned about this economically important class of rifts. For example, the South Atlantic opened about 125 m.y. ago, and many failed rifts associated with this event have passed through normal sequences of events becoming occupied by marine sediments in the Aptian about 15 m.y. later. However, the Rio Salado and Colorado rifts (Urrien and Zambrano, 1973) continued to be filled with nonmarine sediments until much later (about 80 m.y. ago), and the Benue Trough in Nigeria behaved like a small ocean and appears to have opened and shut in a small-scale Wilson cycle of its own over the interval between 125 m.y. and 80 m.y. ago (Burke and Dewey, 1974).

A small population of rifts appears to have developed at Atlantic-type ocean margins some time after initial continental rupture, The Sirte rift in Libya, formed during the Lower Cretaceous facing a Tethyan margin that ruptured 100 m.y. earlier in Triassic time, is the best documented example. It appears to owe its origin to intraplate strains established during the events that led to opening of the South Atlantic. This is again a case for which accurate timing of structural events is essential in understanding. The Sirte rift is one of many rifts without reported contemporary volcanism, which was at one time widely thought to be an essential feature of rift-valley development. The rapid subsidence rates revealed by studies of rift sedi

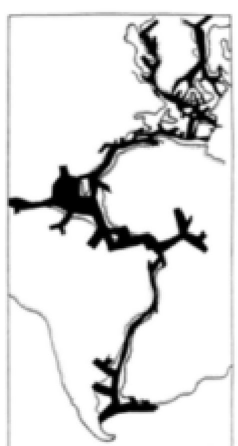


Figure 4.3
Map of a closed Atlantic Ocean showing areas (in black) in which rift-valley lithosphere, formed mainly in association with continental rupture, is well developed.

ments (rates as high as 0.5 km per m.y. are reported by BRGM *et al*., 1975) indicate that fast contemporary cooling rates existed beneath the rifts. Whether igneous rocks are actually erupted at the surface during rift development probably depends on a variety of factors of less than fundamental importance.

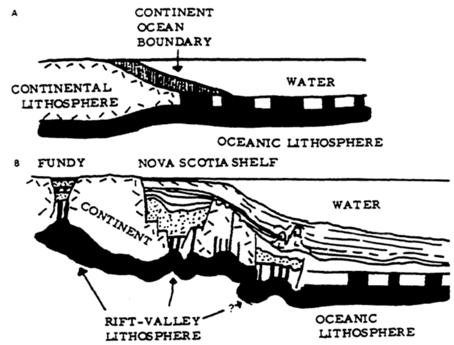


Figure 4.4
Section view comparing an idealized continent-ocean boundary (A) with an interpretation of the boundary off Nova Scotia with rift valleys and transitional lithosphere in a zone about 500 km wide between continent and ocean (B).

CONVERGENT BOUNDARY RIFTS

Rift developments at Andean continental margins and in island arcs are most prominent at the place where the lithosphere is thinnest, that is, along the line of the volcanoes themselves. These rifts, known in such active Andean arcs as New Zealand and Sumatra-Java, achieve their greatest tectonic significance only when the volcanic arc splits to form a marginal basin as, for example, happened in the Japan Sea at the beginning of the Miocene (Sillitoe, 1977). Rifts of Basin and Range type appear associated with imperfect transform motion along continental boundaries, but few fossil examples have, as yet, been recognized.

COLLISIONAL RIFTS

Perhaps the greatest variety of rift development is associated with continental collisions. Not only do the failed rift systems formed at rupture become reactivated as rifts striking into the collisional mountain belts forming the population of objects that Shatski (1947) called *aulacogens*, but a whole new set of rifts, of which the Lake Baikal and Rhine rifts seem the best active examples, are set up as a result of intracontinental strains established during the collision. These objects have been called *impactogens*, and they can be readily distinguished by their geological history from aulacogens.

Besides these two major classes of collision-related rifts, final closing of an ocean is accompanied by complex rift production. This is well illustrated in the Mediterranean, where formation of the rifts of Corsica and Sardinia 20 m.y. ago accompanied rotation of these islands away from France (Dewey *et al*., 1973) and where formation of the Aegean rifts within the last 20 m.y. has been related to the westward motion of Turkey with respect to the Black Sea and the Mediterranean (Dewey and Sengor*, in press). If the closure of the Mediterranean continues, rifts of the Aegean and Corsican-Sardinian types are likely to be compressed out of recognition, and such structures are probably preserved only as obscure areas within mountain belts that show signs of late-stage extension before final compression. For this reason, rifts of this kind are of less general interest in continental geological history than the four major rift classes: continental rupture rifts, failed rifts at Atlantic margins, aulacogens, and impactogens. A generalization of this type always has exceptions, and the Devonian rift basins of Norway (Horn, Solund, and Hornelen) form such an exception (Steel, 1976). They appear to have formed within the Caledonides at continental collision and closely resemble the Thakkola rift of the high Himalaya today.

Aulacogens

Although the aulacogen concept is over 30 years old, widespread use of the term among scientists in the United States dates only from attempts to interpret the aulacogen concept in plate-tectonic terms (Burke, 1977a). Shatski used the word only for rifts striking into fold belts, and it seems unwise to use it for all ancient rifts that have failed to develop into oceans. Shatski also emphasized that the depositional and structural history of the aulacogen begins at the same time as that of the fold belt into which it leads, thus eliminating the structures now being called impactogens. An unfortunate fact in the history of the use of the word aulacogen is that one of Shatski's type examples, the Dneiper-Donetz structure, is a failed rift that strikes not into a fold belt but into the North Caspian depression, an area probably underlain by ocean floor formed in the Devonian that has been covered with a great thickness (~14 km) of sediments but has, so far, escaped both subduction and obduction.

Although aulacogen histories can be diverse, a twofold division can often be made into an earlier continental rupture and ocean-opening phase with typical rift features, including alkaline igneous rocks, basal coarse clastics, and stratigraphically higher evaporites and a distinct postcollisional phase with coarse clastic sediments derived from the collisional mountain belt and strike-slip faulting. Structurally, the outstanding features of many aulacogens are the enormous thicknesses of sediments and volcanics that they contain. The best-documented thickness is that of the southern Oklahoma aulacogen, where more than 13 km of rock accumulated during the Paleozoic (Ham and Wilson, 1967), roughly two thirds being deposited in the first rifting event and the remainder in association with collision. Rift reactivation is a widespread phenomenon, and the deposition of a thick sequence at collision over the initial rift sequence, as happened in Oklahoma, indicates that the first event modified the local lithosphere in a way that made it behave abnormally even hundreds of millions of years later. Shatski pointed out that the upper member of an aulacogen sediment pair commonly extended over an area wider than that of the initial rift, and this seems a particular case of the more general observation that numerous major intracontinental basins, such as the Chad, Paris, and Michigan basins, overlie older rifts (Burke, 1976b).

Impactogens

In map view an impactogen resembles an aulacogen as it is also a rift structure striking into a fold belt. Distinction between them starts from ascertaining whether rift history goes back as far as the opening of the ocean the closure of which is represented in the fold belt. If it does not, and rifting dates only from the time of ocean closure in the fold belt, the rift is an impactogen. For example, both the Rhine graben and the Polish trough strike from northern Europe into the Alpine fold belt. The start of the geological history of the former is contemporary with the mid-Eocene Meso-Alpine collisional event (Sengor* *et al*., 1978), while that of the latter started in the Triassic. Thus the Rhine graben is an impactogen, while the Polish trough is an aulacogen.

Although the Rhine graben reaches to the edge of the Alpine system in the Jura, impactogenal rifts associated with the Himalayan collision, the Baikal and Shansi graben develop far from the main collision zone. Molnar and Tapponnier (1975) suggested that these rifts formed as a consequence of the eastward escape of China from crushing between the viselike jaws of India and Asia. The Baikal and Shansi rifts are linked to the main collisional mountains in Tibet and the Himalaya through a system of strike-slip faults and folds.

Timing of the Himalayan collision is not yet as well established as that of the main Alpine collision so that it is hard to know how closely the start of faulting in Shansi and Baikal coincides with the collision marked by the Indus suture.

Impactogens seem particularly well developed in association with the Hercynian fold belt of Europe. The Hercynian Ocean appears to have opened in two stages: one in medial Devonian times along a rift extending from western Britain to the Urals—the timing of this event indicates that the rift may have been formed as an impactogen related to collision in the Caledonides—and the other in early Carboniferous times along a rift with a very different strike in Iberia. This second rift could have been an Acadian impactogen (Burke and Sawkins, 1977). Although evidence of the formation of the Hercynian Ocean is incomplete, evidence of its closure at the beginning of the Permian is much stronger. Rifts developed over much of northwestern Europe at the beginning of the Permian (the Oslo graben being the best known such rift), and these rifts are strong candidates for interpretation as impactogens. The rifts of northwest Europe established in this Permian event have been repeatedly reactivated in Triassic, Jurassic, Cretaceous, and Cenozoic times (Whiteman *et al*., 1974). This is most obvious in the North Sea, where geological development of the huge oil province has been dominated by rift reactivation (Woodland, 1975). A possible explanation of renewed subsidence in a long inactive rift, so typical of the North Sea, is that a thermal pulse in the mantle induces transition from a basaltic to an eclogitic mineralogy in a subrift axial dike system.

NORTH AMERICAN INTRACONTINENTAL RIFTS AND AULACOGENS

Active Rifts

Active rifts in North America are dominated by those of the Basin and Range, discussed in Chapters 8 and 9, but the Rio Grande rift is a structure whose development although closely linked to that of the Basin and Range is sufficiently isolated to be considered on its own (Figure 4.5). Recent studies are showing that the Rio Grande is an intracontinental rift with many typical features perhaps including an axial dike system. Because volcanic rocks with a wide range of ages are relatively abundant in the rift and because sediment-filled basins with determined stratigraphy are developed at various places along the 1000-km length of the system, an unusually good opportunity exists in the Rio Grande rift for establishing a developmental history through much of Neogene time, and this may help to elucidate the way in which a rift system can develop through a few million years for comparison with the only other well-documented history, that of the 45-m.y.-old Rhine rift.

North America's major active rift, the Rio Grande, is clearly a prime target for many types of geological and geophysical studies that will help to improve understanding of its deep structure and that of continental rifts in general.

Reactivated Rifts

A number of ancient rift systems are at present showing signs of reactivation. This reactivation is most obvious in seismic activity (e.g., in the Reelfoot rift) or in seismicity and the occurrence of youthful faulting [e.g., the Lake George-Lake Champlain rift (Burke, 1977b; Isachsen *et al*., 1978)]. Studies in these systems may help to illuminate the mechanisms by which intraplate stresses are relieved on pre-existing structures and because the occurrence of earthquakes in these isolated tectonic areas may constitute environmental hazards (Sykes, in press).

Figure 4.5

Major Mesozoic and Cenozoic rifts of North America. Most are associated with the development of Atlantic-type ocean margins (based on Burke et al., 1978).

Failed Rifts at Atlantic-Type Margins

An Atlantic-type margin extends from the Yucatan Peninsula around the Gulf of Mexico, along the eastern seaboard of the United States and Canada to Baffin Bay and continues past the Arctic Islands of Canada to the north slope of Alaska. The northern margin was formed in a rupture event of roughly medial Jurassic age, and the Atlantic and Gulf of Mexico boundaries in an event of Late Triassic age. The Labrador Sea-Baffin Bay boundary dates from Late Cretaceous time. Failed rift systems associated with these various events show great diversity, and comparative studies are likely to be fruitful because different rifts show different features. Thus the McKenzie River delta is the best example in North America of a great river flowing down a failed rift to build a prograding delta over ocean floor (the Mississippi, although an example of this phenomenon, occupies a more complex structural environment).

The Labrador and Baffin failed rift systems are largely offshore, and knowledge of them depends mainly on petroleum exploration, although the rift margin of Disko Island has received attention especially for its unusual metallic iron-bearing basalts.

The best-exposed failed rifts at the Atlantic margin are the Triassic rifts of the Appalachian region. Detailed structural studies of some areas are completed, and rift sedimentation is being investigated. The opportunities for an improved understanding of rift structure and development are great. A peculiarity is the lateness of a major (tholeiitic) basaltic episode in these rifts which took place when rift activity had progressed for about 20 m.y. and was apparently almost contemporary with the beginning of Atlantic ocean-floor formation.

Failed rifts of Late Triassic age around the Gulf of Mexico are deeply buried under the salt deposited soon after ocean opening in that area and few research results have, as yet, been published on these relatively inaccessible features.

Aulacogens

The early development of aulacogens associated with the opening of the Iapetus Ocean along the site of the Appalachians in late Precambrian times has been considered by Rankin (1976), and there is clearly an opportunity in many of these structures for integrated stratigraphic, structural, petrological, gravity, magnetic, reflection, refraction, and teleseismic studies (Figure 4.6). The progress of research has been uneven in rifts of this age. For example, the southern Oklahoma aulacogen is structurally and stratigraphically well known (it contains over 75,000 oil and gas wells), but the petrology of igneous rocks at its base is poorly known and its deep structure has not been analyzed geophysically. Proterozoic rifts that are widely interpreted as aulacogens are known from various places along the western Cordillera between southern California and Alaska. The Amaragosa feature in Death Valley (Wright *et al.*, 1974) is perhaps the best described. Emphasis in this review has been on the importance of understanding the relative timing of events in rift studies. In the Precambrian the poor resolution of most dating is a serious drawback in rift studies, and many ambiguities are at present irresolvable.

The Keweenawan rift system provides a good example. The Keweenawan rift is linked westward to the feature causing the midcontinent gravity and magnetic anomalies, which are of typical rift type (Halls, 1978). At its eastern end, the Keweenawan rift can be linked to the rift known from gravity, magnetics, and drilling to underlie the lower peninsula of Michigan. Farther south this feature becomes diffuse within the Grenville province and may have been damaged in the Tibetan-style events (Dewey and Burke, 1973) accompanying the Grenville collision. It seems possible that the Keweenawan rift system was formed in association with the opening of an ocean that closed in formation of the Grenville province, and occurrence of the Seal Lake and Gardar rifts to the northeast of the Keweenawan system is perhaps supporting evidence. However, resolution of these varied Precambrian events in time is too poor to warrant detailed attempts at historical analysis, and the Keweenawan could, on present evidence, be an impactogenal system formed at the Grenville collision.

Figure 4.6

Major Proterozoic and Paleozoic rifts of North America. Stippled areas are rifts formed at the end of the Proterozoic in association with opening of the Iapetus Ocean, most of which became aulacogens during the Paleozoic (based on Burke et al., 1978).

In spite of this limitation, the Keweenawan rift and its subsurface extensions are features of peculiar interest. Their width is noticeably greater than that of most rifts that fail before the ocean-opening stage, and a great variety of volcanic, intrusive, and sedimentary rocks occupy these wide structures. Analysis of relationships within and between these units by as many geological and geophysical techniques as possible is likely to be very rewarding. Geochemistry of Keweenawan igneous rocks is already thoroughly investigated, and no doubt further refinements are possible.

The oldest well-described aulacogens in North America are those of the Canadian Shield, especially the Athapuscow structure of Great Slave Lake (Hoffman, 1973), which is now being recognized to show strong signs of convergence and is perhaps a Proterozoic Benue trough. Less well known are rifts striking into the circum-Ungava collisional fold belt near Richmond Gulf and Cambrien Lake, but limited evidence indicates that these are aulacogens rather than impactogens.

With recognition that the Penokean orogeny embodies collision of continental objects and shows reactivation to the south rather than to the north, features of the southern Canadian Shield (perhaps including the Huronian) may prove to have been deposited in aulacogens.

CONCLUSIONS

Active intracontinental rifts and rifts with ages extending back as far as Early Proterozoic occur in North America, and study of them should help to show how the continental lithosphere ruptures in extension and how, once ruptured, the lithosphere continues to develop in places where oceans fail to form. Relating these numerous rifts to the various stages of the Wilson cycle appears feasible, and analysis in these terms seems the most potent way of applying the revolutionary understanding of plate tectonics to the tectonics of the continents.

Rifts are so varied that almost all display some features that are absent or poorly developed in others, so that what is worth studying varies greatly from rift to rift. However, in most places an integrated approach to rift studies involving thorough field or subsurface geological assessment followed by application of selected appropriate geophysical and geochemical techniques appears likely to be most rewarding.

References

BRGM, ELF, ESSO, SNPA (1975). Geologie du bassin d'Aquitaine (Atlas with 24 plates), Paris.

Burke, K. (1976a). Development of graben associated with the initial ruptures of the Atlantic Ocean, *Tectonophysics 36*, 93-112.

Burke, K. (1976b). The Chad Basin: an active intracontinental basin, *Tectonophysics 36*, 197-206.

Burke, K. (1977a). Aulacogens and continental breakup, *Ann. Rev. Earth Planet. Sci. 5*, 371-396.

Burke, K. (1977b). Are Lakes George and Champlain in Neogene grabens reactivating early Paleozoic rifts?, *Geol. Soc. Am. Abstr. Programs 9*, 247.

Burke, K., and J. F. Dewey (1974). Two plates in Africa during the Cretaceous?, *Nature 249*, 313-316.

Burke, K., and F. J. Sawkins (1977). Were the Rammelsberg, Meggen, Rio Tinto and related ore deposits formed in a Devonian rifting event?, *Geol. Soc. Am. Abstr. Programs 9*, 916-917.

Burke, K., and J. T. Wilson (1972). Is the African Plate stationary?, *Nature 239*, 387-390.

Burke, K., L. Delano, J. F. Dewey, A. Edelstein, W. S. F. Kidd, K. D. Nelson, A. M. C. Sengor*, and J. Stroup (1978). Rifts and Sutures of the World. Unpubl. Rep. to Geophysics Branch, ESA Div. NASA/GSFC, Greenbelt, Md.

Dewey, J. F., and K. Burke (1973). Tibetan, Variscan, and Precambrian basement reactivation: products of continental collision, *J. Geol. 81*, 683-692.

Dewey, J. F., and K. Burke (1974). Hot spots and continental breakup: some implications for collisional orogeny, *Geology 2*, 57-60.

Dewey, J. F., and A. M. C. Sengor* (in press). Aegean and surrounding regions: *Geol. Soc. Am. Bull*.

Dewey, J. F., W. C. Pitman, W. B. F. Ryan, and J. Bonnin (1973). Plate tectonics and the evolution of the Alpine system, *Geol. Soc. Am. Bull. 84*, 3137-3180.

Gregory, J. W. (1921). *The Rift Valleys and Geology of East Africa,* Macmillan, New York, 479 pp.

Halls, H. C. (1978). The late Precambrian geology of the central North American rift system, in *Tectonics and Geophysics of Continental Rifts*, I. B. Ramberg and E. R. Neumann, eds., D. Reidel, NATO, Dordrecht, Holland, pp. 111-121.

Ham, W. E., and J. L. Wilson (1967). Paleozoic epeirogeny and orogeny in the central United States, *Am. J. Sci. 265*, 332-407.

Hoffman, P. F. (1973). Evolution of an early Proterozoic continental margin, *Phil. Trans. Roy. Soc. London A273*, 547-581.

Isachsen, Y. W., E. P. Geraghty, and S. F. Wright (1978). Investigation of Holocene deformation in the Adirondacks Mountain dome, *Geol. Soc. Am. Abstr. Programs 10*, 49.

Jacobs, J., R. D. Russell, and J. T. Wilson (1972). *Physics and Geology,* 2nd ed., McGraw-Hill, New York, 622 pp.

McConnell, R. B. (1974). Evolution of taphrogenic lineaments in continental platforms, *Geol. Rundsch. 63*, 389-430.

Molnar, P., and P. Tapponnier (1975). Cenozoic tectonics of Asia: effects of a continental collision, *Science 189*, 419-426.

Rankin, D. W. (1976). Appalachian salients and recesses: late Precambrian continental breakup and the opening of the Iapetus Ocean, *J. Geophys. Res. 81*, 5605-5619.

Sengor*, A. M. C., K. Burke, and J. F. Dewey (1978). Rifts at high angles to orogenic belts: tests for their origin and the tipper Rhine graben as an example, *Am. J. Sci. 278*, 24-40.

Shatski, N. S. (1947). Structural correlations of platforms and geosynclinal folded regions, *SSR Akad. Nauk Izv. Geol. Ser. 5*, 37-56.

Sillitoe, R. D. (1977). Metallogeny of an Andean continental margin in South Korea: implications for opening of the Japan Sea, in *Island Arcs, Deep Sea Trenches and Back-Arc Basins*. M. Talwani and W. C. Pitman, eds., Maurice Ewing Series 1, Am. Geophys. Union, Washington, D.C., pp. 303-310.

Steel, R. J. (1976). Devonian basins of western Norway—sedimentary response to tectonism and to varying tectonic context, *Tectonophysics 36*, 207-224.

Sykes, L. R. (in press). Intra-plate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism and other tectonism post-dating continental fragmentation, *Rev. Geophys. Space Phys*.

Urrien, C. M., and J. Zambrano (1973). The geology of the basins of the Argentine continental margin, in *The Ocean Basins and Their Margins*, A. E. M. Naim and F. G. Stehli, eds., Plenum, New York, pp. 135-166.

Whiteman, A. J., G. Rees, D. Naylor, and R. M. Pegrum (1974). North Sea troughs, *Tectonophysics 26*, 39-54.

Wilson, J. T. (1968). Static or mobile earth, the current scientific revolution, in Gondwanaland revisited, *Am. Philos. Soc. Proc. 112*, 309-320.

Woodland, A. W., ed. (1975). *Petroleum and the Continental Shelf of North-West Europe*, Halsted-Wiley, London, 501 pp.

Wright, L. A., B. W. Troxel, E. G. Williams, M. T. Roberts, and P. E. Diehl (1974). Precambrian sedimentary units of the Death Valley Region, in *Guidebook, Death Valley Region*, Death Valley Publishing Co., Shoshone, Calif., pp. 27-35.

5

Evolution of Outer Highs on Divergent Continental Margins

Martin A. Schuepbach and Peter R. Vail
Exxon Production Research Company

INTRODUCTION

The development and use of multichannel seismic profiling and deep-drilling technology have led to an increased understanding of the earth's crust. This paper discusses the application of these developments to the nature and evolution of the boundary between continental and oceanic crusts at divergent (passive) margins. Seismic profiles across divergent continental margins and regional geological studies indicate the occurrence of highs near the boundary that probably formed during the late rift phase of divergent margin evolution. Our seismic and regional geological studies have provided us with sufficient data to develop a model for the evolution of these outer highs. In order to test the model on the evolution of outer highs on divergent margins, extensive geophysical surveys, emphasizing regional multichannel seismic profiles, need to be undertaken to investigate the tectonic and stratigraphic variations visible on geophysical data. Near-surface data from outcrops, piston cores, and dredge samples can provide additional important constraints for more accurate geophysical interpretations. From the combination of these geophysical and geological data, optimum drill sites can be located to test the geophysical interpretations and to obtain samples and logs of the earth's crust and overlying sedimentary section.

The results of such studies on the outer portions of divergent margins should provide us with a knowledge of (1) the character, age, and position of the continental-oceanic crustal boundary for better plate reconstructions; (2) the existence, nature, and evolution of the outer high for a better understanding of mantle processes active during the formation of a divergent margin; (3) the nature of the overthickened oceanic crust commonly present in the area adjacent to the oceanic-continental crustal boundary, for better understanding the role of volcanism in seafloor spreading; and (4) the type and controls of late-rift-phase sedimentation, for determining the reasons for the widespread occurrence of salt and organic-rich shales during this phase of seafloor spreading. A knowledge of the above factors in a number of different areas would provide the needed information to understand accurately the nature and evolution of the continental-oceanic boundary, on divergent margins.

STRUCTURAL-STRATIGRAPHIC RELATIONS AT DIVERGENT CONTINENTAL-OCEANIC CRUSTAL BOUNDARIES

Geological Observations

The outer high is defined as the relatively high area between the oldest oceanic crust and the onlapping late rift sediments deposited on adjacent continental crust. Based on a number of geological and seismic profiles across divergent continental margins, the outer high seems to be a common feature formed during the late-rifting phase of pull-apart before the onset of "passive" drifting. In many cases, the outer high does not exist as a present-day high, because of subsidence and tilting during the drift phase. The widespread occurrence of outer highs leads to the assumption that they occur along most pull-apart divergent margins and thus are a part of their tectonic evolution. Divergent margins that evolved through shearing, such as the Ivory Coast, Africa, do not have similar appearing outer highs. Figure 5.1 shows the location of regions studied for this report.

Observations of cross sections across divergent margins indicate that outer highs are variable in size, elevation, and amount of faulting. Figure 5.2 shows three diagrammatic examples that illustrate some of the variations of outer highs. They range from a very-low-relief outer high [Figure 5.2(a)] that is only recognizable by subtle onlap on the landward side and the presence of oceanic crust on the seaward side to a large high [Figure 5.2(c)] that is easily recognizable by the large tilted fault blocks.

Outer highs result from broad uplifts or arching along zones 10 to 20 km wide in areas having earlier horst-and-graben (and graben-fill) topography. We believe that the outer highs are more than single horsts formed during the early-rift phase, for three reasons: (1) Present-day aborted rifts do not have throughgoing horsts in their centers, but outer highs are continuous. (2) If the outer high were an initial horst formed during the early-rift stage, continental separation would most likely take place along one side of the initial horst, thus leaving an outer high on only one margin; however, we observe outer highs on both margins. (3) The top of the outer high is higher than the top of (other) horst blocks at the end of the late-rift phase, as evidenced by the onlapping or pinching out of late-rift-phase sedimentary rocks.

In most examples studied where extensive extrusive volcanics are lacking, the top of the outer high is approximately at the same level as the top of the oceanic crust. This indicates that the oldest oceanic crust was formed near sea level, since the shallow marine late-rift-phase sediments typically onlap the continental outer high. Veevers (1977) came to a similar conclusion.

In many areas, the outer highs are complicated by extensive volcanic extrusives and intrusions. The extrusive volcanics may overlie the outer high, the adjacent oceanic crust, or both. The distribution of volcanic extrusives is

Figure 5.1
Areas considered in this investigation.

very irregular. For example, areas with extensive volcanics, such as the Faeroe Island area (Talwani and Eldholm, 1977) and eastern Greenland (Haller, 1971) contrast with areas having virtually no extrusive volcanics in the area of the outer high, such as the west Africa examples discussed later in this chapter. The distribution of volcanics in present-day rifts is similarly irregular. For example, the Rhine Graben has two large volcanic centers, Vogelsberg and Kaiserstuhl, separated by areas of limited volcanic activity (Rhine Graben Research Group for Explosion Seismology, 1974).

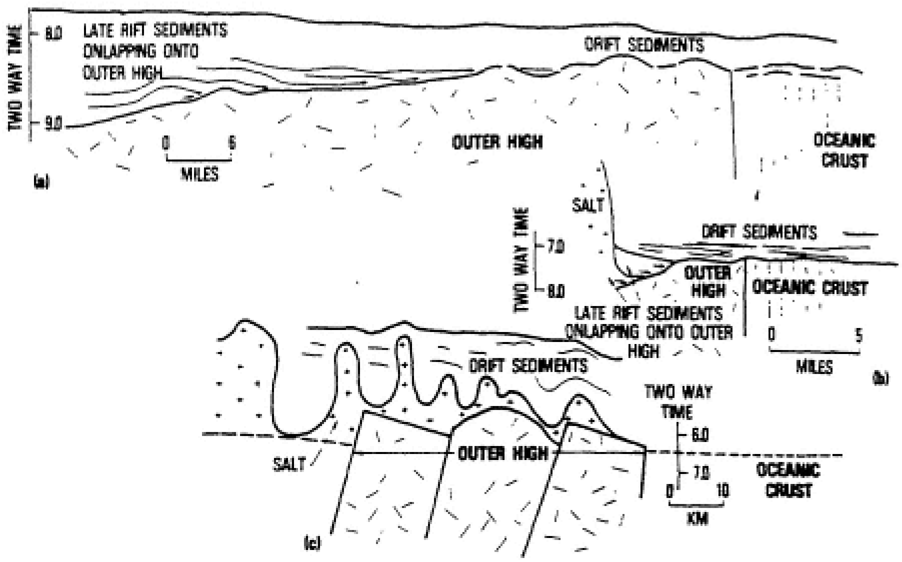


Figure 5.2
Diagrammatic examples of outer highs on divergent continental margins.

In general, divergent margins have a characteristic progression of sediment types that vary as an area progressively undergoes rifting and then drifting. Rifts that form in areas of thick continental crust, such as cratonic shields and areas that have undergone an orogenic event, are in most cases first filled with nonmarine sediments that grade upward to shallow marine sediments that commonly include evaporites, as shown in the West Africa example discussed later in this chapter. Rifts that form in areas of thinned continental crust, such as older inactive rifts, commonly contain marine or interbedded marine and coastal deposits. Figure 5.3 from offshore east Greenland (Haller, 1971) is an example of divergent margin evolution in which the crust has been thinned by earlier rifting.

Post-pull-apart or drift sediments are usually marine, except where large prograding delta complexes are present. In many areas, the marine sediments of the drift phase are prograding carbonate banks and reefs. Marine terrigenous elastics commonly overlie the carbonates or extend throughout the entire drift phase, where carbonates are absent.

Geophysical Evidence for Outer Highs

Outer highs are best recognized on reflection seismic data by looking for the relatively high area in between the seismic expression of oceanic crust and seaward onlap of the reflectors over the predrift unconformity. The reflector at the top of the outer high may be continuous or highly faulted. In addition, reflectors and faults may or may not be observable below the reflector at the top of the outer high. The diagrams shown in Figure 5.2 are patterned after seismic examples and thus illustrate typical reflection patterns of outer highs on divergent margins.

An example of the seismic expression of oceanic crust is shown in Figure 5.4. Note the high amplitude with numerous diffractions that characterize this reflection. In this example, no good reflectors occur below it. This is the common case. In other examples, however, dipping reflectors occur below the top of the oceanic crust. Areas where these reflectors occur are commonly paleotopographic highs, which may indicate local thickening of oceanic crust by volcanic extrusives.

Two multichannel seismic lines from offshore West Africa (A, Figure 5.5, and B, Figure 5.7) are presented to illustrate the seismic expression of outer highs. Both lines extend from the present-day shelf to near the continental-oceanic crustal boundary. This paper only discusses the interpretations in the vicinity of the outer highs. A detailed discussion of seismic line A (Figures 5.5 and 5.6) is presented in Vail *et al*. (in press). A dashed line indicates the interpreted boundary between rift and drift reflectors, and a heavy solid line indicates the top of the outer highs.

On both seismic lines the interpreted presalt early-rift reflectors are bordered on the seaward side by outer highs consisting of blocks bounded by steep faults. Seismic line A (Figure 5.5) shows a relatively smooth surface at the top of the outer high with underlying faults and reflectors. Seismic line B (Figure 5.7) shows large tilted fault blocks. Late-rift salt reflectors onlap or pinch out against the outer highs. Onlap patterns on seismic line A indicate that the paleotopographic high point of the outer high was on its eastern side and the top of the interpreted oceanic crust actually was higher than the western portion of the outer

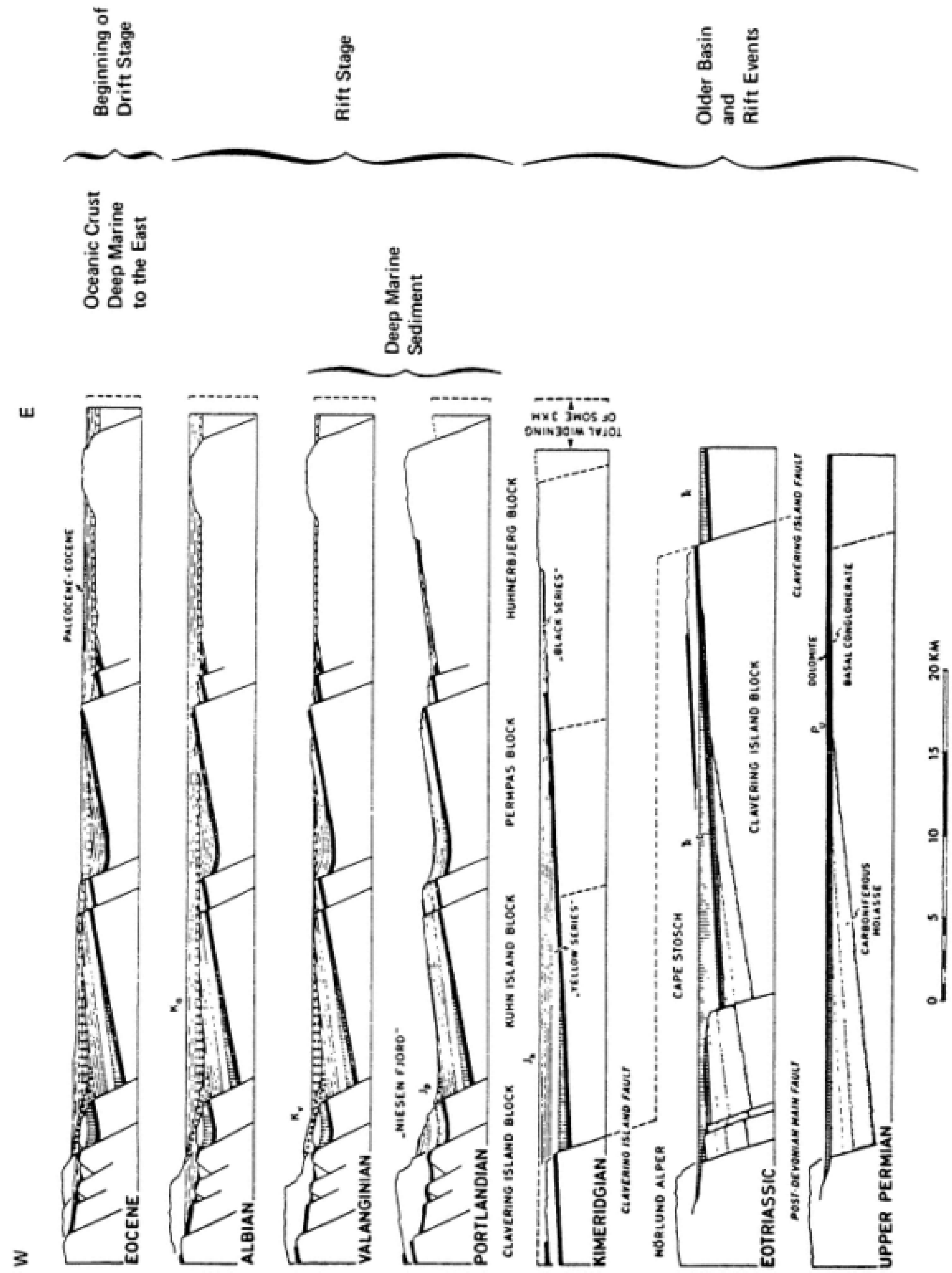


Figure 5.3
Tectonic evolution of the landward part of east Greenland margin (from Haller, 1971). Initiation of the present divergent margin began with Late Jurassic-Early Cretaceous rifting. The resulting grabens are filled with marine sediments, because pre-Late Jurassic rifting thinned the continental crust. The drift phase initiated by the final separation of the continents and emplacement of oceanic crust began in early Tertiary.

high in the early drifting phase. Unfortunately, seismic line B did not extend across the continental-oceanic crustal boundary, but the magnetic quiet zone, interpreted as oceanic crust, is known to exist just west of this seismic line. The interpreted salt interval thins, but does not completely pinch out at the western end of this line.

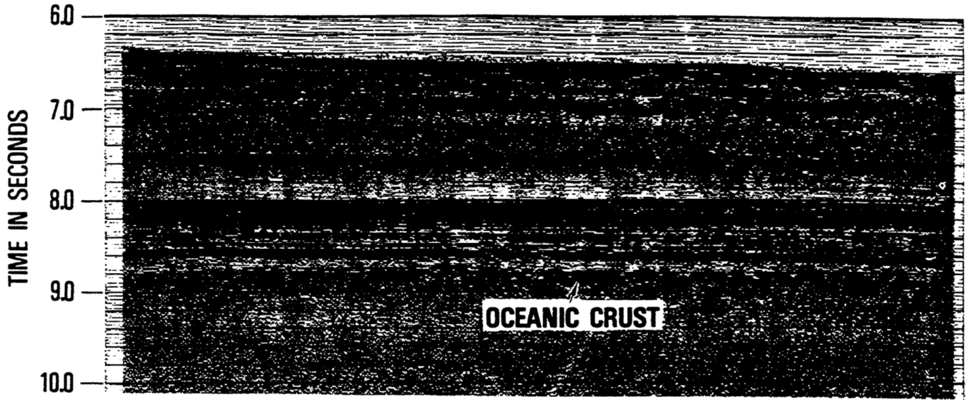


Figure 5.4
Example of the seismic expression of oceanic crust from the Blake-Bahama Basin (from Shipley and Buffler, 1978, courtesy American Association of Petroleum Geologists).

The outer high on seismic line A is bordered by what we interpret to be oceanic crust. The boundary is placed between reflection patterns similar to those characteristics of oceanic crust and to the outer high.

Both seismic lines are controlled by drilling on the shelf and by regional outcrop geology. Based on these data, line A rift sediments include Triassic and Lower Jurassic (Hettangian) red beds and Lower Jurassic carbonates and anhydrites. Based on its stratigraphic position, we interpret the thick salt on line A to be Early Sinemurian in age, but it could be significantly older or younger. The rift sediments are overlain by drift sediments consisting of a Middle Jurassic to Cretaceous Berriasian prograding carbonate bank and a Cretaceous-Tertiary prograding elastic wedge (see Vail *et al*., 1977).

Wells penetrating the shelf near seismic line B indicate that the rift sediments are Neocomian-Aptian elastics overlain by late Aptian salt. Drift sediments are post-Aptian Cretaceous carbonates and elastics and Tertiary clastics.

No good evidence for volcanics is shown on these sections in the vicinity of the outer highs. However, on seismic sections from other areas, the reflection pattern of volcanics is common and quite variable. Volcanics exhibit considerable variation in paleotopography at the top and/ or base of the unit as evidenced by onlap. Interval reflections within the unit commonly dip at a greater angle than the top or base. In many cases these dipping reflections are high amplitude and continuous for short distances. In other examples, extrusives appear as lens-shaped lobes with a relatively high-amplitude reflection at the top and little to no reflections within the lobe. Interval velocities tend to be intermediate to high in the first example and high in the second. We interpret the first example to be largely interbedded volcanics and the second example to be massive flows. Intrusives commonly appear on seismic data as irregular high-amplitude reflectors, sometimes crossing and sometimes subparallel to reflections from stratal surfaces. Reflections originating from intrusives commonly occur sporadically in the vicinity of outer highs.

Only limited magnetic and gravity data were available over the outer highs described in this study. Where magnetic data were available, the presence of magnetic stripe anomalies correlates well with the typical seismic expression of oceanic crust discussed earlier. However, the steep magnetic gradients often associated with the continental-oceanic crustal boundary commonly occur well landward of the outer high. In areas where extensive volcanics are present, the relation of the magnetic patterns to the continental-oceanic crust boundary was variable. In some cases, where volcanics were present, the steep magnetic gradient was seaward of the outer high.

Gravity data commonly show positive isostatic gravity anomalies extending along the interpreted outer edge of the continental margin (Emery *et al*., 1970; Talwani and Eldholm, 1972; Rabinowitz, 1972). Figure 5.8 is an example. A gravity low is commonly present landward of the outer high, indicating a deep sedimentary basin. The assumed shallow basement of the outer high causes a rather strong gravity anomaly but not a large magnetic anomaly (Talwani and Eldholm, 1972).

SUMMARY OF EVOLUTION OF DIVERGENT (PASSIVE) CONTINENTAL MARGINS

If the formation of outer highs during the late rift stage is as prevalent as we believe, it would add another phase to the evolution of divergent (passive) continental margins.

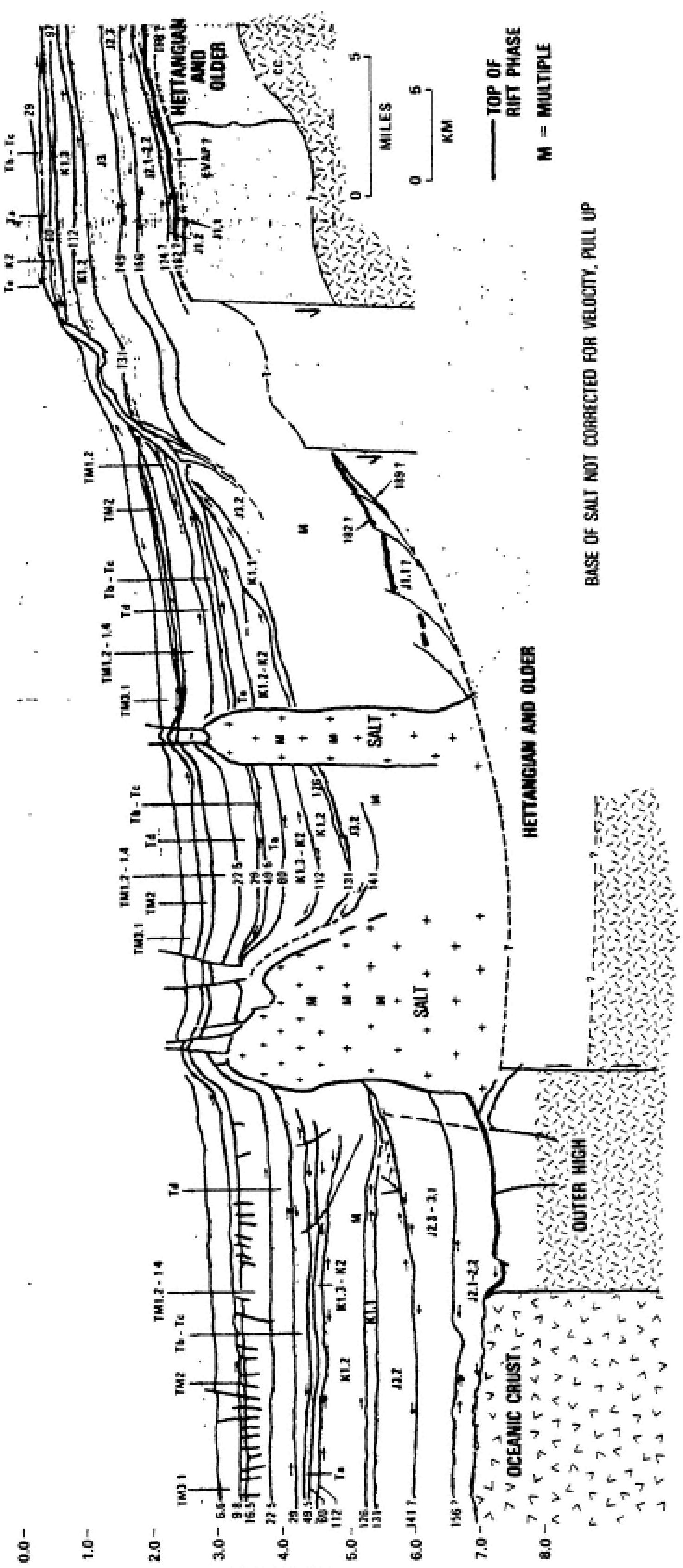


Figure 5.5
Seismic line A, west Africa (see text for explanation).

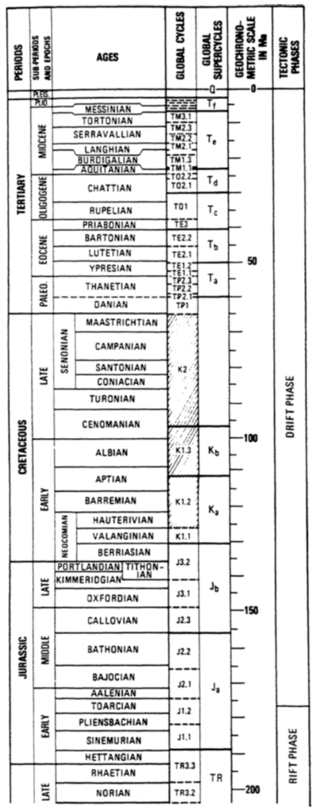


Figure 5.6
Stratigraphic column represented in seismic line A (Figure 5.5).

This section summarizes the evolution of divergent continental margins as a basis for placing the formation of the outer high as an evolutionary phase in divergent continental-margin development.

A vast amount of literature has contributed to the understanding of divergent margins. Just a few publications are cited here: Drake *et al*. (1959); Belmonte *et al*. (1965); Burk (1968); Emery *et al*. (1970); Talwani and Eldholm (1972, 1977); Rabinowitz (1974); Boeuf and Doust (1975); Exon and Willcox (1978).

A summary of the evolution of divergent (passive) continental margins (Figure 5.9) is presented by the IPOD Committee on Passive Continental Margins (Curray, in press). In that paper, the evolution of divergent margins is subdivided into the following phases: (1) variable degrees of doming, (2) rifting, and (3) drifting. Doming of graben margins, such as occurs in the East African rift (Pilger and Rosler, 1976) and the Rhine Graben (Illies, 1977) might not always be present and in some cases may occur while the rifting was taking place (Carey, 1958). Rifting occurs as a single graben or as a complicated subparallel system of grabens and horsts. The drifting is characterized by the formation of oceanic crust, seafloor spreading, and subsidence.

TECTONIC-SEDIMENTARY MODEL FOR THE EVOLUTION OF OUTER HIGHS ON DIVERGENT MARGINS

The tectonic-sedimentary model presented here (Figure 5.10) for the evolution of the outer high differs from the Curray (in press) model for divergent margins by adding a late-rift phase consisting of a longitudinal predrift uplift within the rift graben or graben complex. Our model shows two phases of rifting and two phases of drifting.

The initial rifting phase forms a single graben in this model [Figure 5.10(a)]. However, as discussed in the previous section, the initial rifting may be an intricate graben system. The initial grabening phase may be caused by crustal thinning over a zone several times wider than the graben itself (Rhine Graben Research Group for Explosion Seismology, 1974). Figure 5.11, reproduced from the Rhine Graben Research Group report, shows a cross section of the Rhine Graben. This section is an example of the relationship between early rifting and crustal thinning.

Whether the sediments deposited during the early-rift phase are marine or nonmarine depends primarily on the thickness of the continental crust in which the rift forms. Where the crust is thick, the early rift tends to be topographically high and nonmarine sediments tend to be deposited. Where the crust is thinner, the level of the ripe floor tends to be close to or below sea level, resulting in marine or interbedded marine and nonmarine sediments. Sea-level fluctuations as described in Vail *et al*. (1977) and access to marine waters are important factors in controlling the sediments deposited in these types of rifts.

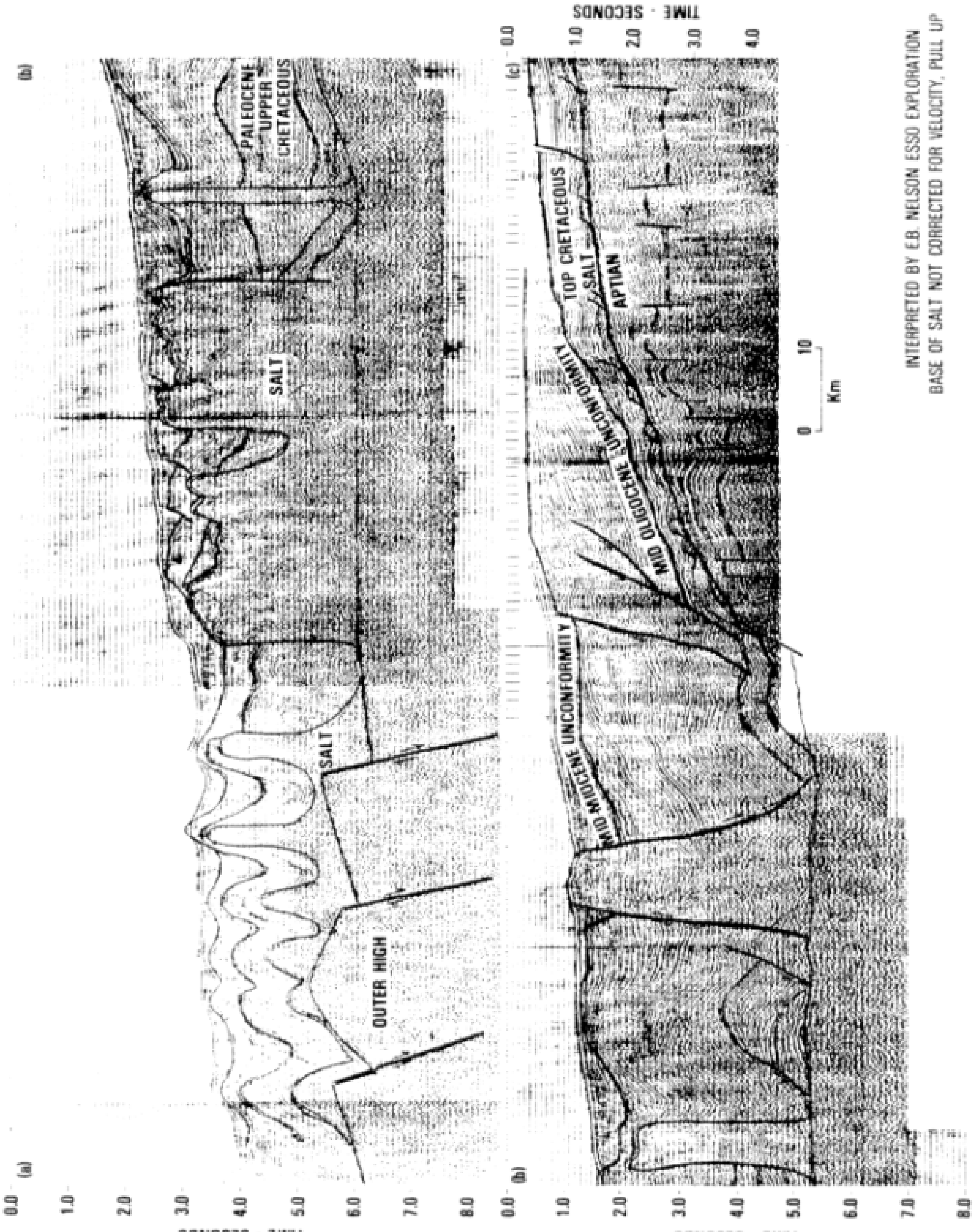


Figure 5.7
Seismic line B, west Africa (see text for explanation).

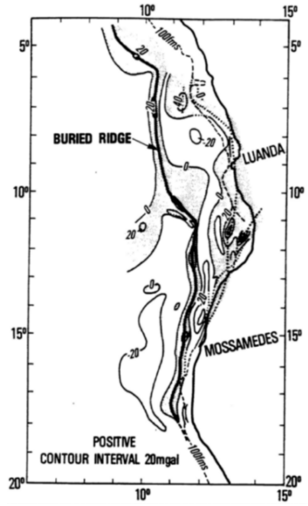


Figure 5.8
Isostatic gravity anomaly map (from Rabinowitz, 1972) of the Angola continental margin showing the outer high as a positive anomaly.

During the late-rifting phase [Figure 5.10(b)], a narrow irregular longitudinal uplift forms within the graben. In cross-sectional view, the narrow uplifted strip has the appearance of a basement high. Also, during the late-rift phase, the area of thinned continental crust on both sides of the basement high continues to subside. Our data indicate that the late-rift phase sedimentary sequences in these two subsiding basins commonly consist of facies varying from moderately deep-water sediments to evaporites. In general, erosion removes the early-rift-phase nonmarine sediments from the uplifted median basement high, and late-rift-phase marine strata onlap it. Our geophysical studies indicate that the outer high is commonly injected by basic intrusions and is in many cases covered by basic extrusives. The upward movement of the median basement high appears to us to be caused by a thermalmechanical event that immediately precedes the formation of oceanic crust The process of the thermal-mechanical event is not well understood and is an area recommended for further study.

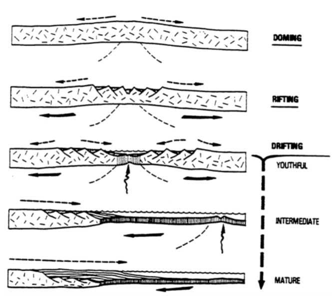


Figure 5.9
Passive margin model from Curray (in press).

At the beginning of the drift phase [Figure 5.10(c)], the outer high splits longitudinally into two outer highs separated by oceanic crust. The outer highs subside rapidly from close to sea level to the present-day depths following the subsidence curves for thinned continental crust and oceanic crust discussed by Parsons and Sclater (1977). Cooling of the mantle below the thinned continental crust and the newly formed oceanic crust and sedimentary

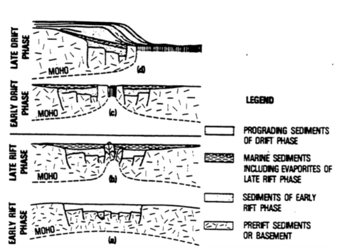


Figure 5.10
Tectonic sedimentary model showing the evolution of outer lights on divergent margins.

loading results in tilting and subsidence of the continental margin. In the proper climate, the rapid subsidence and initially small drainage areas for elastics provide a depositional site ideally suited for the development of thick prograding carbonate banks and reefs. As subsidence slows down through time, the drainage area caused by subsidence expands. This tends to cause terrigenous elastics to build seaward, commonly covering the carbonate bank margins.

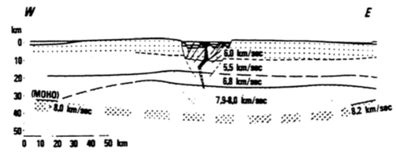


Figure 5.11
Crustal cross section through Rhine Graben (from Rhine Graben Research Group for Explosion Seismology, 1974). Thinning of the continental crust and mantle upwelling occur in a zone several times wider than the width of the graben.

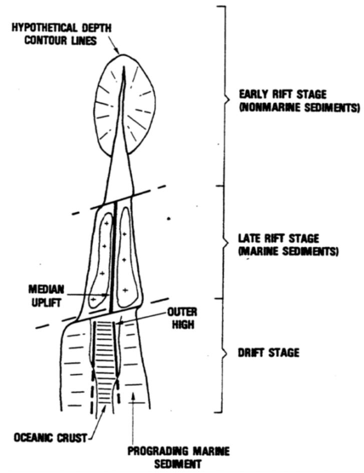


Figure 5.12
Schematic evolution of divergent margin opening in a pivotlike fashion. All stages of evolution. from initial rifting to the final drift stage, may occur simultaneously along a margin opening in a pivotlike fashion. The marine basins of the late-rift stage may be restricted and completely separated from the early-rift and drift phase by the outer high and transform faulting.

Full development of the features associated with rifting is in part dependent on the total extension across the rift. A divergent margin that undergoes a pivotlike opening (Figure 5.12) will fully develop its characteristic features sooner at large distances from the pivot point than at small distances, because large extensions are achieved sooner there. Thus, even though extension along the length of the opening is synchronous, basin formation and filling are diachronous. Also, the early-drift basin may be subdivided into subbasins by highs caused by transform faulting. Subbasins thus formed can restrict the circulation within the late-rift-and early-drift-phase basins, thus causing conditions favorable to the deposition of black shales and/or evaporites.

PROBLEMS AND RECOMMENDATIONS

The combination of regional geological studies and multichannel seismic profiles used in conjunction with other geophysical data has provided the information for the definition of some key problems related to the evolution of outer highs on divergent continental margins. We recommend that a long-term program consisting of geophysical studies, emphasizing regional multichannel seismic lines, and regional geological studies, including outcrop, piston, and dredge samples, be undertaken to locate optimum drill sites for the best possible confirmation of the geophysical interpretations and understanding of the geological problems. Such work is proposed for the 1980's by the JOIDES program (White, 1979). The key geological problems that we have defined concerning the outer highs on divergent continental margins are the following:

1. The character, age, and location of the continental-oceanic crustal boundary;
2. The existence, nature, size variations, and evolution of median highs within grabens to outer highs on divergent continental margins;
3. The nature and evolution of the overthickened oceanic crust and volcanics commonly present in the area adjacent to the continental-oceanic crustal boundary;
4. The type, age, and controls of late-rift-phase sedimentation, particularly organic-rich shales and evaporites;
5. The relation of subsidence and heat flow to crustal type and thickness;
6. The paleoenvironmental history of the sedimentary section through the drift phase including age, paleontology, hiatuses, unconformities, lithofacies, paleobathymetry depth, physical properties, and other significant parameters.

Verification of the model relies on extensive geophysical surveys involving multichannel seismic, magnetic, and gravity data and testing of optimum sites by drilling.

If the model is verified, it will better explain many of the geological problems associated with divergent continental margins, including a better understanding of the nature of the oceanic-continental crustal transition, associated deposition including salt and organic-rich shales, and the relation of volcanism to the onset of seafloor spreading.

SUMMARY

Observation of a large number of geological and seismic profiles across divergent continental margins has demonstrated the widespread occurrence of structural highs that were formed during the late-rift phase of divergent margin evolution and are now the boundary between continental and oceanic crust. This feature originates as a median high, which divides the rift zone into two subsiding basins. This median high later evolves into an outer high along the newly formed continental-oceanic boundary during the drift phase. The outer high commonly extends the whole length of a divergent margin but can be segmented by such factors as transform faults and/or volcanics. In many cases, the outer highs are complicated by extensive intrusives and extrusive volcanics. Early-rift sediments are commonly nonmarine, but in areas of thinned continental crust they may be marine. Marine sediments or volcanics are almost always deposited in the late-rift basins on opposite sides of the median high. Salt and organic-rich shales are commonly deposited during this phase.

This chapter has presented a model for the development of an outer high on divergent continental margins that relates the major structural episodes with the associated sedimentation. It consists of (1) early-rift-phase graben formation, (2) late-rift-phase development of a median high subdividing the graben complex into two subsiding basinal areas, and (3) drift-phase formation of oceanic crust and the outer high as a remnant of the median high and subsidence.

ACKNOWLEDGMENTS

The writers are indebted to E. B. Nelson of Esso Exploration, Inc., and J. Viellart of Esso REP for their seismic interpretations; to J. P. Verdier and R. Lehmann of Esso Production Research—European Laboratories for their paleontological contributions; to H. R. Gould of Esso Production Research Company for his guidance; to C. G. Zinkan, D. R. Seely, R. W. Murphy, J. K. Davidson, D. B. Macurda, K. T. Biddle, R. G. Vierbuchen, R. P. George, Jr., all of Exxon Production Research Company, who reviewed the manuscript; and to Esso Exploration, Inc., for permission to publish the seismic lines.

References

Belmonte, Y., P. Hirtz, and R. Wenger (1965). The salt basins of the Gabon and the Congo (Brazzaville): a tentative paleogeographic interpretation, in *Salt Basins Around Africa*, Elsevier, Amsterdam, pp. 55-74.

Boeuf, M. G., and H. Doust (1975). Structure and development of the southern margin of Australia, *Aust. Petrol. Explor. Assoc. J. 15*, 33-43.

Burk, C. A. (1968). Buried ridges within continental margins, *N.Y. Acad. Sci. Trans. 30*, 135-160.

Carey, S. W. (1958). A tectonic approach to continental drift, in *Continental Drift, a Symposium*, S. W. Carey, ed., Geology Department, University of Tasmania, Hobart, Tasmania, Australia, pp. 177-349.

Curray, J. R. (in press). The IPOD Programme on Passive ContinentalMargins , in *The Evolution of Passive Continental Margins in the Light of Recent Drilling Results*, Philosophical Transactions of the Royal Society, London, England.

Drake, C. L., M. Ewing, and G. H. Sutton (1959). Continental margins and geosynclines: the east coast of North America, north of Cape Hatteras, in *Physics and Chemistry of the Earth*, Vol. 3, L. H. Ahrens, ed., Pergamon, New York, pp. 110-158.

Emery, K. O., E. Uchupi, J. D. Phillips, C. O. Bowin, E. T. Bunce, and J. T. Knott (1970). Continental rise off eastern North America, *Am. Assoc. Petrol. Geol. Bull. 54*, 99-108.

Exon, N. F., and T. B. Willcox (1978). Geology and petroleum potential of Exmouth Plateau and off western Australia, *Am. Assoc. Petrol. Geol. Bull. 641*, 40-72.

Haller, J. (1971). *Geology of the East Greenland Caledonides*, Interscience, New York, 413 pp.

Illies, J. H. (1977). Ancient and recent rifting in the Rhine graben, *Geol. Mijnbouw 56*, 329-350.

Parsons, B., and J. G. Sclater (1977). Analysis of the variation of ocean floor bathymetry and heat flow with age, *J. Geophys, Res. 82*, 803-827.

Pilger, A., and A. Rosler (1976). Temporal relationships in the tectonic evolution of the Afar Depression (Ethiopia) and the adjacent Afro-Arabian rift system, in *Afar Between Continental and Oceanic Rifting*, Interunion Commission on Geodynamics, Rep. No. 16, E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, pp. 1-25.

Rabinowitz, P. D. (1972). Gravity anomalies on the continental margin of Angola, Africa, *J. Geophys. Res. 77*, 6327-6357.

Rabinowitz, P. D. (1974). The boundary between oceanic and continental crust in the western North Atlantic, in *The Geology of Continental Margins*, C. A. Burk and C. L. Drake, eds., Springer-Verlag, New York, pp. 67-84.

Rhine Graben Research Group for Explosion Seismology (1974). The 1972 seismic refraction experiment in the Rhine graben—first results, in *Approach to Taphrogenesis*, J. H. Illies and K. Fuchs, eds., Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, pp. 122-137.

Shipley, T. H., and R. T. Buffler (1978). Seismic stratigraphy and geologic history of Blake Plateau and adjacent western Atlantic continental margin, *Am. Assoc. Petrol. Geol. Bull. 62*, 792-819.

Talwani, M., and O. Eldholm (1972). Continental margin off Norway: a geophysical study, *Geol. Soc. Am. Bull. 83*, 3575-3606.

Talwani, M., and O. Eldholm (1977). Evolution of Norwegian Greenland Sea, *Geol. Soc. Am. Bull. 88*, 563-999.

Vail, P. R., R. M. Mitchum, Jr., R. G. Todd, J. M. Widmier, S. Thompson III, J. B. Sangree, Jr., J. N. Bubb, and W. G. Hatlelid (1977). Seismic stratigraphy and global changes of sea level, in *Seismic Stratigraphy—Applications to Hydrocarbon Exploration,* C. E. Payton, ed., Am. Assoc. Petrol. Geol. Mem. 26, Tulsa, Oklahoma, pp. 49-212.

Vail, P. R., R. M. Mitchum, Jr., T. H. Shipley, and R. T. Burlier (in press). Unconformities of the North Atlantic, in *The Evolution of Passive Continental Margins in the Light of Recent Drilling Results,* Philosophical Transactions of the Royal Society, London, England.

Veevers, J. J. (1977). Paleobathymetry of the crest of spreading ridges related to the age of ocean basins, *Earth Planet, Sci. Lett. 34*, 100-106.

White, S. M. (1979). Deep-sea drilling, *Geotimes*, January 1979, 21-23.

III

INTRAPLATE TECTONICS

6

Tectonics of Noncollisional Regimes

B. Clark Burchfiel
Massachusetts Institute of Technology

INTRODUCTION

An early hypothesis of plate tectonics was that plates moved as rigid pieces of lithosphere and that the relative motion between plates was taken up at narrow zones along their boundaries. Later studies suggested that interaction at plate boundaries could produce deformation, magmatic activity, and metamorphism for a considerable distance from those boundaries. Simply stated, the problems to be addressed now are "what intraplate features and events are the result of plate-boundary interactions, what are the interrelations between these features and events, and what are the processes that cause them." All three types of plate boundaries, transform, subduction, and spreading, show widely distributed intraplate effects. In some cases the boundary can be defined along a narrow zone, but in others it is a broad diffuse zone of plate interaction. This latter type of diffuse or soft boundary is common in which one or both of the plates is composed of continental lithosphere. The interrelations between all the geological features and events at these diffuse boundaries are unclear. The study of intraplate or diffuse-plate boundary features and events within continental crust should be an important focus for any program in crustal dynamics. It is in this setting that most continental crust is formed and modified.

The subduction of oceanic lithosphere is considered a stable process because it is denser than the asthenosphere; thus, subduction could extend over long time periods until it is terminated by changes in relative plate motion or by collisional processes. It is evident from studies of modem plate boundaries that subduction of oceanic lithosphere can cause a variety of structural expressions in an overriding plate. Extension in the overriding plate is expressed by the formation of back-are basins and marginal seas. Oblique subduction can produce strike-slip faulting. Horizontal compression in the overriding plate can lead to the development of complex orogenic belts. All three of these types of tectonic activity in an overriding plate can be coupled in various ways to produce a wide range of complex deformational styles. Even though the geological and geophysical evidence indicates that these types of structural effects exist in over riding plates, the condition that led to different tectonic settings is effectively unknown. This chapter examines the nature and extent of plate-boundary-related effects within continental lithosphere. The focus will be principally on the setting where oceanic lithosphere is being subducted beneath continental lithosphere and the leading edge of the continental lithosphere is under compression.

To demonstrate the presence of compression in the overriding plate, two lines of evidence can be examined: (1) modern intraplate seismicity and (2) geological interpretation of orogenic belts. From the seismic evidence it is reasonably clear that an overriding plate can be under compression. The interpretation of the geological evidence is much more complex and must be examined in detail. Interrelations between all the elements of an orogenic belt and their relation to the plate boundary must be established before a conclusion concerning compression within the continental lithosphere can be established.

SEISMIC EVIDENCE FOR INTRAPLATE COMPRESSION

Parts of the Andes are a modem example of a noncollision orogenic belt. Stauder (1975), from a study of seismicity in the Peruvian Andes, has concluded that the Nazca plate is thrusting under the South American plate beneath central and northern Peru along a surface of shallow (10-15º) dip. Of particular importance to an understanding of Andean orogenesis is Stauder's additional conclusion that the leading (western) edge of the South American plate is under east-west horizontal compression for a distance of approximately 700 km east of the Peru Trench (Figure 6.1). Numerous hypocenters are present in that plate at depths of 1-90 km and for distances up to 700 km east of the trench. Most lie at distances from the trench of 200-600 km, depending on latitude. Focal mechanisms for nine of ten of these intraplate earthquakes yield horizontal compressive stress axes oriented approximately east-west The tenth shows normal faulting and is near the trench. Of the nine solutions that show horizontal compressive stress axes, three solutions indicate predominantly strike-slip faulting and six indicate reverse dip-slip faulting.

The Andes, a chain characterized by voluminous magmatic activity, are bounded on the west by the Peru-Chile trench, a zone of ongoing plate convergence, and on the east by an active fold and thrust belt, which separates orogen from craton. Rates of subduction are on the order of 7-12 cm per year and can reasonably be extrapolated back at least into the Late Miocene on the basis of marine magnetic anomaly studies. Since Middle Miocene time the central Andes, including Peru, have experienced intense igneous activity and accompanying orogenesis. Folds, reverse faults, and thrust faults with a relative eastward sense of motion have developed along the eastern margin of the Andes in the sub-Andean zone (Audeband *et al*., 1973). The Andes were uplifted to their present height during this late Cenozoic episode of orogenesis. Thus one interpretation is that the Andean orogenic belt is the result of deformation related to horizontal compression in a noncollisional subduction system.

Stauder's (1975) data axe preliminary, and studies of this type need to be considerably expanded to establish the state of stress within the leading edge of the plate and its relation to areas of modem magmatism and deformation. Even though horizontal compression can be established, how the stress field is generated is unknown. Studies directed toward understanding the origin of the stress fields within the continental crust at convergent boundaries should be encouraged.

GEOLOGICAL EVIDENCE FOR INTRAPLATE DEFORMATION RELATED TO NONCOLLISIONAL PLATE CONVERGENCE

The geological evidence for compression with an overriding continental plate is complex and less clear than the seismic evidence. The Mesozoic Cordilleran orogenic belt of the western United States is an older and more deeply eroded Andean-type mountain belt (Hamilton, 1969; Burchfiel and Davis, 1972, 1975). South of the latitude of central Oregon, the geological history of the Cordilleran orogenic belt suggests that the eastward subduction of an oceanic plate beneath the North American plate occurred contemporaneously with deformation and magmatic activity that at times extended more than 1000 km into the North American plate. Evidence suggests that major collisional events were rare or nonexistent along this part of the plate boundary during Mesozoic time.

The Mesozoic Cordilleran orogenic belt can be divided into four terranes: (1) a western terrane of accreted oceanic rocks, (2) a central magmatic are or superposed arcs, (3) an eastern terrane of east-directed thrust faults and related structures, and (4) a locally developed terrane lying east of the thrust belt of thrusted Precambrian basement rocks in the Colorado-Wyoming area (Figure 6.2). The first three terranes shift spatially with time, but events in each terrane have a crude contemporaneity (Figure 6.2). The fourth terrane is restricted in time, latest Cretaceous to earliest Tertiary, to a period when the Cordilleran orogenic belt underwent significant but poorly understood changes. Even though the geological features and events in these four belts are broadly contemporaneous, their interrelations and their relation to plate-boundary interaction are not well established, particularly for the terranes farther removed from the plate boundary.

In earliest Mesozoic time, the western boundary of the North American plate lay along a line from central Oregon through the central Klamath Mountains, central Sierra Nevada and west of the San Gabriel Mountains into northeastern Baja California (Figure 6.2). The plate boundary is at present very sinuous, which is a result of post-

Mesozoic deformation. Removing later deformation would straighten the boundary considerably. Nearly all the rocks west of this line are Mesozoic in age and represent rocks of oceanic origin that were accreted by subduction processes and partially or wholly reworked to form either transitional or continental crust.

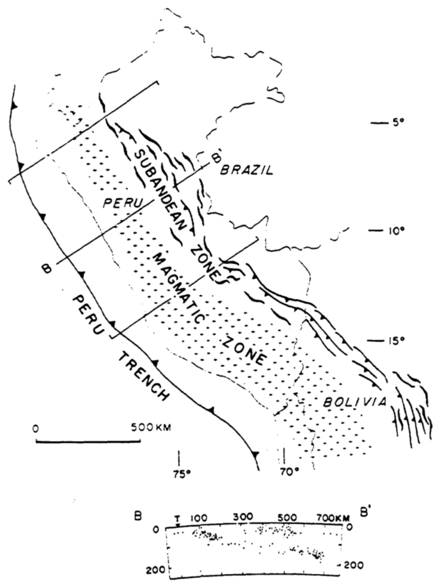


Figure 6.1
Major structural elements of the Peruvian Andes. The belts of Miocene to recent magmatic activity and deformation in the sub-Andean zone are within the South American plate and lie east of the Peru trench, which marks the site of subduction of the Nazca plate. Section B-B' incorporates selected hypo-centers within the outlined region (from Barazangi and Isacks, 1976). T shows the location of the trench. Hypocenters define a shallow dipping subduction zone and distribution of intraplate seismicity within the South American plate.

Western Terrane

Geological data from the western terrane indicate that accretion of oceanic rocks to the North American plate began in Triassic time (Davis *et al*., 1978). Triassic cherts locally associated with late Paleozoic and possible Triassic ophiolites were tectonically disrupted and em-placed by subduction processes probably during Middle Triassic to Early Jurassic time (Figure 6.3). Locally these rocks are associated with blueschist metamorphic mineral assemblages, which were formed 220-210 million years (m.y.) ago. In parts of these accreted sequences, exotic Permian "Tethyan" fusulinid faunas are present in some limestones that are mixed with the Triassic and older rocks suggesting that far-traveled oceanic rocks were incorporated into the accretionary wedge and reworked by later deformational events. Jurassic ophiolites, associated sedimentary rocks, and volcanic rocks lie above and to the west of the rocks accreted in earlier Mesozoic time and represent new additions from oceanic lithosphere to the North American plate. Most of the Triassic and Jurassic rocks can be interpreted as accreted by eastward subduction of oceanic lithosphere beneath the North American plate. Some workers have argued that one or more arc collisions may have occurred during Jurassic, particularly in Late Jurassic time (Schweickert and Cowen, 1975). The evidence is equivocal that arc-continent collisions took place at this time, and considerably more work is needed to clarify this question. Even if an are collision did occur in the Late Jurassic, eastward underthrusting of oceanic lithosphere dominated Triassic and Jurassic plate-boundary activity. Cretaceous and early Tertiary rocks form the western part of the western terrane and were accreted to the North American plate during east-directed subduction of Cretaceous and early Tertiary time (Figure 6.4). Like all accreted terranes, it is uncertain whether subduction was continuous or episodic. Blueschists of Cretaceous age are common within these rocks. Geological data demonstrate that rocks of the western terrane became younger toward the west, indicating retrogression of the subduction boundary during Mesozoic time (Davis *et al*., 1978).

Central or Magmatic-Arc Terrane

The central terrane consists of a magmatic arc or superposed arcs of Mesozoic age. Arc plutons intrude Precambrian crystalline rocks in the south and Paleozoic arc rocks belonging to an arc accreted in latest Paleozoic to earliest Triassic time in the north. All host rocks for the plutons were part of the North American plate; thus the arc was built on the western edge of the North American plate, and its structural setting was similar to the modem Andes. The Mesozoic volcanic-plutonic are began to develop in Early Triassic time, and the oldest plutonic rocks date to approximately 230-240 m.y. ago (Figure 6.3). Igneous activity occurred throughout most of Mesozoic time, but the degree of continuity of activity is controversial. Several workers have discussed this problem (Lanphere and Reed, 1973; Armstrong and Suppe, 1973), and the data suggest at least three intrusive epochs: (1) 79-106 m.y. ago; (2) 132-158 m.y. ago; and (3) an undefined epoch older than 160 m.y. ago, with the oldest dates at 230-240 m.y. ago. Because a few concordant age pairs fall in the intervals between those epochs, the alternative hypothesis of continuous magmatism cannot be eliminated. Younger igneous rocks are also present in the western United States, but an important change in magmatic and structural events occurred at about 75 m.y. ago as discussed below. Magmatic activity in the arc terrane is broadly over the same time span as subduction activity in the western terrane, which has led many workers to couple the two terranes into an Andean-type are-trench system (Hamilton, 1969; Burchfiel and Davis, 1972).

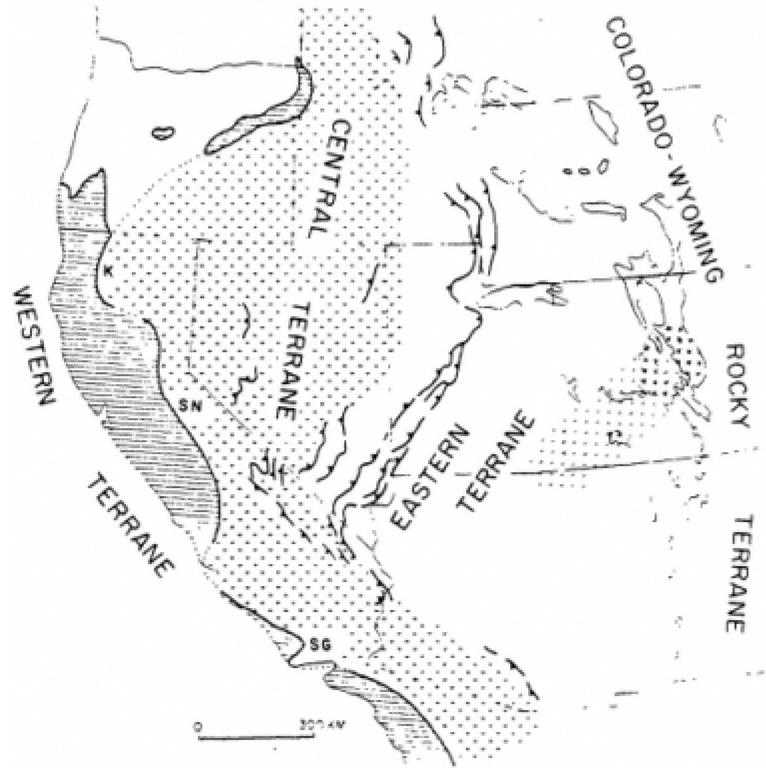


Figure 6.2
The four Mesozoic to early Tertiary terranes of the Cordilleran orogenic belt of the western United States. The western terrane consists of rocks accreted to the North American plate during Mesozoic time, the central terrane is formed by rocks of one or more magmatic arcs, and the eastern terrane consists of an east-directed fold and thrust belt. The Colorado-Wyoming Rocky Mountain terrane developed in latest Cretaceous-early Tertiary time during important rearrangement of Cordilleran tectonic elements. The line marking the eastern edge of the western terrane is the approximate western edge of the North American plate at the beginning of the Mesozoic. K, Klamath Mountains; SN, Sierra Nevada; SG, San Gabriel Mountains. No attempt has been made to remove intraplate deformation such as in the Basin and Range province except for reversing movement on the San Andreas and some associated faults.

Eastern Terrane

The eastern terrane lies east of the magmatic arc and consists of relatively east-directed thrust faults and associated folds. In early Mesozoic time in western Nevada thrust faults developed within Paleozoic "eugeosynclinal" rocks and early Mesozoic back. arc sedimentary and volcanic rocks (Figure 6.3). The age of these thrust faults is Early and Middle Jurassic, and post-Middle Jurassic and pre-Middle Cretaceous. Recent work has demonstrated the presence of early Mesozoic thrusting and folding in northeastern Nevada, which could range in age from Middle Triassic to Early Jurassic. In the miogeosyncline on either side of the California-south-central Nevada state line are several large thrust faults and associated folds that involve rocks as young as Early and Middle Triassic and are cut by plutons 185 m.y. old. These thrusts belong to an early Mesozoic period of thrusting that may he earlier than or synchronous with deformation in western and northeastern Nevada. Farther southeast in southeastern California, thrusts are cut by plutons 200 m.y. old and a newly discovered thrust that is unconformably overlapped by the Upper Triassic (?)-Lower Jurassic Aztec Sandstone. Thrust faults in southeastern California involve Paleozoic miogeosynclinal and cratonal facies and their Precambrian crystalline basement. Recent data from southern California have demonstrated that some deformational events are older than 230-240 m.y. Whether these various events belong to one or more episodes of deformation is not yet known. At present we group them in an undifferentiated period of early Mesozoic deformation, which ranges from Middle Triassic to Middle Jurassic.

During late Mesozoic time, are magmatism encroached eastward into the region of early Mesozoic deformation, with the thrust and fold belt involving rocks even further east (Figure 6.4), except in southeastern California where early and late Mesozoic deformation is superposed. The thrust faults developed from west to east until rock units transitional between miogeosyncline and craton were involved in thrusting. In southeastern California, thrust faults strike south and southeast, leaving the Paleozoic geosynclinal terrane and cutting through the craton. Nearly all late Mesozoic thrust faults in this region involve Precambrian crystalline rocks and strike parallel to and along the eastern edge of the late Mesozoic magmatic arc. This major change in structural trend and style occurs near the California-Nevada state line and presumably continues into Mexico, although data to the southeast are scanty.

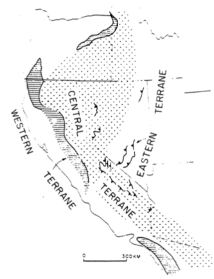


Figure 6.3
Spatial distribution of the western, central, and eastern terranes from Early Triassic to Late Jurassic time. The western and eastern terranes were probably continuous on both sides of the central terrane, but only their present outcrop distribution is shown.

Late Mesozoic plutonism occurred farther eastward in Nevada and to the north than in early Mesozoic time, but in southeastern California and to the south early and late Mesozoic igneous activity is superposed. Associated with the Mesozoic plutonic and volcanic rocks, but characteristically lying east of their major areas of development, thus transitional between the central and eastern terranes, are numerous isolated areas of metamorphic rocks that probably represent exposed culminations of a metamorphic belt that may be continuous at depth. Metamorphism of amphibolite grade affects miogeosynclinal and cratonal rocks of Precambrian and Paleozoic age as well as Mesozoic and early Tertiary (?) granitic, volcanic, and sedimentary rocks. Meager age data suggest that metamorphism began at least 180 m.y. ago and continued into the Tertiary, but it is not known if this represents one long period of metamorphism or several spatially and temporally distinct periods. In some areas of southeastern California and adjacent Arizona, metamorphism is mid-Cenozoic in age and is associated with extensional tectonics, not thrust faulting (see Chapters 8 and 9). It is likely that metamorphism is synchronous with magmatism, spans a long period of time, and will be tied ultimately to magmatic epochs.

The eastern or frontal thrust belt of the Cordilleran orogen lies to the east of the Mesozoic metamorphic terranes mentioned above. In this belt, east-directed thrust plates of Late Jurassic to Late Cretaceous age define a zone of crustal dislocation that extends from British Columbia and Alberta, Canada, to southeastern California and probably into Sonora, Mexico. During the Late Cretaceous an important change took place in the location of deformation by thrust faulting. Formation of folds and thrusts in the eastern thrust belt ceased before the end of the Cretaceous in a sector from central Utah to southeastern California (Armstrong, 1968). North and south of this sector of the thrust belt, deformation continued through the Late Cretaceous into the early Tertiary and ceased in Middle

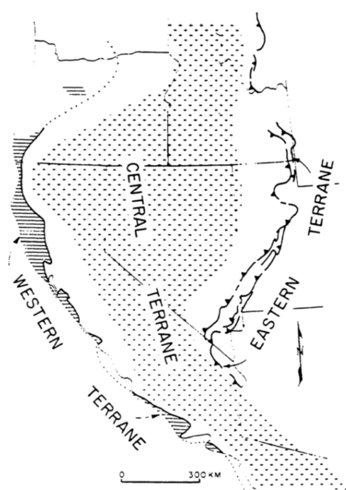


Figure 6.4
Spatial distribution of the western, central, and eastern terranes from Late Jurassic to latest Cretaceous time. The western and eastern terranes were probably continuous on both sides of the central terrane, but only their present outcrop distribution is shown.

or Late Eocene time. The period from latest Cretaceous to Eocene (75-50 m.y. ago) is the time of the classic Laramide orogeny. Low-angle thrust faults developed during Laramide deformation have a structural style similar to those developed in the earlier Cretaceous events but generally lie somewhat farther east.

Colorado-Wyoming Rocky Mountain Terrane

The fourth terrane is only locally developed both spatially and temporally and consists of large uplifts of Precambrian crystalline rocks that extend from southern Montana into New Mexico. These uplifts, commonly referred to as the Colorado-Wyoming Rocky Mountains, developed only during the Laramide Orogeny of latest Cretaceous and early Tertiary time. The structural geometry of the faults bounding the uplifts has been controversial, one group postulating thrust faults that steepen with depth, a second group postulating that the boundary faults are moderate to low angle (for a review see Sales, 1968; Stearns, 1971). Recent seismic reflection lines across one of the uplifts, the Wind River Range, by the Consortium on Continental Reflection Profiling (COCORP), has demonstrated a gently dipping thrust fault bounds the uplift (see Chapter 10). Some of the uplifts can be interpreted to be bounded by thrust faults, but several are still equivocal. The geometry of these uplifts is critical because their origin has been ascribed to horizontal compression or vertical displacement without horizontal crustal shortening. The COCORP data suggest that the former interpretation may be correct, in which case it may be possible to relate the structural origin of this terrane to plate-boundary interaction even though the boundary is more than 1000 km to the west.

Igneous activity in the western part of the North American plate also underwent a significant change during the Late Cretaceous (Figure 6.5). During most of Mesozoic time, a plutonic-volcanic Andean-type arc was active as discussed above. Although the problem of the continuous or episodic character of magmatism has not yet been resolved satisfactorily, it appears that during most of Mesozoic time magmatism was characteristic of much of the western part of the North American plate in the United States and southern Canada. About 75 m.y. ago a change occurred in the pattern of igneous activity. Igneous rocks intruded during Laramide time (75-50 m.y. ago) are present in Canada and Idaho and in southeastern California and Arizona (and presumably western Mexico), but they are very rare in central and northern California, Nevada,

Figure 6.5
Spatial distribution of the western, central, eastern, and Colorado-Wyoming Rocky Mountain terranes from latest Cretaceous to early Tertiary time. The western terrane was probably more extensive than its present outcrop distribution and may have been continuous along the entire west coast, but it is now largely in the offshore region.

and western Utah (Armstrong, 1974; Burchfiel and Davis, 1975). Igneous rocks of Laramide age are also present further east in southwestern Montana (Adel Mountain, Elkhorn Mountain, and Livingston volcanic rocks), central and southwestern Colorado (Colorado Mineral Belt), northeast and southern Arizona, and southwestern New Mexico (Figure 6.5). These latter areas of igneous activity lie east of areas of earlier Cretaceous magmatism in a region that had lacked igneous activity since the Precambrian. The Laramide igneous rocks of Colorado and parts of Arizona lie east of the area where Laramide magmatic activity of the main batholith trend of the Cordillera ceased (Figure 6.5). Of considerable importance is the general spatial coincidence of this Laramide igneous activity with the region of basement uplifts.

In summarizing the areal distribution of deformation and magmatism during Late Cretaceous-Early Tertiary time, the following data are important: (1) low-angle thrust faults developed from Canada to Mexico along a continuous belt of Late Jurassic to Late Cretaceous age; (2) the belt of low-angle thrust faults lies along the eastern margin of a terrane or zone of batholiths that were em-placed synchronously with thrusting (230-75 m.y. ago); (3) low-angle thrusting ended in Late Cretaceous time in a sector from central Utah to southeastern California at approximately the same time as magmatism ceased in corresponding latitudes in the batholith belt to the west; (4) to the north and south, Laramide low-angle thrusting continued in regions where magmatism persisted in the main plutonic belt from 75 m.y. to approximately 50 m.y. ago; and (5) in the central Cordillera, largely adjacent to the gap in Laramide magmatism of the main batholith belt, Laramide igneous activity shifted eastward and coincides generally with the region of Rocky Mountains basement uplifts of the same age. These data suggest that the development of the Laramide structures in the Colorado-Wyoming Rocky Mountains is related to events that took place throughout the orogenic belt and are ultimately related to plate-boundary interaction.

PROBLEMS

The geological data from the four terranes within that part of the North American plate considered here demonstrate an approximate contemporaneity for Mesozoic events. Studies of modem plate boundaries clearly indicate that plate motions and plate-boundary geometries can change rapidly, on the order of a few million years. Our dating accuracy of events in the Mesozoic are at best within several million years. Stratigraphic units and deformational, magmatic, and metamorphic events require precise dating in each terrane before their effects can be correlated across the orogenic belt, so that a complete picture of all synchronous events at and related to the convergent plate boundary can be developed. Each of the four terranes presents different geological problems that require solution before a complete picture of the terrane evolves and before adequate correlation between terranes and the convergent boundary can be established.

The western terrane is characterized by the presence of ophiolites, which offers geologists an exposure of oceanic crust and its deep marine sediments. Study of ophiolites has shown that they have great variability in their internal structure and composition. Continued study of these rocks is necessary to understand their genesis, tectonic setting, and mode of incorporation into a subduction complex. Accurate dating of ophiolite sequences is necessary to establish their time of formation. Recent breakthroughs in the preparation of and study of radiolarians have made great advances possible in the dating of not only ophiolite formation but the sedimentary record of deep-ocean sedimentary sequences. Subduction is not an instantaneous process but may continue for long periods of geological time; however, our techniques for unraveling sequential events in an accretionary wedge need special attention. Tectonic styles and processes are still poorly understood and must be examined in light of modem subduction systems. Blueschist metamorphism is associated with subduction processes, but many problems still remain in understanding blueschist formation and preservation in subduction terranes. Island-arc-type volcanic rocks are sometimes associated within accretionary wedges and incorporated into them by later plate-boundary activity. Geological studies aimed at determining the ancient and modem paleogeographic settings of island-arc sequences and how they become incorporated into accreted boundaries still require a great deal more effort. All of these studies should be correlated with comparison of geophysical and geological studies of modem subduction systems.

Study of the central or magmatic are terrane still must have as its primary objectives the origin of the magmas and their temporal and spatial relations. There remains the principal problem of how magma generation is linked to subduction, and what are the influences of the upper-mantle and continental crust on magma composition. It is within the realm of the magmatic are that pre-existing rock sequences are reworked to form continental crust. During this activity, continental crust, as it is known seismically, begins to develop. Through geological, geochemical, and geophysical study of ancient arc systems such as the Mesozoic are of the western United States, correlated with similar studies of active arc terranes, a more comprehensive understanding of these terranes will evolve. It is of great importance, because vast quantities of our mineral wealth are related to magmatic are development. A proper geological framework of magmatic arcs will greatly aid our exploration of their mineral reserves.

The thrusted and folded terrane east of the magmatic are locally has been mapped in considerable detail, but numerous problems remain concerning the geometry and origin of its structure. East-directed thrust faults dominate the structures of this terrane, and the geometry of the thrust faults is reasonably well understood locally, principally because of seismic profiling by oil companies. The origin of these thrust faults remains a great problem. It is still unclear what the relation of thrust faults that involve only sedimentary rocks is to metamorphic core areas that lie to the west of them. Geologists have been preoccupied with the eastern or frontal parts of thrust faults and have neglected the western or rear parts of these structures. Until we know the relation between metamorphic core and basement rocks to the thrust faults at their rear in the source terranes, adequate mechanical models for thrust faulting cannot be made. Evidence suggests that thrust faults of the eastern terrane may be related to crustal shortening within the leading edge of an overriding continental plate, but more precise geological and time relations are necessary to establish the correct three-dimensional models. The attack must be an integrated study of the metamorphic, structural, and geochronological history of the entire thrusted and folded terrane. More geophysical studies, particularly reflection and refraction studies, in complex areas of the rearward parts of the thrusted terranes are necessary to give a better three-dimensional picture of crustal and structural relations.

The fourth terrane of thrusted Precambrian basement rocks of the Colorado-Wyoming area is of considerable economic interest because of its relation to accumulations of oil and gas and mineral deposits of Precambrian, Mesozoic, and Tertiary age. The Colorado-Wyoming structural province has been regarded by some workers as unique; however, many similar structural provinces are known throughout the world, such as the Amadeus Basin of central Australia and the Caucasus Mountains of southern Russia. Four major problems can be defined in this terrane: (1) Why is it localized where it is? (2) What is its structural configuration with depth? (3) What is the relation and origin of the limited contemporaneous magmatic activity to this province? (4) Is it related to the convergent margin nearly 1000 km farther west?

Within the four terranes of the Mesozoic Cordilleran orogen, numerous problems of timing of events and structural relations exist; thus timing and structural relations between terranes and within the orogen as a whole are not well understood. Only broad correlations between events can be made at present. Greater detail in establishing contemporaneity of events needs to be compared, and relations between subduction, magmatic, metamorphic, and back-arc structures need to be established. Present data suggest that events in all terranes are contemporaneous and all are related to activity at a convergent noncollisional plate boundary: the effects of convergence extend 1000 km from the plate boundary. Greater understanding of the interrelations across the orogenic belt is necessary before accurate models can be developed. At present, a consistent model can be constructed that involves compression in an overriding continental plate in response to subduction of oceanic lithosphere. The nature of lower crust and mantle involvement is unknown, and only through the study of modern belts such as the Andes can we hope to understand the deeper workings of such a plate boundary. The plate boundary and its effects within the overriding plate should be regarded as a single dynamic system. The processes involved in the development of the Mesozoic Cordilleran orogen are processes that lead to the formation and evolution of continental crust.

References

Armstrong, R. L. (1968). Sevier orogenic belt in Nevada and Utah, *Geol. Soc. Am. Bull. 79*, 429-458.

Armstrong, R. L. (1974). Magmatism, orogenic timing and orogenic diachronism in the Cordillera from Mexico to Canada. *Nature 247*, 348-351.

Armstrong, R. L., and J. E. Suppe (1973). Potassium-argon geochronology of Mesozoic igneous rocks in Nevada, Utah and Southern California, *Geol. Soc. Am. Bull. 84*, 1375-1392.

Audebaud, E. H., R. Capdevila, B. Dalmayrac, J. Debelmas, G. Laubacer, C. Lefevre, R. Marocco, C. Martinez, M. Mattauer, F. Megard, J. Paredes, and P. Tomasi (1973). Les traits géologique essentiels des Andes central (Perou-Bolivie), *Rev. Geogr. Phys. Geol. Dynam. XV*, 73-114.

Barazangi, M., and B. L. Isacks (1976). Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America, *Geology 4*, 686-692.

Burchfiel, B. C., and G. A. Davis (1972). Structural framework and evolution of the southern part of the Cordilleran orogen, western United States, *Am. J. Sci. 272*, 97-118.

Burchfiel, B. C., and G. A. Davis (1975). Nature and controls of Cordilleran orogenesis, western United States: extensions of an earlier synthesis, *Am. J. Sci. 275-A*, 363-396.

Davis, G. A., J. W. H. Monger, and B. C. Burchfiel (1978). Mesozoic construction of the Cordilleran "collage," central British Columbia to central California, in *Pacific Coast Paleogeography Symposium 2*, Pacific Section, Society of Economic Paleontologists and Minerologists, pp. 1-32.

Hamilton, W. (1969). Mesozoic California and the underflow of Pacific mantle, *Geol. Soc. Am. Bull. 80*, 2409-2430.

Lanphere, M. A., and B. L. Reed (1973). Timing of Mesozoic and Cenozoicplutonic events in circum-Pacific North America , *Geol. Soc. Am. Bull. 84*, 3773-3782.

Sales, J. K. (1968). Cordilleran foreland deformation, Am. Assoc. *Petrol. Geol. Bull. 52*, 2016-2044.

Schweickert, R. A., and D. S. Cowan (1975). Early Mesozoic tectonic evolution of the western Sierra Nevada, California, *Geol. Soc. Am. Bull. 86*, 1329-1336.

Stauder, W. (1975). Subduction of the Nazca plate under Peru as evidenced by focal mechanisms and by seismicity, *J. Geophys. Res. 80*, 1053-1064.

Stearns, D. W. (1971). Mechanisms of drape folding in the Wyoming province, *Twenty-Third Annual Field Conference Guidebook*, Wyoming Geological Association, pp. 125-143.

7

Models for Midcontinent Tectonism

William J. Hinze and Lawrence W. Braile
Purdue University
G Randy Keller
University of Texas, El Paso
Edward G. Lidiak
University of Pittsburgh

INTRODUCTION

The midcontinent region of the United States has long been regarded as part of the stable craton. Geological evidence has led to the assumption that this area has undergone only minor tectonism during the past several hundred million years and that this tectonism has largely taken the form of broad, slow, vertical movements. However, during the past decade there has been accumulating geological evidence and increasing awareness that the midcontinent region has been and is at present tectonically active. This change in geological thought has come about because of studies of earthquake activity and enhanced discrimination of lateral crustal variations by geophysical techniques.

Earthquake activity has focused attention on the central midcontinent, in the vicinity of the New Madrid seismic zone at the head of the Mississippi Embayment, and has encouraged studies of contemporary tectonics as a means of predicting seismicity and the areal limits of the potential seismic activity. This chapter reviews the major published tectonic hypotheses for the contemporary geodynamics of the midcontinent region. However, to set the framework for these hypotheses, the geological history is summarized with emphasis on the structural development and related tectonic events. This summary is important because the tectonic events that have acted upon the midcontinent in the past are only interpretable based on the structural, sedimentary, and thermal events reflected in the geological history of the area and nearby plate margins. If geological events involve orderly processes, then we can anticipate that the past in a general way is a clue to the present. Although noncyclic processes are important in early history and crustal conditions have changed to some degree, previous tectonism provides guidelines for subsequent dynamic processes (Allen, 1975). It is important to use this structural knowledge to decipher contemporary tectonic processes.

GEOLOGICAL HISTORY

The geological history of the midcontinent region has been the subject of many discussions (e.g., Bristol and Buschbach, 1971), and the major tectonic events are shown schematically as a function of time in Figure 7.1. The early history of the area is poorly known because

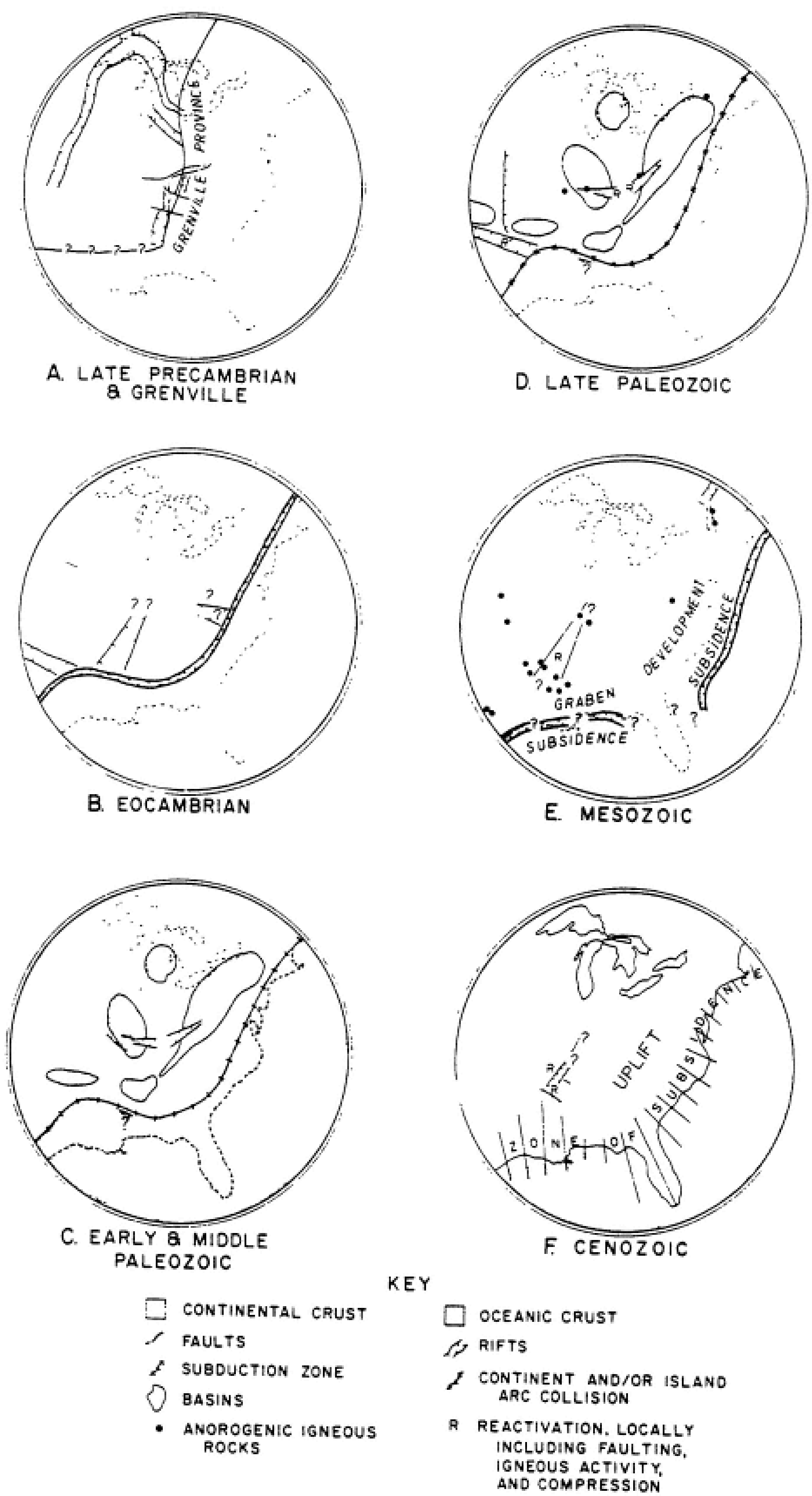


Figure 7.1
Schematic diagrams of major tectonic events that have affected the midcontinent region of the United States through geological time.

Precambrian rocks are generally buried beneath Phanerozoic sedimentary rocks. The Precambrian events have most recently been summarized by Denison *et al*. (in press). The Precambrian rocks of the continental interior are characterized by linear orogenic belts prior to about 1.6 billion years (b.y.) ago. Continental stabilization occurred subsequently and is indicated by shelf-type sedimentation, anorogenic igneous activity, continental rifting, and epeirogenic deformation. Orogenic activity continued along the eastern and southern continental margin in late Precambrian and Phanerozoic time.

Figure 7.1A shows the Grenville basement rock province, which is bordered on the west by an anorogenic igneous terrane and a series of basaltic rift zones. The anorogenic terrane consists mainly of intrusive and extrusive rocks of felsic composition dated at about 1.3-1.5 b.y. old. The basaltic rift zones extend from Kansas north to Lake Superior and south into Michigan and perhaps Indiana, Ohio, Kentucky, and Tennessee. Basaltic igneous activity in these rifts has been dated in the Lake Superior region at 1.0-1.2 b.y. ago (Goldich, 1968). These rifts may reflect a thermotectonic event that attempted to disrupt central North America—an event that is at least partly contemporaneous with tectonic activity occurring within the Grenville province. The third Precambrian terrane is the subsurface extension of the Grenville province of Canada. This province is underlain by medium-to high-grade metamorphic rocks, granites, and anorthosites. The last major period of metamorphism occurred 1.0 b.y. to 1.1 b.y. ago, an event that has been related to continental collision during closing of a Precambrian proto-Atlantic Ocean (Baer, 1976). These events were followed by uplift and stabilization of the Grenville Front, whose extension into the southeastern United States has not been defined.

By latest Precambrian to Eocambrian time, the ancient continental land mass of eastern North America began to split (Figure 7.lB). Each of the continental margins formed was probably of the Atlantic type. The grabens associated with the initial rifting were filled with clastic sedimentary and volcanic rocks. The volcanic rocks of the Catoctin Formation (820 m.y. old) probably represent this rifting event (Brown, 1970). Development of aulacogens also occurred on the craton during this period.

The proto-Atlantic or Iapetus Ocean (Rankin, 1976) continued to grow during early Paleozoic time and by Ordovician time had reached substantial width (Figure 7.1C). The continental margin was transformed into an Andean-type margin along the entire eastern continental margin and probably extended along the southern margin into the Ouachitas. This subduction is indicated by tectonic and thermal events associated with the Taconic orogeny, which affected an area from Newfoundland to Alabama. Widespread cratonic sedimentation occurred in the midcontinent during early and middle Paleozoic time (Figure 7.1C), and the development of numerous basins and arches (Figure 7.2) was initiated. Locally, thick deposits accumulated in the major basins, perhaps as a result of contemporaneous faulting and regional epeirogenic movements. The arch-basin relationship continued to develop throughout the Paleozoic. The axes of the arches and the centers of the basins shifted through time, but the arches were rarely above sea level. Deposition in basins generally kept pace with subsidence, resulting in predominantly shallow-water marine deposits.

By Mid-Devonian to Early Mississippian time (Figure 7.1D) plate consumption was terminated in the northern Appalachians by a collision of continental plates in the Acadian orogeny. During this time, the central and southern Appalachians were involved in continued subduction, and thermal activity and significant tectonic activity was under way in the Ouachitas. Activity in these areas reached a peak during Pennsylvanian time and is reflected in the Alleghenian-Ouachita orogeny, which also involved continental collision.

Mesozoic time (Figure 7.1E) is marked by the rupturing of Pangaea and the opening of the present Atlantic Ocean. Graben development and subsidence occurred along the eastern and southern margins of North America. The Reelfoot Rift of the Mississippi Embayment was reactivated during this time, and this reactivation continued into the Cenozoic era (Figure 7.1F).

CONTEMPORARY GEODYNAMICS

The above discussion of geological history sets the framework for consideration of the contemporary geodynamics of the midcontinent region. These geodynamic processes are inadequately known and poorly understood at present. However, the reality of tectonic activity is dramatically illustrated by the recent faulting and earthquake activity in the New Madrid seismic zone. Evidence for this activity is largely indirect, as surface indications are rare because of an overburden of recent sedimentary deposits. Also, evidence is limited by the short (~200 year) duration of the historical record. However, observations of current patterns of seismicity, stress distribution, Cenozoic faulting, and vertical movements are relevant and are therefore summarized below.

Seismicity

A review of seismicity studies of the midcontinent (e.g., Nuttli, 1974) suggests the following major conclusions: (1) The pattern of seismicity is rather diffuse (Figure 7.3), but there are zones in which activity is concentrated (the New Madrid seismic zone, southern Appalachians, St. Lawrence Valley, and the Boston-Ottawa trend). (2) The most intense historical earthquake activity has been centered in the New Madrid seismic zone in southeast Missouri and adjacent areas (Figure 7.4). (3) The historical earthquake record of the New Madrid seismic zone is dominated by the 1811-1812 earthquake sequence and associated aftershocks, which represent a major earthquake occurrence on any scale of comparison. Nuttli (1974) has shown that the three major shocks of the 1811-1812 se quence had approximate magnitudes (m_b) of 7.2, 7.1, and 7.4. (4) The seismicity of the eastern United States is considerably less than the west, but for a given earthquake magnitude an earthquake is felt over a larger area in the east.

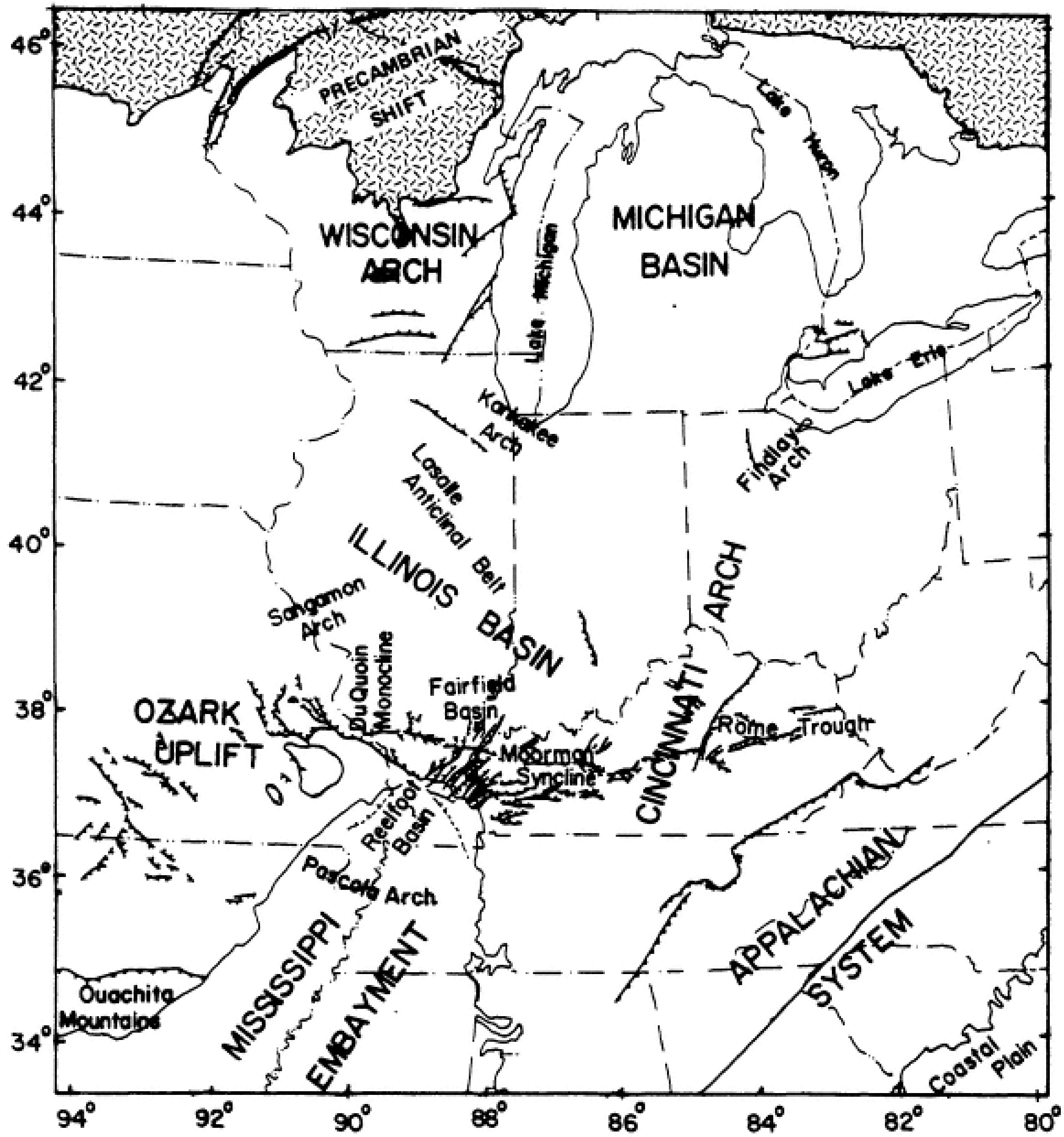


Figure 7.2
Major structural elements of the central midcontinent, United States. Modified from King (1969).

Although detailed seismicity data are generally unavailable, the value of these studies is demonstrated by recent results from a telemetered microearthquake array centered in southeast Missouri (Stauder *et al*., 1977). The pattern of seismicity (Figure 7.4), which has become evident during 21 months of recording, displays several linear trends in NE-SW and NW-SE directions, which are interpreted as indications of the pattern and extent of currently active faults.

Stress Distribution

A knowledge of the distribution of stress within the mid-continent should provide valuable insight into the origin of forces responsible for local contemporary tectonic activity. The regional stress distribution in the eastern United States as measured by several methods (Sbar and Sykes, 1973; Haimson, 1976) shows that the maximum compressive stress is nearly horizontal and trends east to northeast. Sbar and Sykes further suggest that plate motions are responsible for this stress and that earthquake zones in eastern North America are controlled by the existence of unhealed fault zones that are subject to high deviatoric stress. However, a focal mechanism study by Street *et al*. (1974) suggests that the stress distribution in the eastern United States can be locally complex. On the other hand, recent microearthquake focal mechanism studies by Herrmann and Canas (1978) indicate that the two prominent NE-SW trending lines of epicenters in the New Madrid region have focal mechanisms consistent with right-lateral strike-slip faulting along a NE-SW

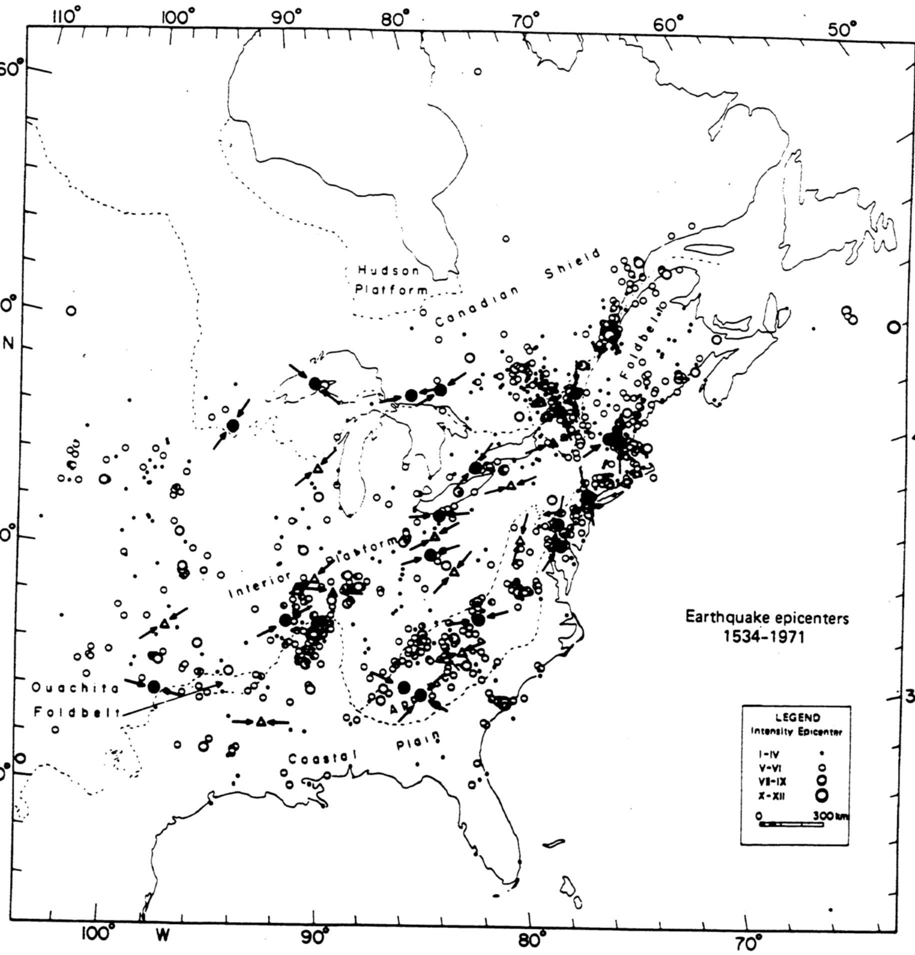


Figure 7.3
Distribution of reported earthquakes (1534-1971) in eastern North America (from York and Oliver, 1976) and selected tectonic features (from Sbar and Sykes, 1973; Haimson, 1976). Fault plane solutions of earthquakes (solid triangles), strain relief in situ stress measurements (solid circles), and hydrofracture in situ stress measurements (open triangles) are also shown. Strike of horizontal component of maximum or minimum compressive stress is shown at each locality.

trending fault plane (Figure 7.4). This faulting is consistent with the previously mentioned regional stress patterns. Thus, regional stress field and intraplate features and forces that may locally perturb or dominate it must be considered.

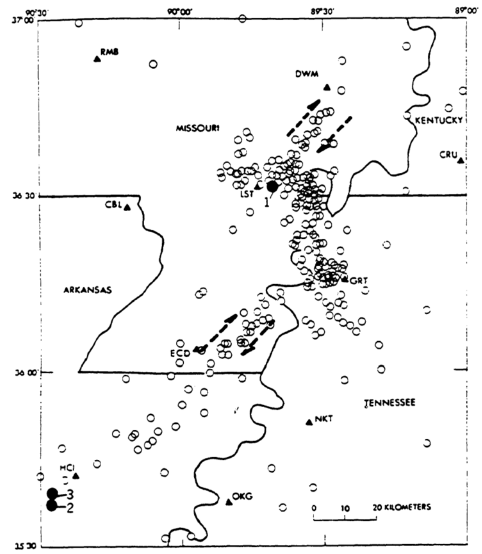


Figure 7.4
Epicenter map showing location of all earthquakes located within 1.5° × 1.5° area by the St. Louis University seismic network during the 21-month period of July 1974 to March 31, 1976 (from Stauder et al., 1977). Larger earthquakes are indicated by solid circles: 1, June 13, 1975, $m_b = 4.3$; 2, March 23, 1976, $m_b = 5.0$; and 3, March 25, 1976, $m_b = 4\text{-}5$. Generalized fault-plane solution of earthquakes after Hermann and Canas (1978).

Cenozoic Faulting

The level of tectonic activity is relatively low; therefore we must use evidence from the geological record to understand the contemporary tectonics of the midcontinent. Although sparse, data on Cretaceous and Cenozoic faulting become particularly important (York and Oliver, 1976).

Seismicity data are very valuable, but space-time magnitude relationships for earthquakes derived from the relatively short (~200 year) historical record are questionable. In fact, the use of the earthquake record to predict future seismic hazards has led Alien (1975) to state: "The very short historic record in North America should, therefore, be used with extreme caution in estimating possible future seismic activity. The geologic history of late Quaternary faulting is the most promising source of statistics on frequencies and locations of large shocks."

Recent Vertical Movements

Long-term, broad-wavelength, vertical crustal movements have an important role in continental tectonics, and their existence in the geological record is readily evidenced by uplifts and basins. Quantitative analysis of vertical movements from the geological record is difficult because of uncertainties in the age of marker units, inability to account for possible oscillatory movements or erosion cycles, and complicating effects related to glacial rebound. Analysis of precise leveling data provides an important source of information on vertical movements of the earth's crust. However, since the leveling measurements are repeated over relatively short periods of time (tens of years), vertical movements inferred from these data are indicative of "instantaneous" velocity and may not be representative of long-term tectonic motions.

An analysis of vertical crustal movements in the eastern United States determined from leveling data (Brown and Oliver, 1976) shows, in general, that modem vertical movements appear to be related to earlier Phanerozoic tectonic trends. However, the rates of modem movements are much larger than average rates over the last 130 m.y., and, thus, modem movements must be episodic or oscillatory. Furthermore, Brown and Oliver (1976) suggest a correlation between zones of vertical crustal movements and patterns of seismicity.

TECTONIC MODELS

The geological history and the contemporary geodynamics of the midcontinent region testify to its complex and continuing tectonic development. A wide variety of tectonic models (Figure 7.5) has been suggested to explain the contemporary tectonics of this region, and thus, it is important to consider the hypotheses leading to these models.

Resurgent Tectonics

Many models assume that much of the contemporary tectonic activity is controlled by pre-existing geological features. These models suggest that crustal rifts, zones of weakness and crustal boundaries, and local crustal inhomogeneities serve to localize in a passive manner the deformation resulting from stresses generated by a variety of tectonic forces. These forces may be and probably are completely alien to those initially responsible for the features, and, therefore, these models are grouped under the general term "resurgent tectonics."

Crustal Rifting

Rifting of the continental crust and its commonly associated igneous events are a major source of large-scale crustal disturbance and are therefore particularly susceptible to resurgent tectonics. It has become increasingly clear that rifting of the crust has played a major role in the geological history of central North America. This type of deformation, in which vertical movement predominates over horizontal displacement, has been operative for at least 1.6 b.y. since the stabilization of the continental interior.

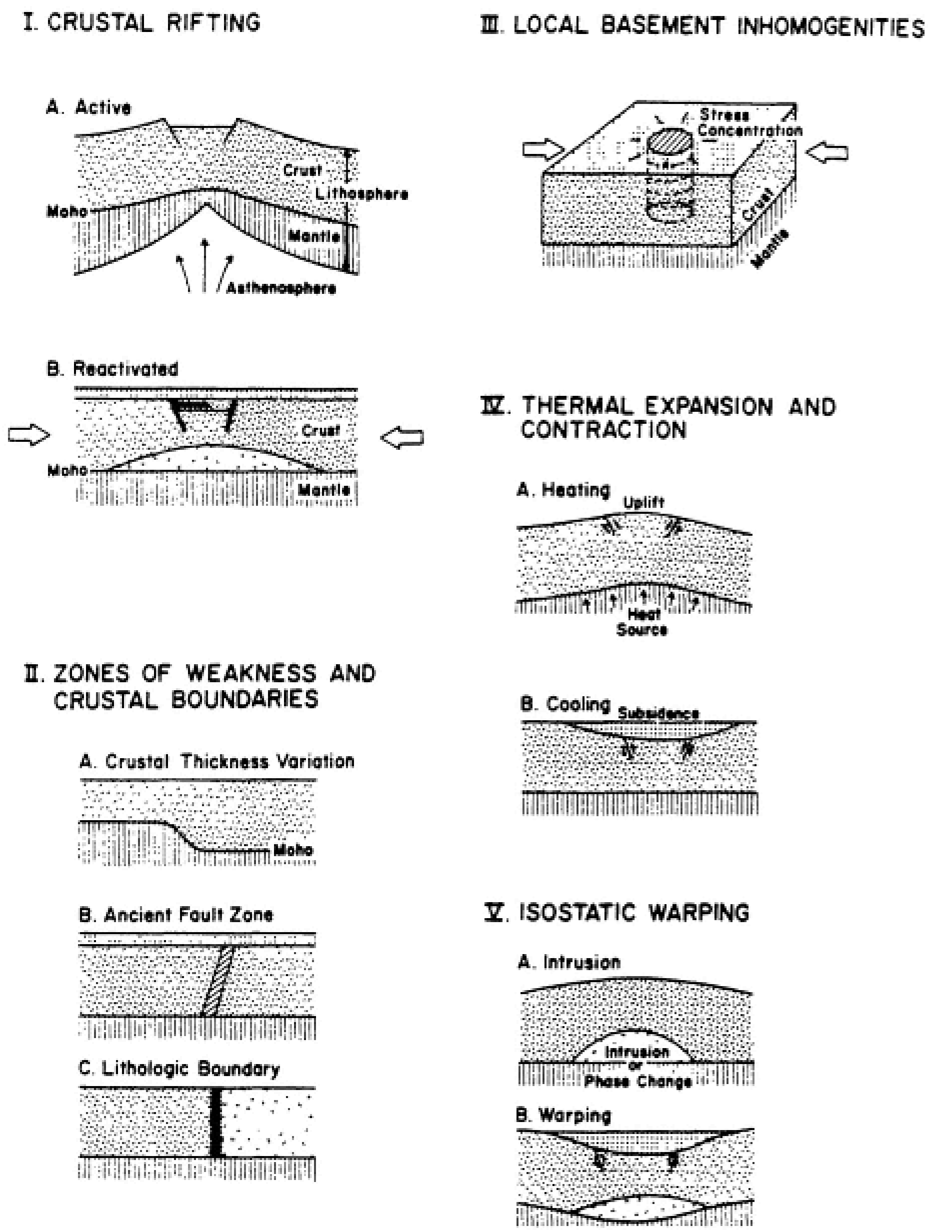


Figure 7.5
Schematic diagrams of proposed tectonic mechanisms.

The feature most widely accepted as being due to rifting is the linear midcontinent gravity high, which extends from Kansas to Lake Superior (Figure 7.1A). Recent studies by King and Zietz (1971) and Ocola and Meyer (1973) indicate that it is associated with a major, deep-seated rift zone of late Precambrian age. Strong arguments have also been made for extension of the rift zone to the southeast into the Michigan basin area (Hinze *et al*., 1975) and possibly farther south into Ohio, Kentucky, and Tennessee.

An interesting model for intraplate rifting has been proposed by Burke and Dewey (1973), which has several plume-generated, triple junctions in the midcontinent during late Precambrian time. These triple junctions became inactive before rifting proceeded to create an ocean basin. Although clearly an oversimplification, this model has the advantage of providing an integrated causal mechanism for this episode of widespread rifting. Sawkins (1976) explains that the rifting also may be associated with continental collision, leading eventually to continental breakup.

Triple junctions that did form ocean basins have also been postulated as having an important role in the tectonic development of the midcontinent. For example, the

Mississippi Embayment and southern Oklahoma aulacogen (Figure 7.1B) are interpreted as failed arms of triple junctions (aulacogens) formed as new episodes of ocean spreading begin. Burke and Dewey (1973) originally suggested that the Mississippi Embayment was a Mesozoic failed arm from a triple junction. Ervin and McGinnis (1975) evaluated this proposal in terms of the available geological and geophysical data and concluded that this model was a latest Precambrian-Eocambrian failed arm of the same origin as the southern Oklahoma aulacogen (Burke and Dewey, 1973) and was thus formed as a Precambrian land mass broke up prior to the formation of the Appalachian-Ouachita mountain belt. According to the model of Ervin and McGinnis (1975), this older feature (the Reelfoot Rift) was reactivated in Mesozoic time to form the present-day Mississippi Embayment. Evidence for a rift zone that coincides with the New Madrid seismic zone has been presented by Hildenbrand *et al*. (1977) in the form of gravity and magnetic data.

Kumerapeli and Saull (1966) make a strong case for the interpretation of the St. Lawrence Valley system as a rift zone that was probably initiated in Mesozoic time but is still active today. They suggested that this rift zone may extend into the midcontinent and connect with the New Madrid seismic zone. The implication of this model is that major earthquakes could be expected anywhere along a NE-SW trend extending from Arkansas to the St. Lawrence Valley.

Rifting has clearly played an important role in the tectonic development of the midcontinent, and the boundary and intrarift faults of these paleorifts are prime candidates for resurgent tectonics.

Zones of Weakness and Crustal Boundaries

Old zones of weakness are widespread in the midcontinent, and it is well established that they may have significant influence on subsequent structural development. For example, in northwestern Ohio the north-trending Bowling Green Fault coincides with the subsurface extension of the eastern boundary of the Grenville orogenic belt (Quick *et al*., 1976). The surface fault zones in eastern and central Kentucky (Figure 7.2) exhibit similar behavior in that the east-trending Kentucky River Fault zone and the Irvine-Paint Creek Fault zone appear to be the reactivated northern boundary of the Rome Trough (Ammerman and Keller, in press). These faults have been active as recently as the Pennsylvanian, but the major movement associated with the Rome Trough was Cambrian in age.

The fault complex of western Kentucky, southern Illinois, and southeastern Missouri (Heyl, 1972) is another example of resurgent tectonics. These faults outline a complex crustal block that encompasses the Rough Creek graben (Moorman syncline), Hicks Dome, the Illinois-Kentucky mineral district, and a concentration of mafic and ultramafic dikes. Many are believed to be reactivated Precambrian structures with movement on the major surface faults occurring during late Paleozoic and Cretaceous time. Of particular interest are the inferred northeast-trending faults at the northern extremity of the Mississippi Embayment and their extension to the northeast into the Wabash Valley Fault zone. Some of these faults coincide with or parallel geophysical anomalies, suggesting that they reflect buried basement structures (Lidiak and Zietz, 1976) perhaps associated with the rift system proposed for the northern portion of the embayment.

These examples demonstrate the importance of resurgent tectonics in the development of many of the structures of the midcontinent. Old zones of weakness appear to exercise control on the location and trend of many younger structures. However, there is not necessarily a relation between younger faulting and pre-existing structures. Some old fault zones have not been reactivated by younger faults (e.g., the southern boundary of the Rome Trough), and younger faults may not reflect old zones of weakness (e.g., the faults along the southern margin of the Moorman Syncline). It is clear that the resurgent character of a fault can be neither implied nor ignored.

Knowledge of large-scale crustal structure variations is important because these variations may localize or even generate regional stresses. The structure of the earth's crust has generally been assumed to be uncomplicated in the midcontinent region, but our knowledge is limited because few detailed studies have been performed. Recent studies suggest that the crustal structure of this region is more complicated than generally believed. For example, interpretation of two seismic refraction lines in southern Missouri and northeast Arkansas (McCamy and Meyer, 1966; Stewart, 1968), modeling of long period *P*-wave spectra near St. Louis, Missouri (Kurita, 1973), surface-wave dispersion measurements, and gravity anomalies suggest the presence of a basal high-velocity crustal layer. The geographical extent of this layer and its geological implications are at present unknown, but its possible spatial correlation with the New Madrid seismic zone suggests that it may have an effect on contemporary tectonics.

Local Basement Inhomogeneities

Local basement inhomogeneities in the form of mafic or ultramafic intrusives and that are the probable source of major gravity and magnetic anomalies have been recognized in the New Madrid seismic zone and other areas of eastern North America as being correlative with earthquake epicenters (McGinnis and Ervin, 1974; Long, 1976; Kane, 1977; McKeown, 1978). Generally, the earthquakes occur adjacent to positive gravity gradient areas.

The explanations of the origin of these correlations have taken two general forms: (1) the gravity and magnetic anomalies are related to crustal faults that are locally reactivated; and (2) the source of the anomalies reflects rigidity variations within the crust, which passively control the strain field causing concentration of earthquakes.

Although these correlations are potentially significant, these explanations raise several problems. The statistical validity of these correlations still remains to be proven because relative positive gravity and magnetic anomalies occur widely over the midcontinent, but, as discussed previously, earthquakes do not. Even if a statistically valid correlation exists, it is not clear that there is a cause and effect relationship. The origins stated above assume this type of relationship, but, as McKeown (1978) mentions, the seismicity may be related to reactivated paleorifts that are in part inferred from the presence of alkaline mafic rocks that produce positive gravity and magnetic anomalies. Thus, the anomalies and seismicity may have a common fundamental origin and not be related cause to effect. If the relationship is cause and effect, it is unclear why only a few positive anomalies are involved with seismicity and whether resurgent tectonics or crustal strength variations are involved.

Thermal Expansion and Contraction

Thermally induced forces, manifested in a variety of forms and patterns, are generally recognized as the principal origin of stress within the earth. A major method of translating thermal energy into stress is by thermal expansion and contraction. Thus, it is to be expected that this mechanism could be used to explain the geodynamics of plate interiors and contemporary tectonic activity of the midcontinent. Two general categories of models have been proposed: (1) models based on local thermal variations primarily related to igneous intrusions in the lithosphere or local heat-flow perturbations and (2) models based on mantle penetrative convection resulting in regional tensional and compressional patterns.

As an example of the first category of models, Sleep and Snell (1976) proposed that the gradual subsidence of mid-continent basins involves thermal contraction of the lithosphere complicated by time-dependent regional isostatic compensation. They imposed a creep mechanism as well as faulting to relieve isostatic imbalance in the subsiding basin. However, evidence of the thermal heating event that precedes subsidence is unknown for mid-continent basins.

The second general class of model is based on mantle penetrative convection. Burke and Dewey (1973) gave numerous examples of continental triple-rift junctions, including the Mississippi Embayment, which they suggested were formed in stationary continental lithosphere over mantle plumes. Presumably, rising material in deep-mantle plumes spreads out in the upper asthenosphere, producing stresses on the overlying plates. Similarly, Hinze *et al*. (1972) suggested that late Precambrian rifts are related to rising mantle plumes followed by slow cooling of the upper mantle, which is consistent with continuing tectonic activity over long periods of geological time. As Burke and Dewey (1973) point out, reactivation of paleorifts is common. Thus, regional expansion and contraction associated with surface uplift and subsidence over rising (hotter) and sinking (cooler) mantle may be related to the contemporary tectonics of the midcontinent.

Isostatic Warping

Regional variations in loading or unloading of the crust cause isostatic deviations, which lead to crustal warping and the possibility of related crustal rupture and earthquake activity. As a result, isostatic warping of the crust has been related to the contemporary geodynamics of the midcontinent region. Fox (1970), in discussing the origin of the seismicity of the eastern United States, suggested that rebound from the depression of the earth's surface by the weight of the Pleistocene glaciation may have triggered earthquakes. However, Woollard (1958) finds no relationship between isostatic imbalance and earthquake epicenters.

McGinnis (1963) has studied the frequency of earthquakes in the New Madrid seismic zone as a function of river stage along the Mississippi River. He concludes that the most obvious and influential triggering mechanism for earthquakes in this area is the change in surface load caused by the seasonal change in the amount of water held in alluvial valleys. An alternative explanation is that a corresponding increase in pore pressure may lead to decreased friction, which will trigger earthquakes. Similarly, Fitch and Muirhead (1974) conclude that the load of water in reservoirs may trigger the release of contemporary tectonic stress and result in earthquake activity.

Brecke (1964) suggests that downwarping of the Mississippi embayment under an increasing load of Mesozoic and Cenozoic sediments produces tectonic activity. McGinnis (1970) relates isostatic warping in the midcontinent to the emplacement of high-density intrusive and extrusive rocks in rifts with sedimentary loading in the rift producing additional isostatic subsidence, which culminates in the development of sedimentary basins such as the Illinois Basin and the Mississippi Embayment. A related hypothesis has been proposed by Haxby *et al*. (1976), who suggest that the Michigan Basin formed by elastic flexure due to conversion of the lower crust from gabbro to heavier eclogite by a hot mantle diapiric plume. As the mantle is cooled by conduction, the basin subsides under the load of eclogite.

CONCLUSIONS

A summary of data and interpretations that pertain to the tectonic framework of the midcontinent region of the United States has been presented, which should help to delineate critical information that is needed to obtain an understanding of the contemporary tectonism of this area. Accordingly, some general observations on the validity of the tectonic models are appropriate.

1. The New Madrid seismic zone is the focus of the most intense earthquake activity in the midcontinent re

gion and, therefore, deserves special comment as to the possible tectonic mechanisms responsible for the seismicity. However, sufficient data are not available to make definite statements on mechanism, and, thus, the following comments must be regarded as preliminary.

The area has been a focus of tectonic activity since Precambrian time, involving rifting and igneous activity. Therefore, it is reasonable to presume that significant crustal boundaries are present that may act as zones of weakness for reactivation by the regional stress field. This field appears to be dominated by generally uniform east to northeast horizontal compressive stress resulting from the relative motion of the North American plate. Recent microearthquake studies in the New Madrid zone indicate significant NE-SW trending zones of epicenters having a combined length of over 150 km and fault motion consistent with a NE-SW trending right-lateral strike-slip fault. A major crustal feature that may be the expression of an ancient rift appears to be related to the northern Mississippi Embayment and the New Madrid seismic zone and is evidenced by NE-SW trending gravity and magnetic anomalies.

The implications of these observations are that zones of weakness may exist in the crust in the New Madrid region and that earthquake activity may be expected anywhere along the extent of these features at least where they are oriented appropriately to the direction of regional compressive stress. Because of the contemporary stress field, rifting is not likely to be currently active. In the New Madrid region, the relatively consistent earthquake focal mechanisms along linear trends of epicenters suggest a source mechanism involving resurgence of an older feature (rift?). This conclusion implies that the potential seismic zone can be delineated if the detailed subsurface structure of the region is determined.

2. Any working tectonic model for the midcontinent region should be constructed within the framework of plate tectonics. We know of no major contemporary tectonic activity that cannot be generally explained by plate-tectonic theory. In fact, except for seismicity induced by man's activities, all major seismic zones similar to the New Madrid area appear to have their origin in plate motions. However, this is not to say that all tectonic activity is currently well understood or completely explained by our present knowledge of plate tectonics. Although the mechanisms of contemporary deformation in the midcontinent are unclear, we expect that intraplate tectonism ultimately related to the motion of the North American plate is the cause. The central midcontinent is part of the interior of the North American plate. This area is at present a relatively stable craton. However, it has had a long, complicated history of tectonic activity, which is difficult to unravel. It is obvious that the history of deformation and thermal events, largely pre-Mesozoic in age, has left its imprint on the crust of the midcontinent. Lateral and vertical variations in composition and physical properties, fault zones, and intraplate or province boundaries are evident in the crust. These features may serve to localize stress, enhance deformation, or act as zones of weakness.

ACKNOWLEDGMENTS

This report was supported by U.S. Nuclear Regulatory Commission contract AT(49-24)-0323 and is a condensed version of the Commission's Technical Report NUREG-0382(1977). The authors express their appreciation to T. C. Buschbach, N. B. Steuer, and the Nuclear Regulatory Commission for their interest and cooperation.

References

Allen, C. R. (1975). Geological criteria for evaluating seismicity, *Geol. Soc. Am. Bull. 86*, 1041-1057.

Ammerman, M. L., and G. R. Keller (in press). Delineation of the Rome Trough in eastern Kentucky with gravity and deep drilling data, *Am. Assoc. Petrol. Geol. Bull*.

Baer, A. J. (1976). The Grenville province in Helikian times: a possible model of evolution, *Phil. Trans. Roy. Soc. London A280*, 499-515.

Brecke, E. A. (1964). A possible source of solutions of the Illinois-Kentucky Fluorspar District, *Econ. Geol. 59*, 1293-1297.

Bristol, H. M., and T. C. Buschbach (1971). Structural features of the Eastern Interior Region of the United States, in Background Materials for Symposium on Future Petroleum Potential of NPC Region 9, D.C. Bond, chm., *Ill. State Geol. Surv., Ill. Petrol. 96*, 21-28.

Brown, L. D., and J. E. Oliver (1976). Vertical crustal movements from leveling data and their relation to geologic structure in the eastern United States, *Rev. Geophys. Space Phys. 14*, 13-35.

Brown, W. R. (1970). The Piedmont: investigations of the sedimentary record in the Piedmont and Blue Ridge of Virginia, in *Studies in Appalachian Geology, Central and Southern*, G. W. Fisher *et al*., eds., Wiley-Interscience, New York, pp. 335-349.

Burke, K., and J. F. Dewey (1973). Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks, *J. Geol. 81*, 406-433.

Denison, R. E., E. G. Lidiak, M. E. Bickford, and E. B. Kisvarsanyi (in press). Geology and geochronology of Precambrian rocks in the central interior region of the United States, *Econ. Geol*.

Ervin, C. P., and L. D. McGinnis (1975). Reelfoot Rift: reactivated precursor to the Mississippi Embayment, *Geol. Soc. Am. Bull. 86*, 1287-1295.

Fitch, T. J., and K. J. Muirhead (1974). Depths to larger earthquakes associated with crustal loading, *Geophys. J. Roy. Astron. Soc. 37*, 285-296.

Fox, F. L. (1970). Seismic geology of eastern United States, *Assoc.Eng. Geol. Bull. 7*, 21-43.

Goldich, S. S. (1968). Geochronology in the Lake Superior region, *Can. J. Earth Sci. 5*, 715-724.

Haimson, B. C. (1976). Crustal stress in the continental United States as derived from hydrofracturing tests, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 576-592.

Haxby, W. F., D. L. Turcotte, and J. M. Bird (1976). Thermal and

mechanical evolution of the Michigan Basin, in *Sedimentary Basins of Continental Margins and Cratons*, M. H. P. Bott, ed., *Tectonophysics 36*, 57-75.
Herrmann, R. B., and J. A. Canas (1978). Focal mechanisms studies in the New Madrid seismic zone, *Bull. Seismol. Soc. Am. 68,* 1095-1102.
Heyl, A. V. (1972). The 38th parallel lineament and its relationship to ore deposits, *Econ. Geol. 67*, 879-894.
Hildenbrand, T. G., M. F. Kane, and W. Stauder (1977). Magnetic andgravity anomalies in the northern Mississippi Embayment and their spacial relation to seismicity , *U.S. Geol. Surv. Map MF-914*.
Hinze, W. J., R. F. Roy, and D. M. Davidson (1972). The origin of late Precambrian rifts, *Geol. Soc. Am. Abstr. Programs*, 725.
Hinze, W. J., R. L. Kellogg, and N. W. O'Hara (1975). Geophysical studies of basement geology of Southern Peninsula of Michigan, *Am. Assoc. Petrol. Geol. Bull. 59*, 1562-1584.
Kane, M. F. (1977). Correlation of major eastern earthquake centers with mafic/ultramafic basement masses, *U.S. Geol. Surv. Prof. Paper 1028-0*, 199-204.
King, E. R., and I. Zietz (1971). Aeromagnetic study of the mid-continent gravity high of central United States, *Geol. Soc. Am. Bull. 82*, 2187-2208.
King, P. B, compiler (1969). Tectonic map of North America, *U.S. Geol. Surv. Map*, scale 1:5,000,000.
Kumarapeli, P. S., and V. A. Saull (1966). The St. Lawrence valley system: A North American equivalent of the east African rift valley system, *Can. J. Earth Sci. 3*, 639-658.
Kurita, T. (1973). Regional variations in the structure of the crust in the United States from *P*-wave spectra, *Bull. Seismol, Soc. Am. 63*, 1663-1687.
Lidiak, E. G., and I. Zietz (1976). Interpretation of aeromagnetic anomalies between latitudes 37ºN and 38ºN in the eastern and central United States, *Geol. Soc. Am. Spec. Paper 167*, 37 pp.
Long, L. T. (1976). Speculations concerning southeastern earthquakes, mafic intrusions, gravity anomalies, and stress amplifications, *Earthquake Notes 47*, 29-35.
McCamy, K., and R. P. Meyer (1966). Crustal results of fixed multiple shots in the Mississippi Embayment, in *The Earth Beneath the Continents,* J. S. Steinhart and T. J. Smith, eds., Am. Geophys. Union Geophys. Monogr. 10, pp. 370-381.
McGinnis, L. D. (1963). Earthquakes and crustal movement as related to water load in the Mississippi Valley region, *Ill. State Geol. Surv. Circ. 344*, 20.
McGinnis, L. D. (1970). Tectonics and gravity field in the continental interior, *J. Geophys. Res. 75*, 317-331.
McGinnis, L. D., and C. P. Ervin (1974). Earthquakes and block tectonics in the Illinois Basin, *Geology 2*, 517-519.
McKeown, F. A. (1978). Hypothesis: many earthquakes in the central and southeastern United States are casually related to mafic intrusive bodies, *J. Res. U.S. Geol. Surv. 6*, 41-50.
Nuttli, O. W. (1974). Magnitude-recurrence relation for central Mississippi valley earthquakes, *Bull. Seismol. Soc. Am. 64*, 1189-1207.
Ocola, L. D., and R. P. Meyer (1973). Central North American rift system: 1. Structure of the axial zone from seismic and gravimetric data, *J. Geophys. Res. 78*, 5173-5194.
Quick, R. C., E. F. Pawlowicz, and W. J. Hinze (1976). The Bowling Green Fault—a case of resurgent tectonics? (abstr.), in *American Association of Petroleum Geologists, Eastern Section Meeting*, Lexington, Kentucky, p. 20.
Rankin, D. W. (1976). Appalachian salients and recesses: late Precambrian continental breakup and the opening of the Iapetus Ocean, *J. Geophys. Res. 81*, 5605-5619.
Sawkins, F. J. (1976). Widespread continental rifting: some considerations of timing and mechanism, *Geology 4*, 427-430.
Sbar, M. L., and L. R. Sykes (1973). Contemporary compressive stressand seismicity in eastern North America , an example of intraplate tectonics, *Geol. Soc. Am. Bull. 84*, 1861-1882.
Sleep, N. H., and N. S. Snell (1976). Thermal contraction and flexure of midcontinent and Atlantic marginal basins, *Geophys. J. Boy. Astron. Soc. 45*, 125-154.
Stauder, W., M. Kramer, G. Fischer, S. Schaefer, and S. Morrissey (1977). Seismic characteristics of southeast Missouri as indicated by a regional telemetered microearthquake array, *Bull. Seismol. Soc. Am. 66*, 1953-1964.
Stewart, S. W. (1968). Crustal structure in Missouri by seismic refraction methods, *Bull. Seismol. Soc. Am. 58*, 291-323.
Street, R. L., R. B. Herrmann, and O. M. Nuttli (1974). Earthquake mechanics in the central United States, *Science 184*, 1285-1287.
Woollard, G. P. (1958). Areas of tectonic activity in the U.S. as indicated by earthquake epicenters, *Eos Trans. Am. Geophys. Union 39*, 1135-1150.
York, E., and J. E. Oliver (1976). Cretaceous and Cenozoic faultingin eastern North America , *Geol. Soc. Am. Bull. 87*, 1105-1114.

8

Problems of Intraplate Extensional Tectonics, Western United States

Gregory A. Davis
University of Southern California

INTRODUCTION

Intraplate extensional tectonics within continents has traditionally been viewed from the standpoint of high-angle normal faulting. Other extensional mechanisms (dikes, low-angle normal faults, zones of ductile flow) have received much less attention. Normal faults within continental plates occur in widely varying settings, both orogenic and anorogenic. The spatial association of normal faulting with orogenic belts is so common that many geologists have considered that block faulting is a necessary late or postorogenic phase of a geotectonic cycle (e.g., Roberts, 1972). Nevertheless, support for the concept of a geotectonic cycle has justifiably waned with the advent of plate tectonics (Coney, 1970), and there is little in the disparate examples of normal faulting within orogenic belts to indicate a single underlying cause for them.

One of the most widespread categories of intracontinental normal faults is the type described by Burke (Chapter 4)—normal fault systems that lead, unless arrested in their development, to the complete rifting and separation of continental lithosphere. Zones of continental plate attenuation and rifting appear to develop indiscriminately across both orogenic and cratonic areas. Rifts will not specifically be treated here, although the possibility cannot be dismissed entirely that some examples of Cordilleran normal faulting such as in the Basin and Range province might, from a causal standpoint, be atypical expressions of continental rifting of the Atlantic-Red Sea-East African type.

Normal faults in the Gulf Coast area of the southeastern United States are the dominant Cenozoic tectonic features of that region, but their origin as growth structures accompanying continental-margin sedimentation appears well documented. The Cordillera of the western United States offers many examples of structures that do not appear to fall in the categories cited above. These form the basis for this summary of the problems of continental, intraplate extensional tectonics.

What are the problems of intraplate extension within the western portion of the North American continent? They encompass all aspects of the formulation of structural elements—geometric, kinematic, and dynamic. What, for example, is the geometry at depth of normal faults in the Basin and Range province? What is the total amount of extension represented by faulting in the Great Basin area of that province? Geometrically, how is brittle distension in the uppermost part of the lithosphere accommodated at deeper structural levels? Have patterns of Cenozoic extension been influenced by pre-existing crustal anisotropies? If so, how, and by what controls? What are the kinematics of Basin and Range faulting, and have they changed in time? How are diverse styles of coeval late Cenozoic extension in the northwestern, western, and southwestern United States related geometrically and kinematically? Does such extension represent the response of the North American plate to transform motion along its western edge, to back-arc "spreading," to the inception of subplate asthenospheric plumes, to rifting of Atlantic type, or to undetermined factors? In what ways are stress fields in upper plates along convergent plate boundaries determined by the geometry of subduction and the rate of plate convergence? This list of questions is hardly conclusive, but it serves to illustrate the great diversity and scope of uncertainties regarding the nature of intraplate extension in just one part of the North American continent.

EARLY TERTIARY EXTENSION IN THE PACIFIC NORTHWEST

A large region of the Pacific Northwest experienced a poorly understood extensional event in early Tertiary time with the reinitiation of arc volcanism following a Late Cretaceous-early Tertiary magmatic hiatus (Figure 8.1). Volcanic activity began between 55 and 50 million years (m.y.) ago and was spatially accompanied by regional N-S block faulting across the entire U.S.-Canadian border area from northwestern Washington to northwestern Montana (125-116º W longitude).

Eocene uplift of the Northern Cascade Mountains was accompanied by dip-slip reactivation of earlier strike-slip faults (Hope-Straight Creek and Chewack-Pasayten), formation of the Chiwaukum graben, and intrusion of the Teanaway basalts in N-S-striking dike swarms. The Chiwaukum graben contains interbedded fanglomerates and fluviatile sediments, interpreted by Whetten (1976) to represent deposition during block faulting.

There is also a clear spatial relationship between block faulting and regional andesitic magmatism in areas east of the North Cascades. Formation of the Republic graben in the Okanogan region was accompanied by andesitic volcanism (52-43 m.y. ago; Fox *et al*., 1977) and deposition of graben sediments. To the east, the southern Rocky Mountain Trench, a half-graben at the time, came into being during the Eocene or early Oligocene as recorded by syntectonic sedimentation (Clague, 1974). Unfortunately, the southward extent beneath the Columbia Plateau basalts of the steep bounding faults of the Chiwaukum, Methow-Pasayten, and Republic grabens is not known.

The total amount of crustal extension during this early Tertiary event was probably not great, since the major faults dip steeply, but reasons for the intra-arc setting of the block-faulted area and its great E-W breadth are unclear. Shallow dipping subduction of an oceanic plate could explain the unusual breadth of the arc, but the consequences of such subduction beneath the Andes of central Peru are a shutoff of arc volcanism and an upper-plate compressional stress state oriented more or less perpendicularly to the trench (Stauder, 1975). Miocene crustal extension in the Pacific Northwest (dike development and normal faulting) occurred in areas east of the contemporaneous Cascade volcanic arc, not spatially coincident with it as in the case of the Eocene events.

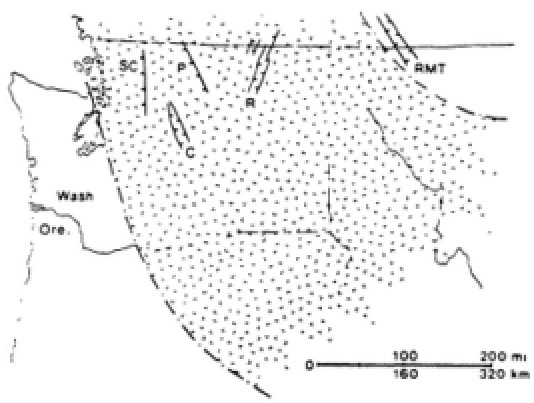


Figure 8.1
Location map of Eocene-Oligocene (55-40.1 m.y. ago) volcanic are in the Pacific Northwest and contemporaneous block faulting. Position of the patterned are from Snyder et al. (1976). SC, Straight Creek Fault; C, Chiwaukum graben; P, Pasayten Fault; R, Republic graben; RMT, Southern Rocky Mountain Trench.

MIOCENE EXTENSION IN THE WESTERN UNITED STATES

The Miocene was a time of profound tectonic change for the western United States in areas extending from southeastern Washington to southeastern California. This entire region experienced diverse but strikingly synchronous extensional deformation beginning approximately 16 or 17 m.y. ago. In the Pacific Northwest, crustal dilation was represented by extensive, primarily NNW-trending basaltic dike swarms in central and northeastern Oregon, southeastern Washington, and western Idaho. The rifted western portion of the Snake River Plain was also formed at this time, but the amount of distension represented by this feature is controversial.

Crustal extension was pronounced in the Great Basin area of Nevada, Utah, and adjoining states, where widespread normal faults characterize the deformation and control the topography. Volcanic activity that accom panied normal faulting was bimodal—predominantly basaltic but with subordinate rhyolite and rhyodacite (see Chapter 14). Farther south in the Colorado River area between California-Nevada and Arizona, block faulting of classic Basin and Range type generally did not occur, but extreme high-level crustal distension appears to be represented by an enigmatic terrane of low-angle Miocene faults of still unknown extent and origin.

Dike Swarms, Columbia River Basalt

The dike swarms of the Columbia River basalt occur principally in the Snake River area between northeastern Oregon and Idaho (Figure 8.2) and constitute the feeders for one of the most voluminous outpourings of basaltic lava in the geological record. Within a brief 3 m.y. period (ca. 16.5-13.4 m.y. ago) more than 200,000 km^3 of Columbia River tholeiitic basalts were erupted as three geochemically distinct groups—Picture Gorge, Imnaha, and Lower Yakima (McDougall, 1976); the latter is by far the most voluminous and really extensive. Watkins and Baksi (1974) conclude that this rate of extrusion exceeds any witnessed in historical times and is four to six times greater than that which occurs at typical midocean spreading centers. Swanson and Wright (1977) disagree. They have calculated that yearly-averaged flow rates for the basaltic province were 0.07 km^3/year, roughly comparable to Hawaiian lava production in the last 3 m.y. (0.08 km^3/year) and greater than the rate of Icelandic volcanism (0.05 km^3 /year) during the last 10,000 years. On the average, Columbia River flows were erupted in a given area every 10,000-20,000 years and with average volumes of 10-20 km^3. Some single flows with eastern Plateau feeders were so voluminous (up to 600 km^3) and were erupted in so brief a time (a few days) that they reached the Pacific Ocean near the present mouth of the Columbia River (Figure 8.2).

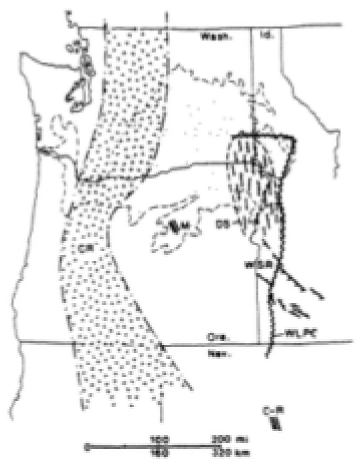


Figure 8.2
Location map of Miocene (20-10.1 m.y. ago) volcanism in the Pacific Northwest. Position of Western volcanic arc from Snyder et al. (1976). Distribution of Columbia River basalts (stippled pattern) and limits of Grande Ronde-Cornucopia dike swarm (DS) from Snavely et al. (1973). CR, Cascade Range; M, Monument dike swarm; WSR, Western Snake River graben; C-R, dike swarms in Cortez and Roberts Mountains (Stewart et al., 1975); WLPC, western limit of Precambrian continental crust based on initial $^{87}Sr/^{88}Sr$ isotopic ratios (Armstrong et al., 1977).

It is notable that the eastern Oregon-western Idaho dike swarms from which most of the basalts were apparently extruded roughly coincide with the north-south contact between eastern Precambrian continental crust and younger "eugeosynclinal" units in the Oregon-Idaho border region (Figure 8.2). It is suggested that this Mesozoic tectonic boundary, or suture, represented a fundamental zone of crustal weakness that was reactivated in Miocene time during ENE-WSW regional extension. Similar and contemporaneous basaltic andesite dike swarms are present in the Cortez and Roberts Mountains of north-central Nevada (Stewart *et al*., 1975); they also strike northerly (ca. N 20º W) and occur near the postulated westward limit of Precambrian crust in the underlying basement (Figure 8.2).

Zoback and Thompson (1978) have proposed continuity of Miocene rifting along a 700-km-long linear zone that includes the Nevada dike swarms described above, the graben of the western Snake River Plain, and the feeder dikes of the Columbia River basalts. The amount of Miocene crustal extension clearly diminishes northward from the Basin and Range province (major normal faulting) to the eastern Columbia Plateau-Blue Mountains region (vertical fissuring). Therefore, it is not surprising that evidence for Miocene extension is not seen still farther north, for example in northeastern Washington.

Great Basin Extensional Tectonics

The extent and general topographic and structural characteristics of the Great Basin area of the Basin and Range province are well known (e.g., Hamilton and Myers, 1966; Stewart, 1971). Nevertheless, many fundamental geological aspects of the Great Basin area are not well understood. No attempt is made to review the geophysical characteristics of the Basin and Range province, as they are treated in a companion paper by Eaton (Chapter 9).

Geometry of Range-Front Faults in the Great Basin

In most areas of the Great Basin (Figure 8.3) range-front faults are observed to dip 50-6º [although Donath (1962) reports that near-vertical faults are dominant in south-central Oregon]. Many workers (e.g., Stewart, 1971) have projected graben-bounding faults downward at constant angles to determine the depths below the surface where pairs of such faults would intersect and define the level of transition between shallow extension by normal faulting and extension by other, deeper mechanisms. Most estimates fall in the range of 14-17 km.

Figure 8.3
Distribution of extensional tectonics in the southwestern United States. Areas affected by normal faulting of basin-range type are indicated by NE-SW ruled pattern. Areas of low-angle detachment faulting are stippled. The inferred distribution of crustal low-velocity layer (Smith, 1978) is shown by NW-SE ruled pattern. Geographic localities from north to south: SRP, Snake River Plain; Y, Yerington; S, Snake Range; F, Funeral Range; P, Pahranagat Shear System; G, Garlock Fault; LV, Las Vegas; E, Eldorado Mountains; W, Whipple Mountains; P, Phoenix; T, Tucson.

However, other workers (e.g., Wright and Troxel, 1973; Proffett, 1977) have supported the early suggestion of Longwell (1945) that at least some basin-range faults must flatten at depth in order to explain rotational tilting of hanging-wall strata toward the faults. Deep seismic reflection studies across selected grabens in the Great Basin confirm a downward flattening of some range faults (A. W. Bally, Shell Oil Company, personal communication, 1978). The "bottoming-out" of basin-range faults may, therefore, occur at shallower levels than those depths determined by projecting faults downward at constant dip. Wright and Troxel (1973) believe that normal faults east of Death Valley, California, may flatten to the horizontal at depths of only 5-10 km, although Proffer (1977) postulates that curviplanar range-front faults in the Yerington area of western Nevada did not become horizontal until depths of at least 13-16 km.

Amount of Great Basin Extension

Estimates of the amount of extension across the Great Basin vary widely and are highly influenced by the assumptions made regarding the geometry of range-front faults. At 40º N latitude the Great Basin is 750 km wide. Stewart (1971) assumed average dips for range-front faults of 60º and calculated 2.5 km of average extension across each major graben in the Great Basin at this latitude, for a total E-W extension in the province of 75 km (approximately 10 percent). Extrapolating from a single basin in northern Nevada (Dixie Valley), Thompson and Burke (1974) postulate 100 km of total Great Basin extension. Alternatively, Proffer (1977) estimates 160-180 km of overall extension (by his calculation, approximately 30-35 percent) based on an assumed downward flattening of normal faults in some portions of the Great Basin and on the observed patterns of crustal thinning beneath the region. Hamilton's most recent estimate of total extension across the Great Basin is from 50 to 100 percent depending on the extent to which the tectonically thinned crust has been thickened by Cenozoic magmatism (Hamilton, 1978, and personal communication, 1978).

There are certainly localized areas of extreme extension within the Great Basin. Wright and Troxel (1973) calculated that a 30-5 percent extension has occurred along downward-flattening normal faults between Death Valley and the Nopah Range, California. Davis and Burchfiel (1973) concluded from a geometric analysis of a larger area that included the area studied by Wright and Troxel that E-W crustal extension north of the Carlock Fault between the Nopah Range and the Spangler Hills has been approximately 100 percent. Finally, Proffer (1977) estimates more than 100 percent extension in the Yerington district near Reno as the consequence of multiple, superposed generations of curviplanar normal faults. Although these examples are not thought by the writer to be characteristic of the entire Great Basin, they are instructive in indicating that local or regional estimates of limited extension (10 percent or less) based on unproven assumptions of steep faults and constant dips may be highly erroneous and much too conservative.

Relations Between Extension and Strike-Slip Faulting

Not all portions of the western United States have experienced the same amounts of Cenozoic extension. The boundaries between areas of differential extension are geometrically interesting and of considerable importance in defining the kinematics of intraplate deformation. Davis and Burchfiel (1973) and Lawrence (1976) have defined the southwestern and northwestern boundaries of the Great Basin, respectively, as intraplate transform or strike-slip faults that separate the distended Great Basin from southern and northern terranes that lack basin-range structure. The ENE- to E-striking Garlock Fault (Figures 8.3 and 8.4) is the most impressive of the transform boundaries. It represents a major lithospheric structural element that abruptly separates two regions (Great Basin and Mojave Desert) with dissimilar crustal thicknesses and seismic velocity characteristics (Davis and Burchfiel, 1973).

An unresolved, fundamental problem of Great Basin genesis is why lithospheric extension terminated southward so abruptly along an ENE- to E-striking boundary that was to become the Garlock Fault. Significant variations in initial strontium isotopic ratios from Mesozoic granitic rocks in southern California indicate to Kistler and Peterman (1978) that the Garlock Fault developed along a major pre-Mesozoic boundary in the crust. Thus, mechanical or compositional crustal anisotropies may have influenced the pattern of Cenozoic extension in this area. A similar general conclusion was tentatively reached above for the localization of Columbia River basaltic dike swarms in eastern Oregon and western Idaho.

Variable amounts of distension within the Basin and Range province may also be accommodated by internal strike-slip or transform fault zones. The Pahranagat shear system of southeastern Nevada (Liggett and Ehrenspeck, 1974) is a left-lateral displacement that appears to connect nonaligned northern and southern areas of major normal faulting (Figure 8.3). The transform zone strikes approximately N 60º E and thus defines the relative direction of crustal dilation in this part of the Great Basin at the time of its formation. A kinematic analysis of the entire Great Basin might be made by looking for other examples of strike-slip faults that developed synchronously with normal faulting as the consequence of differential extension.

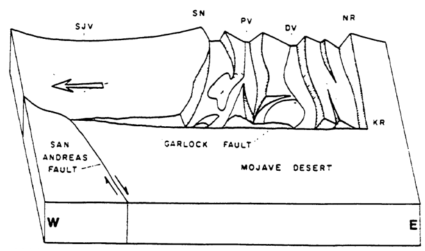


Figure 8.4
Northward diagrammatic view of Garlock Fault, southern California, as a boundary between a northern, distended crustal block (Basin and Range province) and a southern, nondistended crustal block (Mojave Desert). Topographic relations are highly generalized and are shown only north of the Garlock Fault. Geographic localities: SJV, San Joaquin Valley; SN, Sierra Nevada; PV, Panamint Valley; DV, Death Valley; NR, Nopah Range; KR, Kingston Range. (From Davis and Burchfiel, 1973, reprinted from the Bulletin of the Geological Societal of America, with permission).

Mechanisms of Crustal Extension at Depth

The Great Basin is characterized by a crust considerably thinner than that of surrounding regions (Smith, 1978; Thompson and Burke, 1974), a relation best explained as the consequence of Cenozoic tectonism. Attenuation of the 18- to 30-km-thick Great Basin crust cannot be due solely to normal faulting because various estimates for the depths to which normal faulting occurs fall in the range of 5-17 km. An area of possible exception is the west side of the Great. Basin, where Smith (1978) reports crustal thickness as little as 18 km, and Proffett (1977) proposed normal faulting to depths of at least 13-16 km. In general. however, mechanisms other than normal faulting must be sought for extension of the lower Great Basin crust.

Thompson and Burke (1974) and Wright and Troxel (1973), among others, have proposed that extension by normal faults in the upper crust may be accommodated wholly or in part by the intrusion of igneous dikes and plutons at depth. The spatial and temporal relations between basin-range faulting and volcanism. however. are so inconsistent as to cast doubt on the overall utility of this mechanism. Igneous intrusion could account for deep crustal extension, but there is no *a prior* reason to believe that it would produce attenuation of the lower crust.

Extension and thinning could, alternatively, be the consequence of the penetrative, ductile flow of lower crustal rocks, a phenomenon made more probable in light of the high heat flow of the province and the occurrence in western Utah and eastern Nevada of metamorphic and mylonitic terranes that experienced metamorphic temperatures as recently as the Miocene (Compton *et al*., 1977). Smith (1978) reports that there is evidence for the existence of a crustal low-velocity zone in this eastern Great Basin region today. The zone extends from a depth of 5-15 km and might conceivably coincide with a transition from highly brittle deformation to ductile deformation. Proffett (1977) has recently proposed a geometric model for explaining the occurrence of such a transition between downward flattening basin-range faults and deeper levels of laminar flow.

LOW-ANGLE DETACHMENT FAULTS

Nearly 20 years ago Misch (1960) described an extensive terrane in the border area between northern Nevada and Utah that is, he believed, underlain by a low-angle regional fault (Figure 8.3). He named this fault the Snake Range décollement, and the area in which it occurs the "Northwestern Nevada structural province." The décollement typically separates complexly faulted, unmetamorphosed upper-plate rocks from stratigraphically older, lower-plate crystalline complexes, the latter now known to include both Mesozoic and Tertiary metamorphic assemblages. Misch (1960) noted that while some mountain ranges in the structural province were fault blocks of basin and range type. others were elongate-domal uplifts with only subordinate block faulting. In recent years décollement structures and domal ranges of the type recognized by Misch have also been discovered in more southerly areas of the Cordillera (Figure 8.3), in the Death Valley area of California (Reynolds, 1974), and, more extensively, in southernmost Nevada, southeastern California. and Arizona.

Many of the low-angle faults in the Nevada-Utah region studied by Misch are found in areas that lack Tertiary rocks (or direct evidence for the involvement in faulting of Tertiary units) and are complicated by the superposed disruptive effects of basin-range faulting. As a consequence, the age and origin of these low-angle faults are highly controversial. Nevertheless, Armstrong (1972) has presented a strong case that the Snake Range décollement and higher, related faults (1) are Tertiary in age or, at the least, show Tertiary reactivation; (2) formed in an extensional regime; and (3) result from thinning of supracrustal rocks by normal faulting above a basal detachment surface (denudational tectonics). In contrast, Misch (1971) considers the Snake Range décollement to be Mesozoic in age and compressional in origin. Hose and Dane (1973) offer still another interpretation—the décollement re-suited from regional eastward gravity sliding during Mesozoic time.

The case for a Tertiary origin for the low-angle faults of the northern Nevada-Utah area is greatly enhanced by geological relations in the closely similar (and comparably enigmatic) terrane of low-angle faults that begins near Hoover Dam in the Las Vegas area, follows the Colorado River trough as far south as Parker, then swings southeastward and extends across southern Arizona into areas near Tucson (Figure 8.3). On the flanks of several ranges that characteristically owe their elevation to doming or arching rather than block faulting, basal detachment faults are found that separate allochthonous upper-plate units from crystalline rocks in the cores of the ranges. Tertiary strata as young as Middle Miocene are widely involved in the low-angle faulting, as are Precambrian gneisses, Paleozoic and Mesozoic metasedimentary rocks, and Precambrian and Mesozoic plutons. Lower-plate rocks in the terrane include Precambrian gneisses and plutonic rocks, Mesozoic and Tertiary plutonic rocks, and strongly lineated mylonitic gneisses of Cretaceous(?) and Tertiary age.

Ongoing studies by J. Lawford Anderson and the writer, and their students at the University of Southern California, in the Colorado River portion of the Nevada-California-Arizona dislocational terrane are reviewed below, but with no assurance that conclusions drawn from these studies are applicable to portions of the terrane farther east, e.g., in the Tucson area (Davis, 1975; 1977), or in the Nevada-Utah terrane first described by Misch (1960). Carr and Dickey (1977), Lucchitta and Suneson (1977), and Rehrig and Reynolds (1977; in press) have also investigated areas within the terrane described here.

Colorado River Trough South of Lake Mead

Low-angle faults cutting rocks of Precambrian through Miocene age were mapped in 1960, by geologists of the Southern Pacific Land Company in the mountain ranges surrounding Needles, California (Figure 8.5). They concluded that the faults were late Tertiary thrust faults, but their important studies were never published. Terry. (1972) recognized an extensive subhorizontal fault in the Whipple Mountains, California, 60 km south of Needles, and interpreted it to be a thrust fault of Cretaceous age. Anderson's (1971) study of low-angle faulting in the Eldorado Mountains, Nevada, 100 km north of Needles, was the first to attribute the low-angle faults of the Colorado River area to noncompressional tectonics. He described the Eldorado Mountains as an area of major imbricate normal faulting accompanied by pronounced rotational tilting of Tertiary strata. Miocene normal faults that dip westward were mapped as flattening downward and merging with a subhorizontal basal fault surface below which extension by normal faulting has not occurred. Displacement of Tertiary rocks above the basal surface was thus westward (S 70º W) relative to lower-plate autochthonous units.

Remapping of the Whipple Mountains Fault recognized by Terry (1972) confirms that it also is of Tertiary age and that it underlies a 3000-km^2 normal faulted terrane that includes the Whipple, Buckskin, and Rawhide Mountains of western Arizona (Shackelford, 1976a, 1976b, 1977; Lingrey *et al*., 1977; Davis *et al*., 1977). The surface may be coextensive with a similar basal dislocation surface mapped by Southern Pacific geologists on the flanks of the Chemeheuvi, Sacramento, Homer, and Dead Mountains to the north (Figure 8.5). As in the Eldorado Mountains, the basal detachment or dislocation surface(s) separates an allochthonous upper-plate assemblage of Tertiary sedimentary and volcanic units (and their crystalline basement) from lower-plate crystalline rocks that are mylonitic in some ranges. Northwest-striking, typically northeast-dipping, normal faults occur throughout the upper plate in the area of Figure 8.5 but are absent from lower-plate autochthonous rocks. The normal faults (not shown on the figure) both merge with and are cut by the basal dislocation surface(s), indicating a complex movement history for the upper plate.

Kinematic relations (fault striae, rotational tilting of Tertiary strata, and drag folds) indicate that tectonic transport of upper-plate units in the area of Figure 8.5 was predominantly northeastward (N 50±10º E, opposite to that in the Eldorado Mountains). This striking consistency of upper-plate movement direction clearly indicates that the basal dislocation surface(s) and higher related faults formed prior to the regional doming that produced the present ranges. No evidence has been found that allochthonous units moved radially away from uplifted areas, in contrast to the conclusions drawn by Coney

(1974) for units above the Snake Range décollement in the Snake Range and by Davis (1975) for folded rocks above the Catalina Fault in the Rincon Mountains near Tucson.

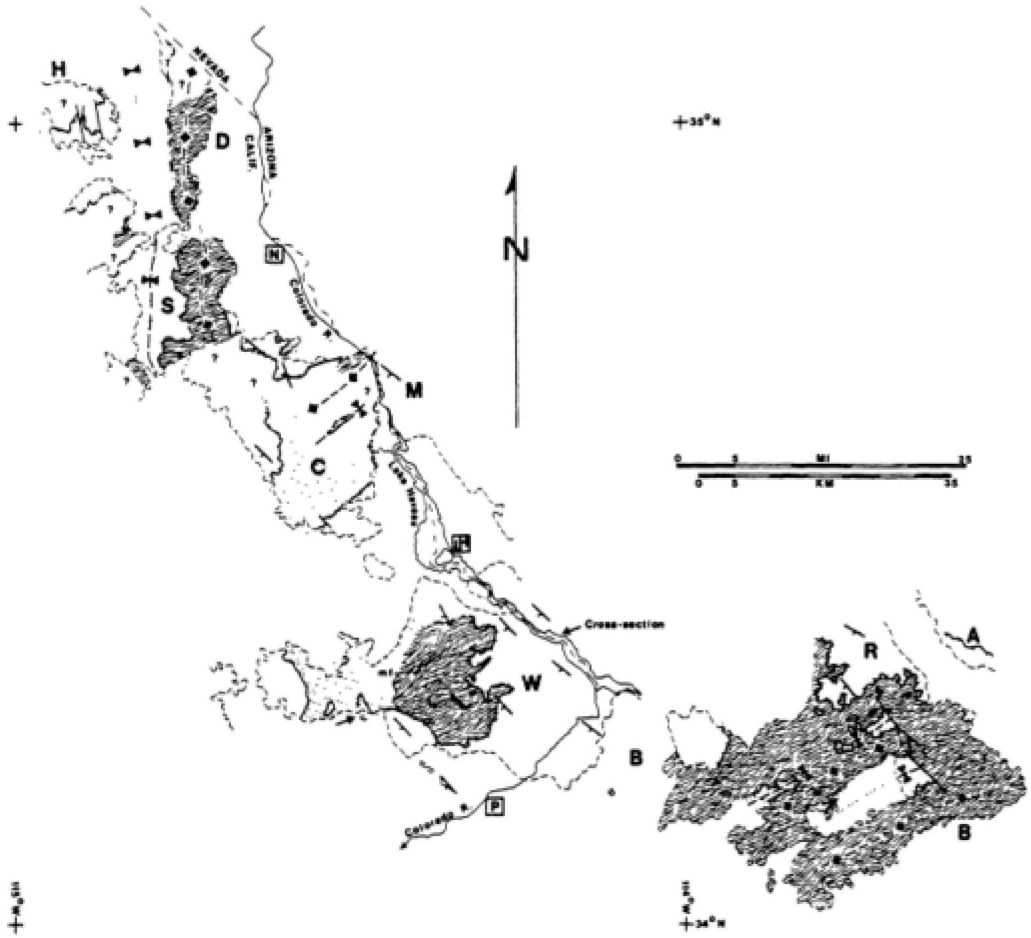


Figure 8.5
Tertiary low-angle fault complex, southeastern California and west central Arizona. Map shows location of regional low-angle fault(s) (hatchured contacts) separating undifferentiated allochthonous upper-plate units (stippled pattern) from autochtonous lower-plate rocks. Lower-plate lithologies include undifferentiated metamorphic and intrusive rocks (short-lined pattern) and their mylonitic equivalents (long-wavy-lined pattern). The two lower-plate assemblages are separated by a "mylonitic front" (mf) in the Whipple Mountains. Heavy dashed lines represent the axial traces of broad antiformal (triangles point outward) and synformal (triangles point inward) folds that warp the basal dislocation fault(s). Communities: N, Needles; LH, Lake Havasu City; P, Parker. Mountain ranges, from NW to SE: H, Homer; D, Dead; S, Sacramento; M, Mojave; C, Chemehuevi; W, Whipple; B, Buckskin; R, Rawhide; A, Artillery. Geological relations north of the Whipple Mountains first ascertained by geologists of the Southern Pacific Land Company (unpublished studies) and confirmed and supplemented by E. Frost and the writer. Geological relations in the Rawhide Mountains from Shackelford (1976a).

Figure 8.6 is a diagrammatic cross section of geological relations in the Whipple Mountains. Mapping in that range indicates that the basal dislocation surface dies out southwestward, that it experienced at least two stages of movement separated by a major interval of erosion and deposition of Miocene sedimentary and volcanic rocks (E. Frost, Univ. of Southern California, personal communication, 1978), and that it has been warped by postfaulting domal uplift of the Whipple Mountains. An earlier suggestion (Davis *et al*., 1977) that low-angle faults in the Colorado River trough may everywhere have developed within a brief, 1- to 2-m.y.-long episode (13-15 m.y. ago) was much too simplistic and is erroneous. Mapping by E. Frost in the southeastern Whipple Mountains reveals that Oligocene or Lower Miocene sedimentary and volcanic strata (Gene Canyon Formation) have been rotated more steeply along northeast-dipping normal faults than unconformably overlying Miocene strata (Copper Basin Formation). Both Frost and Lucchitta and Suneson (1977), the latter in studies between the Whipple and Rawhide Mountains (Figure 8.5), also report sections of rotated Tertiary strata in which the dip of tilted beds decreases progressively upward. These geometric relations are strongly indicative of growth faulting, i.e., Tertiary sedimentation and volcanic activity occurred synchronously with upper-plate normal faulting and the lateral displacement of upper-plate units along the Whipple Mountain basal dislocation surface. Rehrig and Reynolds (in press) reached a similar conclusion in their western Arizona investigations.

The origin of the Colorado River area structures and their relationship, if any, to conventional basin-range faulting in the western United States is unclear. Mountain ranges in the area of Figure 8.5 are definitely not outlined by range-front faults of the type seen in the Great Basin. Significant, but unknown amounts of Miocene extension of upper-plate units apparently occurred throughout the Colorado River trough, but no regional and coeval extensional phenomena have yet been identified in the crystalline rocks of the lower plate(s). Anderson (1971) postulated that extreme "thin-skin distension" at shallow crustal levels in the Eldorado Mountains was compensated at depth by the intrusion of Miocene plutons. But, although present in the Eldorado Mountains, Miocene plutons have not been recognized in the lower plate(s) of the dislocational terrane south of the Eldorado Mountains and, if present, cannot be extensively developed.

It is tempting to speculate that the low-angle fault structures of the Colorado River area may represent the "bottoming-out," as discussed earlier, of typical basin-range faults into a transitional zone between an upper level of brittle deformation and deeper levels of crustal flow. However, field relations do not support this speculation. Mylonitic gneisses occur throughout some lower-plate areas south of the Eldorado Mountains and possess a penetrative NE-SW lineation indication of mylonitic flow in that direction (parallel to upper-plate extension by normal faulting). Field relations in the Whipple-Buckskin-Rawhide Mountains portion of the dislocational terrane clearly indicate that regional mylonitization predated significantly the Miocene dislocational events; tilted and folded mylonitic gneisses are cut with pronounced discordance by the basal dislocation surface in the eastern and central Whipple Mountains. They may have been subject to major erosion prior to Oligocene(?) (or Miocene) deposition and faulting and certainly were being eroded between phases of Miocene faulting (Figure 8.6). Mylonitic cobbles and boulders of lower-plate rock types occur abundantly in some of the Oligocene(?) and Miocene fanglomerate units of the southern and northern Whipple Mountains (both Gene Canyon and Copper Basin Formations). Sphenes from a mylonitic gneiss in one such boulder have yielded an 82.9 ± + 3-m.y.-old fission track age (Dokka and Lingrey, in press). A mylonitic gneiss from the lower plate of the Rawhide Mountains has yielded semiconcordant hornblende and biotite K/Ar ages of 57.4 and 52.3 m.y. ago, respectively (Shackelford, 1977). Furthermore, low-angle faulting and NE-SW extension of upper-plate rocks were not restricted to areas of deeper mylonitic deformation. This is documented in the central Whipple Mountains, where nonmylonitic lower-plate gneisses and plutonic rocks are separated from structurally lower mylonitic

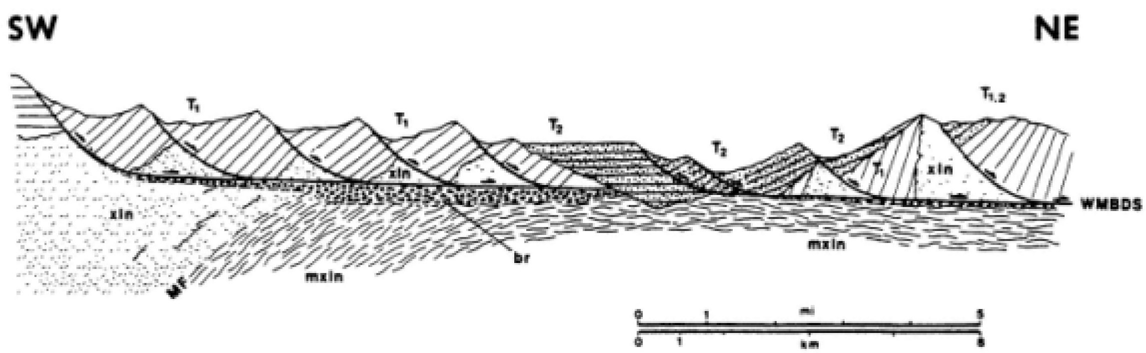


Figure 8.6
Diagrammatic cross section across the Whipple Mountains, southeastern California, illustrating Middle Miocene geological relations prior to domal uplift and warping of the Whipple Mountains basal dislocation surface (WMBDS). The cross section illustrates evidence for two phases of rotational, normal fault displacement along the basal dislocation surface. T_1, older Tertiary sedimentary and volcanic rocks; T_2, younger Tertiary sedimentary and volcanic rocks deposited across the basal dislocation surface prior to their involvement in renewed fault displacement. MF is a "mylonitic front," the abrupt nonfault contact between undifferentiated lower-plate metamorphic and intrusive rocks (xln) and their largely mylonitic equivalents (mxln); br, breccias developed below the basal dislocation surface.

equivalents by an abrupt "mylonitic front" (Figures 8.5 and 8.6); in the Chemeheuvi Mountains (Figure 8.5), where a nonmylonitized Mesozoic(?) quartz diorite comprises most of the lower plate; and in the Eldorado Mountains as described above.

The probable level of erosion in the Colorado River area appears to be much too shallow (<<5 km) to have exposed a brittle-ductile transitional zone of the Great Basin type postulated by Proffett (1977) and others (e.g., Eaton, see Chapter 9). Several lines of evidence suggest that the basal dislocation surface or detachment fault in the Whipple Mountains formed at very shallow crustal levels: (1) the stratigraphic thickness of rotated fault blocks (Tertiary strata plus Precambrian crystalline basement) does not exceed 5 km; (2) Miocene erosion that occurred between two phases of movement on the Whipple basal dislocation surface locally exhumed and breached that fault before deposition of younger, now partly allochthonous strata; and (3) Middle Miocene fanglomerate deposits and interbedded basalt flows (Osborne Wash Formation) partly buried the tilted fault blocks within 3 or 4 m.y. (maximum) after their last movements. These posttectonic surficial deposits locally lie only several hundred meters above the basal dislocation surface and relations (both field and temporal) indicate that the amount of erosion between faulting and fanglomerate deposition was slight.

The origin of the low-angle faults of the Colorado River trough remains obscure. Our studies indicate the detachment of thin upper crustal slabs beneath the area and their synchronous internal distension by normal faulting. Regional gravity sliding of the detached rocks is indicated (a conclusion first reached by Shackelford, 1976a), since lower-plate geological relations are quite diverse and no synchronous lower-plate extensional phenomena of regional extent have been recognized. Lower-plate mylonitic gneisses in portions of the dislocational terrane attest to an earlier phase of NE-SW crustal extension by penetrative, semiductile flow. The tectonic significance and even the age(s) of these puzzling mylonitic rocks remain unresolved; problems of their origin have recently been reviewed by Davis (1977) and Rehrig and Reynolds (1977; in press). It is likely that extensional faulting in the Colorado River trough predated, perhaps significantly, crustal extension by normal faulting in Great Basin areas to the north. A genetic relationship between the two faulted terranes cannot, at present, be demonstrated.

PLATE-TECTONIC SPECULATIONS ON THE CAUSE(S) OF LATE CENOZOIC EXTENSION IN THE WESTERN UNITED STATES

Throughout essentially its entire length, from southeastern Washington to southern Nevada, the interior portions of the western United States experienced variable amounts and styles of E-W to ENE-WSW extension beginning 17-14 m.y. ago. This coincidence in timing applies constraints on plate-tectonic models proposed for the origin of the phenomena. For example, the areas affected by Great Basin normal faulting and the dike swarms that fed Columbia River basalts lay completely north of the Mendocino triple junction at this time (Snyder *et al*., 1976). Thus, extensional tectonics cannot be related simply to northward migration of the Mendocino triple junction as Atwater (1970) originally suggested, a conclusion also reached by others for the transition in the western United States from calc-alkaline, mainly andesitic volcanism to basaltic and subordinate silicic volcanism (Snyder *et al*., 1976; Cross and Pilger, 1978; Lipman, see Chapter 14).

Atwater (1970) proposed that extension in the Basin and Range province was related to oblique (NW-SE) rifting or "stretching" within a broad transform zone that developed south of the Mendocino triple junction and east of the Pacific-North American transform plate boundary. Two lines of evidence refute this suggestion: (1) the relation already mentioned that at the time of inception of normal faulting in the Great Basin (ca. 17 m.y. ago) almost all of the region lay north of the triple junction and, therefore, east of a convergent plate boundary; and (2) increasing kinematic evidence that initial extension across normal faults and dikes in the Great Basin occurred in an E-W to ENE-WSW direction, not along the NW-SE trend as postulated by Atwater. E-W to ENE-WSW distension is defined by the orientation of the Garlock transform fault at the southern end of the Great Basin (Davis and Burchfiel, 1973), by the orientation of the Pahranagat shear system in southeastern Nevada, and by the orientations of a 17- to 14-m.y.-old dike swarm in north-central Nevada and a small orthogonal transform fault associated with it (N 70º E; Zoback and Thompson, 1978).

At present, much of the western Great Basin does appear to be extending in a NW-SE direction parallel to the San Andreas plate boundary to the west. N-S faults in affected areas have typically exhibited oblique slip (slip with both normal, or dip-slip, and right-slip components) during historic seismic events (Gumper and Scholz, 1971). The eastern Great Basin, however, does not apparently "feel" a superposed transform boundary stress regime and continues to extend in an E-W direction (Smith and Sbar, 1974).

Recent workers favor the idea that initial extension and magmatism within the Great Basin area is a back-arc phenomenon (e.g., Scholz *et al*., 1971; Snyder *et al*., 1976; Zoback and Thompson, 1978), perhaps related to the diapiric rise and lateral spreading of North American plate asthenosphere that was heated at depth above a subducting plate. Similarly, the location of the Columbia River basalts east of the Cascade Range (Figure 8.2) and their temporal equivalency to early phases of Basin-Range faulting in areas to the south have suggested to most workers that they, too, are a back-arc response to subduction of oceanic lithosphere beneath the western edge of North America (e.g., Snyder *et al*., 1976; McDougall.

1976; and Christiansen and Lipman, 1972). McBirney *et al*. (1974) have presented data in support of an association of the Columbia River basaltic eruptions with Cascade arc activity. The former coincides impressively, at least in the central Oregon Cascades, with the most intense pulse of andesitic volcanic activity in the Cenozoic period (16.6-14.2 m.y. ago). McBirney *et al*. (1974) state that a coeval mid-Miocene pulse of arc volcanism may also be present in the Indonesian and other circum-Pacific areas. Accelerated global rates of plate convergence in mid-Miocene time may be indicated, and it is perhaps in this general plate context that the Columbia River dike swarms and basalt outpouring should be viewed.

An additional or alternative explanation for Miocene back-arc extension in the western United States is provided by the fact that this extension followed a probable steepening of the Farallon-Juan de Fuca plate approximately 20 m.y. ago. Snyder *et al*. (1976) presented data indicating an east-west narrowing of the volcanic arc in the Pacific Northwest at this time. A more pronounced narrowing occurred in areas to the south. Since the width of a volcanic arc is related to the dip of the subducting plate beneath it, the wider the arc, the lower the inclination of the underthrusting slab. Changes in dip of a sub-ducting plate may govern, or at least influence, stress state in the overlying plate [other factors, e.g., convergence rate, absolute motions of upper plate, and age of descending oceanic plate, are discussed by Cross and Pilger (1978)]. Subduction relationships between the Nazca plate and the South American plate are instructive in this regard. At present, the Nazca plate is being subducted eastward beneath northern and central Peru at a low angle (10-15º; Stauder, 1975; Megard and Philip, 1976). Above this plate the Andean volcanic are is inactive and the stress in the South American plate is characterized by east-west compression. In contrast, beneath southern Peru and northern Chile the Nazca plate dips more steeply (ca. 30º), the upper plate is affected by predominantly extensional tectonics acting perpendicularly to the trench, and the Andean calc-alkaline volcanic chain is active. It is thus conceivable that back-arc extension in the western United States was triggered by (1) an accelerated rate of plate convergence and (2) by steepening of the Farallon-Juan de Fuca plate approximately 20 m.y. ago, with the resultant initiation of an upper-plate stress field favoring ENE-WSW dilation in the Columbia Plateau and Basin and Range provinces.

PERSPECTIVES

It should be evident from this brief review of Cenozoic intraplate extension in the western United States that much work remains to be done and more data need to be gathered before even the structural phenomena are well understood. The problems are not just why extension occurred in the intraplate setting and which plate-tectonics explanations are most satisfying. Before we can hypothesize "why" we must understand the basic geometries and kinematics of "how" extension took place. Much more field work is needed in the extensional terranes of the western continent to resolve these first-order problems. The recognition in only the past 8 years (beginning with Anderson, 1971) of an entire province in southern Nevada, southeastern California, and western and southern Arizona, of late Cenozoic low-angle normal faults with an as-yet cryptic origin, is just one indication of how far *field studies have lagged behind plate-tectonics speculations*. The recently published study of the Yerington district of western Nevada by Proffett (1977) is an admirable example of how much new data and field-based interpretation can come from detailed mapping studies in the western states. Proffett's conclusions were not based solely on field mapping but were supplemented by a total of 30,000 m of drill-hole data, cores, and cuttings. His work indicates the importance of subsurface information in identifying the geometry of normal faulting within the Basin and Range province. The need for deep seismic reflection data in the province is obvious, particularly for resolving the problems of normal fault geometry at depth. And finally, the possible existence of a newly discovered low-velocity zone in the upper crust of the eastern Great Basin is but one impressive indication of how much fundamental geophysical data are still needed in order to understand the physical state of the lithosphere in western North America.

ACKNOWLEDGMENTS

J. L. Anderson, A. W. Bally, Gordon Eaton, Eric Frost, and Warren Hamilton kindly suggested improvements in several drafts of this paper. Portions of this summary concerning extensional tectonics in the Pacific Northwest were originally prepared for the Washington Public Power Supply System (WPPSS). David Tillson, of the WPPSS, is thanked for his interest in and support of these studies. Research by the writer and his associates in the Whipple-Buckskin-Rawhide Mountains area described herein has been supported by the National Science Foundation, Earth Science Grants GA-43309 and EAR77-09695.

References

Anderson, R. E. (1971). Thin skin distension in Tertiary rocks of southeastern Nevada, *Geol. Soc. Am. Bull. 82*, 43-58.

Armstrong, R. L. (1972). Low-angle (denudation) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah, *Geol. Soc. Am. Bull. 83*, 1929-1954.

Armstrong, R. L., W. H. Taubeneck, and P.O. Hales (1977). Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho, *Geol. Soc. Am. Bull. 88*, 397-411.

Atwater, T. (1970). Implications of plate tectonics for the Cenozoic evolution of western North America, *Geol. Soc. Am. Bull. 81*, 3513-3536.

Carr, W. J., and D. D. Dickey (1977). Cenozoic tectonics of eastern Mojave Desert, *U.S. Geol. Surv. Prof. Paper 1000*, 75 pp.

Christiansen, R. L., and P. W. Lipman (1972). Cenozoic volcanism and plate-tectonic evolution of the western United States. II. Late Cenozoic, *Phil. Trans. Roy. Soc. London A271*, 249-284.

Clague, J. J. (1974). The St. Eugene Formation and the development of the southern Rocky Mountain Trench, *Can. J. Earth Sci. 11*, 916-938.

Compton, R. R., V. R. Todd, R. E. Zartman, and C. W. Naeser (1977). Oligocene and Miocene metamorphism, folding and low-angle faulting in northwestern Utah, *Geol. Soc. Am. Bull. 88*, 1237-1250.

Coney, P. J. (1970). The geotectonic cycle and the new global tectonics, *Geol. Soc. Am. Bull. 81*, 739-747.

Coney, P. J. (1974). Structural analysis of the Snake Range "décollement," east-central Nevada, *Geol. Soc. Am. Bull. 85*, 973-978.

Cross, T. A., and R. H. Pilger (1978). Constraints on absolute motion and plate interaction inferred from Cenozoic igneous activity in the western United States, *Am. J. Sci. 278*, 865-902.

Davis, G. A., and B. C. Burchfiel (1973). Garlock Fault: an intracontinental transform structure, southern California, *Geol. Soc. Am. Bull. 84*, 1407-1422.

Davis, G. A., V. K. Evans, E. G. Frost, S. H. Lingrey, and T. J. Shackelford (1977). Enigmatic Miocene low-angle faulting, southeastern California and west-central Arizona—supra-structural tectonics?, *Geol. Soc. Am. Abstr. Programs 9*, 933-934.

Davis, G. H. (1975). Gravity-induced folding off a gneiss dome complex, Rincon Mountains, Arizona, *Geol. Soc. Am. Bull. 86*, 979-990.

Davis, G. H. (1977). Characteristics of metamorphic core complexes, southern Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 934.

Dokka, R. K., and S. H. Lingrey (in press). Fission track evidence for a Miocene cooling event, Whipple Mountains, southeastern California, in *Cenozoic Paleogeography of the Western United States*, Pacific Section, Society of Economic Paleontologists and Mineralogists.

Donath, R. A. (1962). Analysis of basin-range structure, south-central Oregon, *Geol. Soc. Am. Bull. 73*, 1-16.

Fox, K. F., Jr., C. D. Rinehard, and J. C. Engels (1977). Plutonism and orogeny in north-central Washington—timing and regional context, *U.S. Geol. Surv. Prof. Paper 889*, 27 pp.

Gumper, F. J., and C. Scholz (1971). Microseismicity and tectonics of the Nevada Seismic Zone, *Bull. Seismol. Soc. Am. 61*, 1413-1432.

Hamilton, W. (1978). Mesozoic tectonics of the western United States, in *Mesozoic Paleogeography of the Western United States*, Pacific Section, Society of Economic Paleontologists and Mineralogists, pp. 33-70.

Hamilton, W., and W. B. Myers (1966). Cenozoic tectonics of the western United States, *Rev. Geophys. 4*, 509-549.

Hose, R. K., and Z. F. Dane (1973). Development of the late Mesozoic to early Cenozoic structures in the eastern Great Basin, in *Gravity and Tectonics*, K. A. De Jong and R. Scholton, eds., John Wiley, New York, pp. 429-442.

Kistler, R. W., and Z. E. Peterman (1978). Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks, *U.S. Geol. Surv. Prof. Paper 1071*, 17 pp.

Lawrence, R. D. (1976). Strike-slip faulting terminates the Basin and Range province in Oregon, *Geol. Soc. Am. Bull. 87*, 846-850.

Liggett, M. A, and H. E. Ehrenspeck (1974). Pahranagat shear system, Lincoln County, Nevada, Argus Exploration Co. Report of Investigation, *NASA 5-21809*, 6 pp.

Lingrey, S. H., K. V. Evans, and G. A. Davis (1977). Tertiary denudational faulting, Whipple Mountains, southeastern San Bernardino County, California, *Geol. Soc. Am. Abstr. Programs 9*, 454-455.

Longwell, C. R. (1945). Low-angle normal faults in the Basin-and-Range province, *Eos Trans. Am. Geophys. Union 26*, 107-118.

Lucchitta, I., and N. Suneson (1977). Cenozoic volcanism and tectonism, west-central Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 457-458.

McBirney, A. R., J. F. Sutter, H. R. Naslund, K. G. Sutton, and C. M. White (1974). Episodic volcanism in the central Oregon Cascade Range, *Geology 2*, 585-590.

McDougall, I. (1976). Geochemistry and origin of basalt of the Columbia River group, Oregon and Washington, *Geol. Soc. Am. Bull. 87*, 777-792.

Megard, F., and H. Philip (1976). Plio-Quaternary tectonomagmatic zonation and plate tectonics in the central Andes, *Earth Planet. Sci. Lett. 33*, 231-238.

Misch, P. (1960). Regional structural reconnaissance in central-northeast Nevada and some adjacent areas: observations and interpretations, in *Intermountain Association of Petroleum Geologists 11th Annual Field Conference Guidebook*, pp. 17-42.

Misch, P. (1971). Geotectonic implications of Mesozoic décollement thrusting in parts of eastern Great Basin, *Geol. Soc. Am. Abstr. Programs 3*, 164-166.

Proffett, J. M., Jr. (1977). Cenozoic geology of the Yerington district, Nevada, and implications for the nature and origin of Basin and Range faulting, *Geol. Soc. Am. Bull. 88*, 247-266.

Rehrig, W. A., and S. J. Reynolds (1977). A northwest zone of metamorphic core complexes in Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 1139.

Rehrig, W. A., and S. J. Reynolds (in press). Geologic and geochronologic reconnaissance of a northwest-trending zone of metamorphic complexes in southern Arizona, *Geol. Soc. Am. Mem*.

Reynolds, M. W. (1974). Geology of the Grapevine Mountains, Death Valley, California: a summary, in *Guidebook: Death Valley Region, California and Nevada*, Death Valley Publishing Co., Shoshone, Calif., pp. 91-96.

Roberts, R. J. (1972). Evolution of the Cordilleran fold belt, *Geol. Soc. Am. Bull. 83*, 1989-2004.

Scholz, C. H., M. Barazangi, and M. L. Sbar (1971). Late Cenozoic evolution of the Great Basin, western United States, as an ensialic interarc basin, *Geol. Soc. Am. Bull. 82*, 2979-2990.

Shackelford, T. J. (1976a). Structural geology of the Rawhide Mountains, Mohave County, Arizona, U. Southern Calif., unpubl. Ph.D dissertation, 175 pp.

Shackelford, T. J. (1976b). Juxtaposition of contrasting structural and lithologic terranes along a major Miocene gravity detachment surface, Rawhide Mtns., Arizona, *Geol. Soc. Am. Abstr. Programs 8*, 1099.

Shackelford, T. J. (1977). Late Tertiary tectonic denudation of a Mesozoic(?) gneiss complex, Rawhide Mountains, Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 1169.

Smith, R. B. (1978). Seismicity, crustal structure and intraplate tectonics of the western Cordillera, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. 152*, 111-144.

Smith, R. B., and M. L. Sbar (1974). Contemporary tectonics and seismicity of the western United States with emphasis on the Intermountain Seismic Belt, *Geol. Soc. Am. Bull. 85*, 1205-1218.

Snavely, P. D., Jr., N. S. MacLeod, and H. C. Wagner (1973). Miocenetholeiitic basalt of coastal Oregon and Washington and their relation to coeval basalt of the Columbia Plateau , *Geol. Soc. Am. Bull. 84*, 387-424.

Snyder, W. S., W. R. Dickinson, and M. L. Silberman (1976). Tectonic implications of space-time patterns of Cenozoic volcanism in the western United States, *Earth Planet. Sci. Lett. 32*, 91-106.

Stauder, W. (1975). Subduction of the Nazca plate under Peru as evidenced by focal mechanism and by seismicity, *J. Geophys. Res. 80*, 1053-1064.

Stewart, J. H. (1971). Basin and Range structure: a system of horsts and grabens produced by deep-seated extension, *Geol. Soc. Am. Bull. 82*, 1019-1044.

Stewart, J. H., G. W. Walker, and F. J. Kleinhampl (1975). Oregon-Nevada lineament, *Geology 3*, 265-268.

Swanson, D. A., and T. L. Wright (1977). The Columbia River basalt, product of a major thermal event in the western Cordillera, *Geol. Assoc. Can. Program Abstr. 2*, 50.

Terry, A. H. (1972). The geology of the Whipple Mountains thrust fault, southeastern California, Calif. State U. at San Diego, unpubl. M.S. thesis, 90 pp.

Thompson, G. A., and D. B. Burke (1974). Regional geophysics of the Basin and Range Province, *Ann. Rev. Earth Planet. Sci. 2*, 213-238.

Watkins, N. D., and A. K. Baksi (1974). Magnetostratigraphy and oroclinal folding of the Columbia River, Steens, and Owyhee basalts in Oregon, Washington, and Idaho, *Am. J. Sci. 274*, 148-189.

Whetten, J. T. (1976). Tertiary sedimentary rocks in the central part of the Chiwaukum graben, Washington, *Geol. Soc. Am. Abstr. Programs 8*, 420.

Wright, L. A., and B. W. Troxel (1973). Shallow-fault interpretation of Basin and Range structure, southwestern Great Basin, in *Gravity and Tectonics*, K. A. De Jong and R. Scholton, eds., John Wiley, New York, pp. 397-407.

Zoback, M. L., and G. A. Thompson (1978). Basin and Range rifting in northern Nevada: clues from a mid-Miocene rift and its subsequent offsets, *Geology 6*, 111-116.

9

Geophysical and Geological Characteristics of the Crust of the Basin and Range Province

Gordon P. Eaton
U.S. Geological Survey

INTRODUCTION

Selected geophysical and geological characteristics of the Basin and Range province of the western United States are examined here. They represent the effects of nearly 30 million years (m.y.) of crustal extension, preceded by subduction-related compressional deformation and magma genesis. The effects of this compressional regime appear to have exerted a significant influence on the mechanics and locus of late Cenozoic extension.

In its present state, the crust of the Basin and Range province is atypical of most continental crust, being appreciably thinner, warmer, and more highly fractured and permeable. Some of what is seen or measured geophysically at the surface in the province is influenced by processes or phenomena in the shallow crust. Faulting, for example, appears to be a condition of the upper 10 to 15 km, and some of the phenomena associated with it serve to screen or distort information from deeper levels. As an example, convection of groundwater in the fractured, shallow crust complicates the determination of heat flow from deeper crustal levels and is apparently also capable of creating regional geoelectric anomalies.

One of the remarkable features of the Basin and Range province is the broadly distributed nature of its extension. Most other regions of continental extension consist of singular or branching large rifts, like the contemporary, neighboring Rio Grande rift in New Mexico or the Rhine graben of Europe. Although marginal seas bordering continents, like the Seas of Okhotsk or Japan [both of which are at present inactive with regard to crustal spreading but nevertheless have high heat flow (Karig, 1974)], have a similarity to the Basin and Range province, they have significant contrasts as well. The same is true for actively spreading intraoceanic, back-arc basins, like the Bonin, Mariana, New Hebrides, and Lau-Havre troughs.

The Great Basin consists of continental crust, stands nearly 2 km above sea level, has a very broad, but well-developed. bilateral symmetry in its geophysical characteristics (Eaton *et al*., 1978), and has a perimeter marked by active crustal seismicity and young volcanism. In contrast, marginal seas near the Asian continent and intraoceanic spreading basins consist of oceanic crust, the upper surface of which is submerged *below* sea level. They show little or no distributed or peripheral crustal seismicity. (see Barazangi and Dorman, 1969). No well-developed or extensive pattern of symmetry nor any persistent, peripheral volcanism has been described for them. Although paired linear magnetic anomalies exist locally, they are relatively incoherent compared with those at spreading ocean ridges. and their amplitudes are much smaller.

Extension in the Great Basin *began* within and behind an andesitic volcanic are, typical of where back-are basins form and exist. Although much of that are is now dead, extensional deformation continues. Finally, active continental volcanic arcs persist today in the Cordillera of Central and South America but have no Basin-Range structure behind them. According to Uyeda and Kanamori (1979), this is probably the result of a compressional state of stress in the overriding American plate; it follows from a relatively low angle of subduction and rapid rate of plate convergence. A fundamental distinction is made between what they term a "Chilean," or convergence-related compressional mode, and a "Marianas," or convergence-related extensional mode, of subduction.

Interpretations of crustal thickness of different parts of the western United States vary somewhat (compare Prodehl, 1970; Warren and Healey, 1973; and Smith, 1978), but it appears that the Great Basin crust is no more than 25-30 km thick. The crust of the rest of the Basin and Range province ranges from 20 to 30 km thick. By contrast, the crust of the unextended Colorado Plateaus and Great Plains provinces to the east is 40 and 50 km thick, respectively. Upper-mantle velocities beneath the Great Basin are anomalously low (Smith, 1978) and the lithosphere is anomalously thin ($\leq$65 km). It is one of the few *continental* areas of the globe beneath which a low-velocity zone is known to be present in the mantle.

Not all of the Basin and Range province is tectonically active at present. The Great Basin section is extending, the western Mojave Desert is in horizontal dextral shear, and the rest of the province is tectonically inactive and has been for several millions of years.

PRESENT AND PAST CRUSTAL EXTENSION

Normal faults, the surface evidence of crustal extension, are widespread in the western United States. Their distribution in the Basin and Range province is broad and pervasive, but in the Rio Grande rift [see Figures 9.5(a) and 9.6 for identification of the principal geographic features mentioned in the text] they define a narrow band typical of continental grabens. On the basis of the geomorphology of the faulted region, crustal spreading is a continuing process, still active in some areas, but inactive in others. The Sonoran Desert section of the Basin and Range province has a general elevation and topography characteristic of profound erosion and tectonic inactivity (Fenneman, 1931; Lobeck, 1939; Hunt, 1967). In sharp contrast, Quaternary faults of Nevada are amenable to detailed age classification based on the degree of erosion, the slope angle of the scarp, and the width of the crestal break in slope (Slemmons, 1967; Wallace, 1977).

Figure 9.1 is a map of the contemporary fault-displacement field of the United States. Active normal-fault displacements are restricted almost exclusively to the Great Basin and Rio Grande rift. Local, scattered thrusting events in the West are seen to the northwest, east, and southwest, but displacements of this type are found mostly in the middle and eastern United States.

The azimuths of the *P* (pressure) and *T* (tension) axes cannot be equated with those of the principal stress directions, σ_1 and σ_3, with certainty. McKenzie (1969) demonstrated that in the general case of triaxial stress the only restriction on the spatial relation between the two is that the stress directions must lie in quadrants containing the related displacement axes but may diverge from them by many tens of degrees. He argued, as did Brace (1972a), that because most crustal earthquakes occur on preexisting fault planes, they are unrepresentative of the ideal situation in which initial failure takes place in virgin, homogeneous material. Only in this case will stress directions necessarily bear direct correspondence to displacement. Because of the long history of normal faulting in the Great Basin we may assume that those earthquakes whose fault-plane solutions are shown in Figure 9.1 took place on pre-existing faults and that the orientation of these faults with respect to the stress field determined both the direction and nature of displacement.

To investigate the potential discrepancy between displacement and stress direction, *in situ* measured directions of σ_1 for the eastern United States were plotted in Figure 9.1. Together, the two kinds of data suggest that on a regional basis *P* and σ_1 have much the same orientation—northeast to east.

In a related plot (Figure 9.2), a comparison is made between locally paired, measured directions of *T* and σ_3 in the West. Most of the directions of σ_3 determined from *in situ* stress measurements are not shown in Figure 9.1 because of the density of other data. They were taken from sources listed in the caption. The data selection process required only that the earthquake and *in situ* stress measurement locality in each pair be within 200 km of one another.

Figure 9.2 indicates moderately good agreement between extensional stresses and displacement directions, despite McKenzie's (1969) well-reasoned caution. On this limited basis we may assume that the displacement field of the normal faults is a measure of the direction of current extension. On average, it is east-west. The strike of these faults in relation to that of the displacements is, in much of the region, orthogonal or nearly so, hence the spreading is generally in consonance with that of continental spreading elsewhere in the world (Ranalli and Tanczyk, 1975) and with that of seafloor spreading (Moore, 1973).

Figure 9.1 shows two other significant characteristics of the region of active continental spreading: (1) it stands high, having been uplifted to an elevation of 1100 m and more; and (2) it is hot; heat-flow values nearly everywhere are greater than 1.1 μcal cm^{-2} sec^{-1}, the average value for stable continents. We return to these observations below.

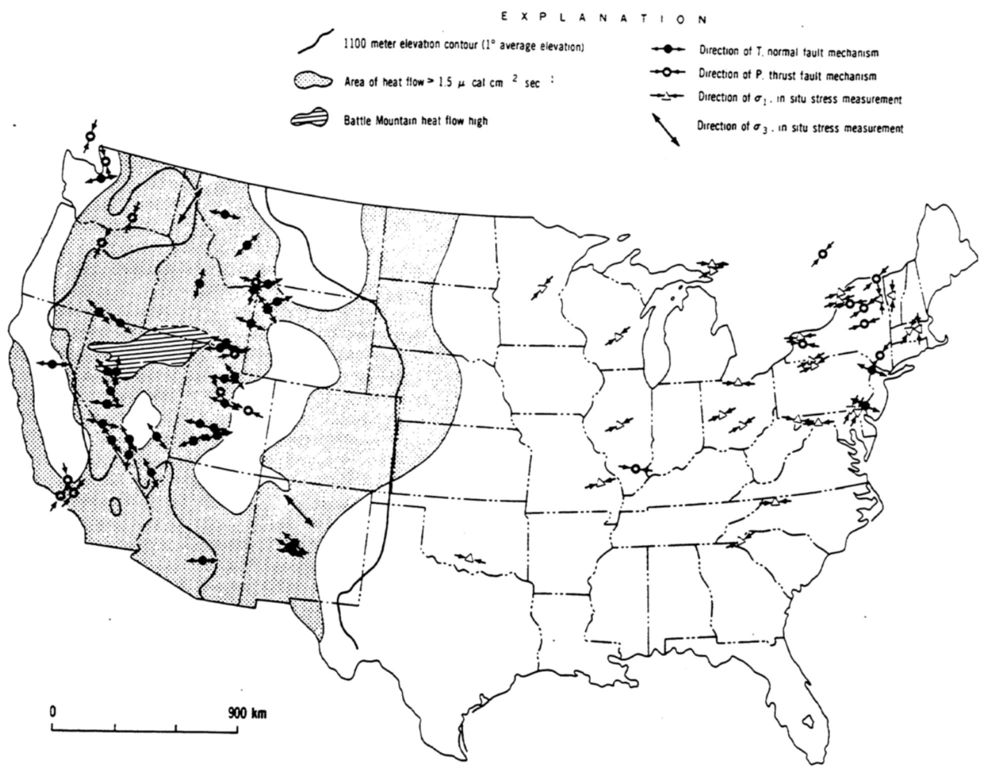


Figure 9.1
Horizontal components of fault displacements and maximum principal stress directions. Compiled from data in Couch et al. (1976), Haimson (1977), Jaksha et al. (1977), Lachenbruch and Sass (1977), Malone et al. (1975), Mott (1976), Rogers (1977), Sanford et al. (1977), Sbar and Sykes (1973, 1977), Smith (1978), Smith and Lindh (1978), and Strange and Woollard (1964). Thrust faulting and horizontal compression dominate in the eastern United States, normal faulting and extension, in the west.

Figure 9.3(a) is a map of most of the major faults in western North America that have steep dips, regardless of age. The longer ones are strike-slip faults, the shorter ones, normal faults. The latter are much more abundant and widespread. *Active* strike-slip faults that have major displacements of several tens of kilometers are today restricted largely to a 500-km-wide corridor paralleling the west coast of the United States. Those farther inland, in Canada, have not been active in late Cenozoic time—the period of crustal spreading with which we are concerned.

Because the azimuthal range of these faults is great, it is difficult to see well-defined patterns from which we might draw immediate conclusions about variations in spreading. To do this one must filter the data. Figure 9.3(b) is temporally filtered, showing only those faults active in the past 10-15 m.y. for which some Quaternary movement is suspected. It distinguishes the Great Basin section of the Basin and Range province from the Sonoran Desert section. It also clearly defines the Rio Grande rift. The fault data, like the geomorphic data, suggest that extensional spreading in the Sonoran Desert region ended some time ago. Eberly and Stanley (1978) alleged that block faulting began to wane about 10.5 m.y. ago in the Sonoran Desert and that throughgoing drainage of the Gila River was well established by 6.0 m.y. ago. The cessation of crustal spreading at these latitudes may be related to the opening of the Gulf of California on the west. If so, it represents the continental equivalent of a kind of ridge jump—from distributed extension on the east to narrowly focused extension on the west.

Figure 9.4 shows that the orientation of basin ranges in the Sonoran Desert is significantly different from that of the Great Basin and Rio Grande rift. If the topography is a reflection of block faulting, this difference, in conjunction with the data of Figures 9.1 and 9.2, suggests that extensional stresses had a markedly different orientation in the Sonoran Desert region than they do in the Great Basin today. They were generally southwest to west-southwest, in contrast to later east-west to west-north-west extension in the rest of the western United States.

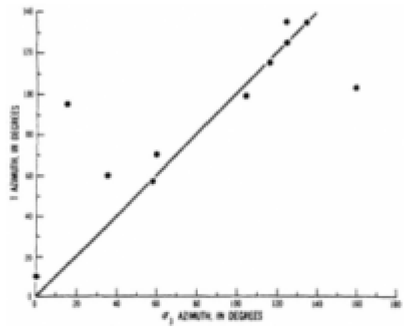


Figure 9.2
Azimuthal relations between horizontal components of minimum principal stress (σ_3) and T axes of normal-fault earthquake mechanisms. Localities in each set of paired measurements (stress measurement and fault displacement) were less than 200 km apart. Data from sources listed in caption of Figure 9.1. Solid line defines locus of stress directions and fault displacements having the same horizontal direction.

THE THERMAL REGIME

Heat lost at the surface of the earth may reflect any of several phenomena: (1) conduction of heat into and through the lithosphere, the base of which may approximate an isothermal surface marking the upper bound of a region of partial melt in the mantle; (2) upward mass transport of heat through the lithosphere by ascending magma, a penetrative form of convection; (3) production of heat in the crust itself by radioactive decay of U, Th, and ^{40}K and/or by conversion from mechanical energy, as in faulting; (4) transient cooling of the lithosphere; and (5) convection of heat in the shallow crust by circulating groundwaters. The last phenomenon is especially troublesome in permeable rocks. It can greatly disturb the heat flow associated with deeper crustal regimes.

Many heat-flow measurements have been made in the extended regions of the western United States (Lachenbruch and Sass, 1977). The average value is roughly twice that of the tectonically stable interior and eastern part of the North American continent, after accounting for crustal radioactivity. Although heat-flow values in the Pacific Northwest are also higher than those in the continental interior, the reduced values (those for which the contribution of heat from heat production within the crust has been subtracted) appear to be somewhat lower than those of the Great Basin (compare Figures I and 14 of Lachenbruch and Sass, 1977). At present, there are an insufficient number of heat-flow measurements in the Sonoran Desert to be certain of what its thermal regime is, but thermal lag may keep crustal temperatures moderately high if extension there ceased as recently as 7 m.y. ago. The Rio Grande rift has heat-flow values as high as those of the Great Basin (Reiter *et al*., 1975), and they decay systematically outward, much as they do at spreading ocean ridges.

Lachenbruch and Sass (1977; 1978) examined the thermal regime of the U.S. continental crust, evaluated various factors affecting crustal temperature, and proposed a model in which the high, but variable, heat flow in the Great Basin is accounted for by (1) regionally distributed basaltic intrusions of the lithosphere, accompanied by lithospheric thinning and possibly magmatic underplating and (2) spatial variations of the extensional strain rate that controls access of this basalt to the lithosphere. The lithosphere is pulled apart and basalt wells up into it from the asthenosphere.

Basaltic intrusions in this model are viewed as vertical, dike-like or bleb-like bodies that accommodate extension while preserving the continuity of the lithosphere. Instead of pulling open at a single place, as it does at most spreading ridges, the lithosphere pulls open in a broadly distributed fashion at a great many places. Spatial and temporal variations in rates of extension control the intensity of the upward flux of basalt and, hence, that of surface heat flow. The model is mechanically somewhat like one proposed by Thompson (1959; 1966), who postulated a Great Basin lithosphere dilated by intrusive dikes. It provides a rationale for the local occurrence of shallow magmatic systems that convey heat from the asthenosphere to the surface by mass transport.

CRUSTAL MAGMAS

The Mesozoic and Cenozoic history of the western United States is one of repeated intrusion of the shallow crust by magmas. Many broke through to the surface, creating large volcanic fields. Figure 9.5 shows the spatial distribution of igneous rocks of post-Paleozoic age. From earliest Mesozoic through early Miocene time, rocks of intermediate, calc-alkaline composition were formed, probably as a result of subduction of the Farallon plate beneath the region (Lipman *et al*., 1972; see Chapter 14). From Miocene time on, however, they were dominantly bimodal (basaltic and rhyolitic) in composition, reflecting extension of the lithosphere that followed cessation of subduction (Christiansen and Lipman, 1972; Christiansen and McKee, 1978). Figure 9.5(a) shows the regional limits of Mesozoic igneous rocks and individual areas of outcrop of Paleocene through Oligocene igneous rocks. Clearly, the Great Basin had abundant igneous activity in Mesozoic and early Tertiary time.

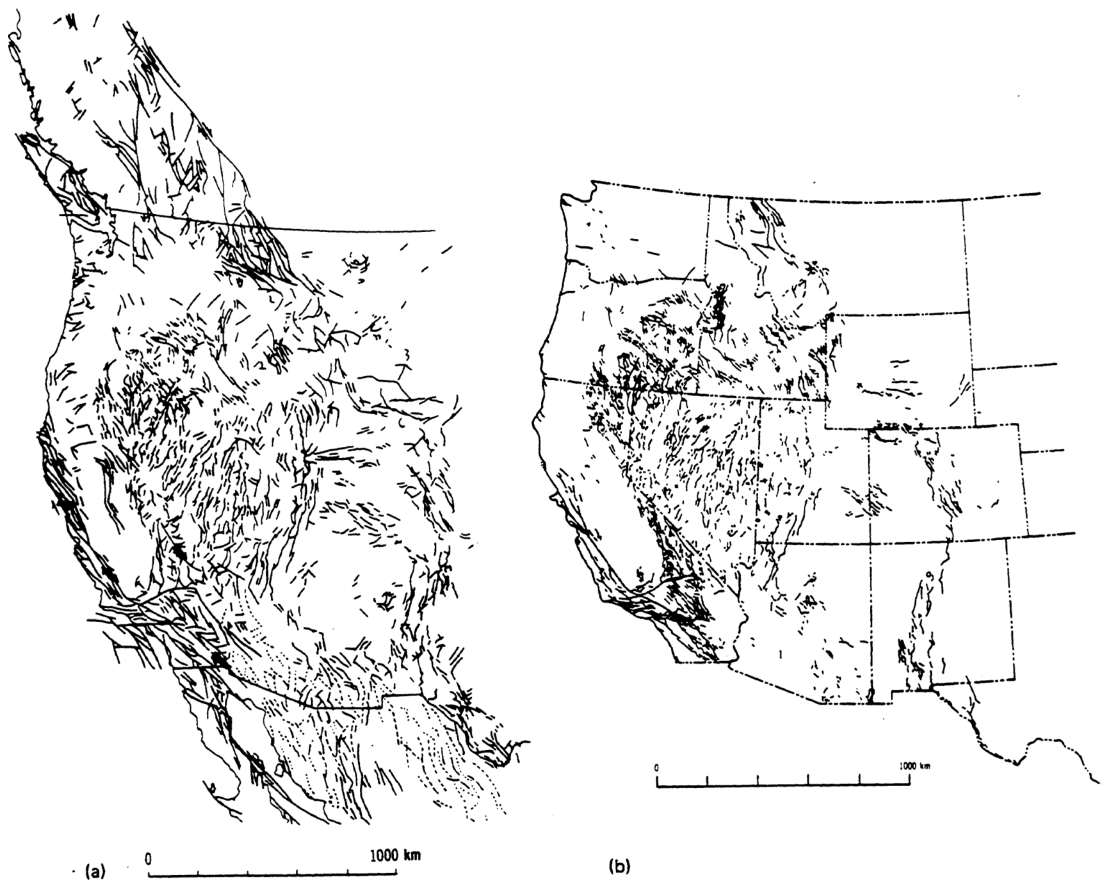


Figure 9.3
(a) Steeply clipping faults in western North America (from King and Beikman, 1974; King, 1969; and Cohee, 1961). Faults are shown without regard to age of initiation or last movement. Longer ones are generally strike-slip faults; shorter ones, normal faults. Dashed lines in southern part of map mark long axes of block ranges whose boundary faults are obscured by upper Cenozoic alluvium. (b) Faults active in the past 10-15 m.y. for which some Quaternary movement is suspected (after Howard et al., 1978).

Igneous rocks of Miocene, Pliocene, and Quaternary age are shown in Figures 9.5(b), 9.5(c), and 9.5(d). An outward restriction of magmatic activity with time toward the margins of the Great Basin is illustrated by these three diagrams. Areas immediately adjacent to the Great Basin had a generally similar history of magmatism, as the four illustrations show. Magmatism alone does not set the Great Basin or Rio Grande rift apart from the rest of the region.

Figure 9.6 shows the spatial distribution of known Cenozoic plutons and ash flow-related calderas. The calderas represent large magma chambers at crustal levels shallow enough to permit catastrophic eruptions of great volumes of pyroclastic material. Such magma chambers are in the upper 5 to 10 km of the crust and give up large amounts of heat there. More than 50 of these features have been identified thus far in the Great Basin and the region immediately north of it. Another 22 are seen in a broad north-south corridor that includes the Rio Grande rift and its western environs.

The distribution of these shallow crustal heat sources coincides generally with that of normal faults active during the past 10 m.y. to 15 m.y. [Figure 9.3(b)]. Although the former span the entire Cenozoic Era, they are spatially distributed more or less like the regions of young extensional faulting, suggesting some sort of genetic relationship. Mackin (1960) proposed direct cause and effect:

withdrawal of large volumes of magma to the surface allowed collapse and spreading of the overlying slab. I do not subscribe to this view.

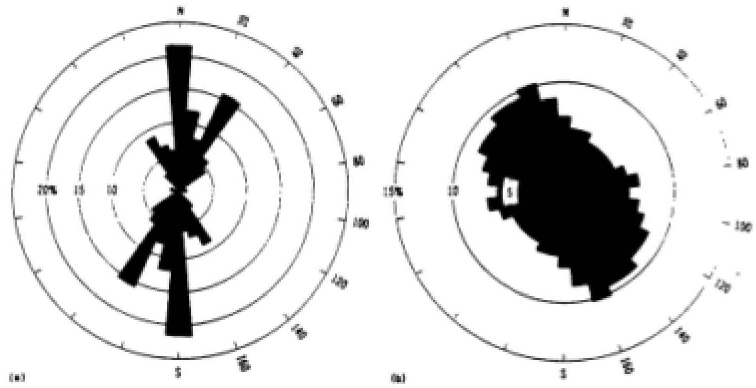


Figure 9.4
Rose diagrams of the orientations of range fronts. (a) 545 range-front segments, 50 km long, in the Great Basin section of the Basin and Range province and the Rio Grande rift (Nevada, Utah, and northern and central New Mexico); (b) 548 range-front segments, 50 km long, in the Mojave Desert. Sonoran Desert, and Mexican Highlands section of the Basin and Range province (southeastern California, southern Arizona, and southern New Mexico). Note difference in maximum frequency values of the outermost rings of the two diagrams and the greater scatter of range-front directions in (b).

HYDROTHERMAL CONVECTION

Heated groundwaters appear at the surface as hot springs and warm springs, the surface manifestations of convecting hydrothermal systems. In some places they are related directly to shallow bodies of magma, as at Yellowstone National Park (Eaton *et al*., 1975; Smith and Shaw, 1975). They mark sites where the regional conductive heat flow is perturbed. The areal distribution of thermal springs is shown in Figure 9.7(a), where boundaries have been drawn about regional clusters of springs. The guideline followed in drawing the boundaries required that individual members of a cluster should be no more than 100 km from one of its neighbors in the same cluster.

The hydrothermal systems in Figure 9.7(a) are active today, yet, with two exceptions, their distribution seems generally to mimic that of extensional faults of the last 15 m.y. [Figure 9.3(b)]. Areas of prominent exception are (1) a broad corridor along the San Andreas Fault, in coastal California [see Lachenbruch and Sass (1973) for an explanation of the high-heat flow there]; and (2) a corridor along the Cascade Range of Washington, Oregon, and northern California, a belt of active, subduction-related andesitic volcanoes. These hot springs are thus associated with regions of (1) crustal extension and basaltic volcanism, (2) active calc-alkaline volcanism, and (3) transform plate motion. Hydrothermal systems in the extensional regime doubtless use the high regional permeability provided by faults and related fractures.

Fossil equivalents of these hydrothermal systems are shown in Figure 9.7(b). They are sites of Tertiary and Mesozoic (and some still older) epigenetic ore deposits, many of hydrothermal origin. To a limited extent their distribution may reflect accidents of erosion. In some regions, however, such as the Colorado Plateaus, subsurface data are abundant enough to indicate a general scarcity, if not absence, of such deposits on a regional scale, both now and in the past. The boundaries drawn in Figure 9.7(a) are repeated in Figure 9.7(b) to illustrate the fact that although some of the older hydrothermal systems are in locations quite different from those active today. a great many are in the same areas. The Great Basin hot springs, those north of the Snake River Plain (in central Idaho and southwestern Montana), and those in and near the Rio Grande rift (in Colorado and New Mexico) have ancestors in very much the same places, dating back tens of millions of years. Clearly, hydrothermal convection has been a feature of the shallow crust in these areas during times of both crustal compression and extension, whether of subduction-related or postsubduction origin.

What can be said of the exceptions, those clusters of springs or areas of spring-absence identified in Figure 9.7(b) by letters? Significantly, the most numerous exceptions are older deposits and hydrothermal systems that have not persisted to the present, rather than younger ones that sprang up anew, as they have in virgin areas B, G, and K. Deposits in areas C and D occur largely in oceanic crust accreted to the continent.

Areas E and F are of special interest because they tell us something about significant contrasts within the Basin and Range province, such as the strong difference between the Great Basin and Sonoran Desert sections. Hydrothermal convection is dying out in the Sonoran Desert region, just as crustal extension has.

Figures 9.5-9.7 show that the Great Basin and Rio Grande rift have been perturbed thermally for a very long time and that crustal temperatures were high in these areas long before extension began. High temperature probably influenced extensional deformation from its outset; at the very least it thermally weakened the crust and thereby helped to determine where extension and thinning would ultimately take place.

THE DISTRIBUTION OF EARTHQUAKES

The spatial distribution of earthquakes in the western United States is shown in Figure 9.8. Many are concentrated in a broad corridor parallel to the San Andreas Fault

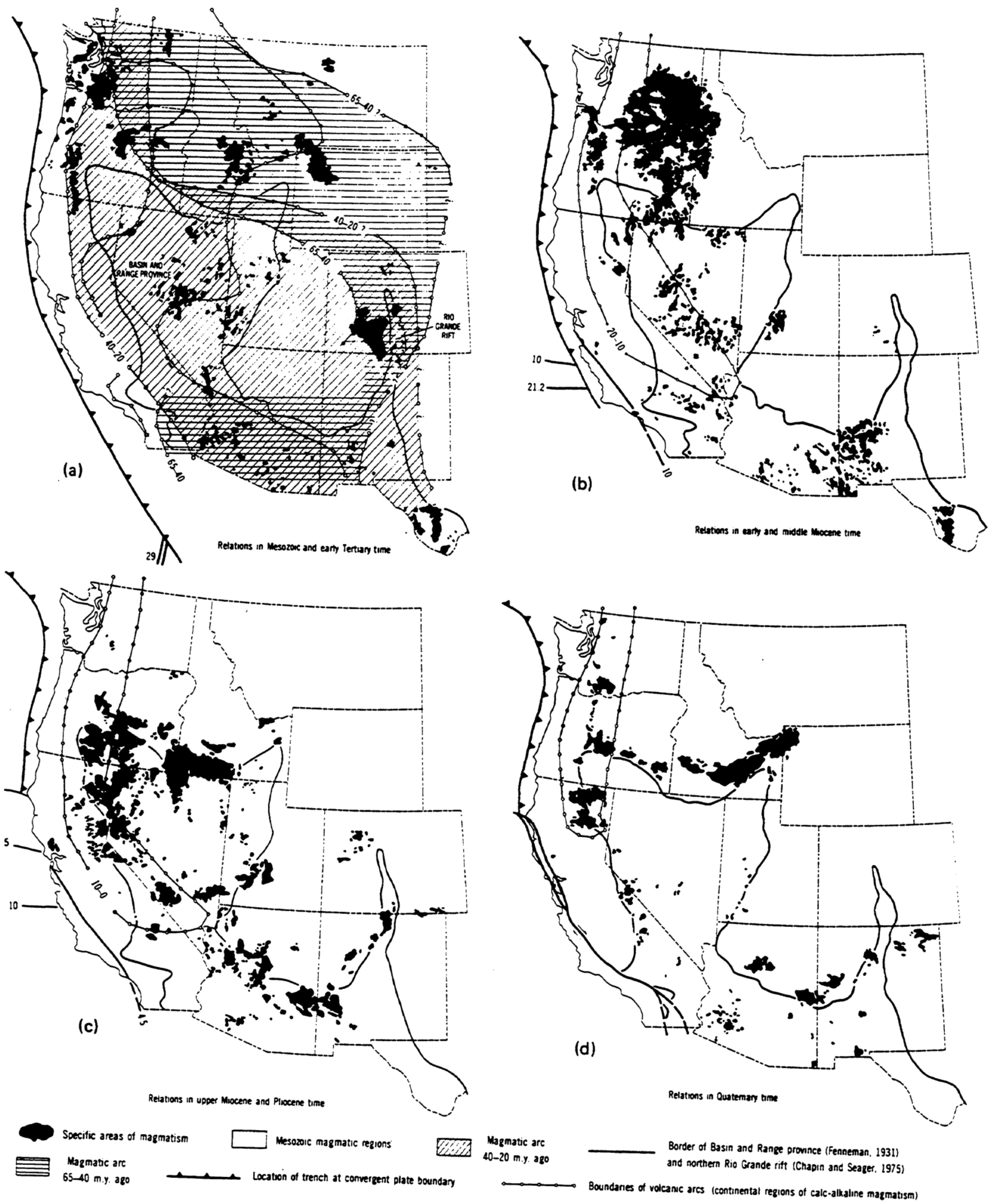


Figure 9.5
Post-Paleozoic magmatism in the western United States (data from King and Beikman, 1974; Snyder et al., 1976). The heavy line is the present-day border of the Basin and Range province, after Fenneman (1931), with addition of the northern Rio Grande rift from Chapin and Seager (1975).

system in coastal California, but many are well inland, some as far as 1700 km from the coast. Atwater (1970) considered the San Andreas to be the key element in the transform boundary between the Pacific and North American plates. She viewed tectonic activity inland, at least to the eastern edge of the Great Basin, as reflecting deformation in a broad, soft zone functioning as part of that boundary.

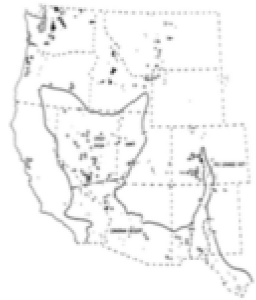


Figure 9.6
Shallow Cenozoic intrusive masses and ash-flow related calderas in the western United States (data from King and Beikman, 1974; E. H. McKee, USGS, unpublished data; T. A. Steven, USGS, unpublished data; and Eaton et al., 1978). Small irregular black areas and tiny dots are exposed intrusive rocks; open circles are sites of ash-flow related calderas. The calderas represent large bodies of magma at crustal levels shallow enough for their roofs to founder. They mark sites where shallow local crustal temperatures were, or still are, anomalously very high. The heavy line is Fenneman's (1931) boundary of the Basin and Range province, plus Chapin and Seager's (1975) boundary for the Rio Grande rift, the southern part of which is dashed. The dash-dot-dot line marks the midpoint of a steep, regional gravity gradient between the Great Basin and Sonoran Desert sections of the Basin and Range province (after Eaton et al., 1978).

The smoothly curving line in Figure 9.8 marks the inland edge of the most abundant earthquakes. Although Smith (1978) used a lower threshold magnitude for plotting California earthquakes, the line I have drawn is, for the most part, within California, hence represents a real boundary. This line approximates the *present* inboard limit of pronounced right-lateral shearing motion between the Pacific and North American plates. East of this line, strike-slip first motions in the southern Great Basin are left-lateral in displacement sense, with the active nodal plane striking west to southwest (see the compilation of Smith and Lindh, 1978).

Strike-slip motion has not been recorded or observed in the eastern Great Basin, but in west-central Nevada, in a zone 100 to 150 km wide east of the boundary shown in Figure 9.8, oblique-slip faults with dextral components of shear are observed. The nature of displacements has been confirmed by fault-plane solution, by geodetic measurement, and by direct observation in the field (Thompson and Burke, 1974). Slemmons (1967) observed strike-slip faults across the entire Nevada section of the Great Basin, but those east of the zone in question are of much smaller displacement, suggesting that the direct effects of dextral shear related to plate interaction die out inland in the western Great Basin. The total seismic-strain energy released in the eastern two thirds of the Great Basin is less, in the aggregate, than that related to strike-slip faulting in the curve-bounded region to the west (Ryall *et al*., 1966; Crampin *et al*., 1976).

Figure 9.9 shows statistical and geographical variations in the focal depths of earthquakes. Hypocenters are distributed from the near surface to a depth of 20 km but are seldom found at depths greater than 15 km. In the eastern part of the Great Basin (histograms A, D, G, and H) earthquakes are limited to the upper 15 km of the crust and concentrated in the upper 10 km, with some of the modal values (e.g., sites D and G) in the upper 5 km. Sites D and G also show a systematic decrease in the numbers of earthquakes downward. Western Great Basin earthquakes (histograms I through N) show a tendency toward somewhat greater depth, and at sites K and L there is a systematic *increase* downward, but, as just noted, this is an area of profound strike-slip faulting, hence the deformational regime is different.

These depth characteristics provide clues to the nature of crustal extension. Because sudden instabilities that create earthquakes in the shallow crust are probably related to stick-slip displacements on pre-existing faults (Brace, 1972a), one may interpret their depths as depths of instantaneous faulting. Extensional faulting accompanied by abrupt stress drop appears to continue in places scarcely deeper than 10 km. If the faults themselves continue to deeper levels their displacements are characterized by stable sliding or fault creep rather than stick-slip (Byerlee, 1968; Brace and Byerlee, 1968; Brace, 1972b).

Almost certainly the extensional faults do *not* continue to deeper levels. Thompson (1966) and Hamilton and Myers (1966) pointed out that if the normal faults in the Great Basin continued as planes having dips like those observed at the surface to depths where facing pairs intersected, the fault blocks could be no thicker than 10-15 km. Many students of the region, beginning with Longwell (1933; 1945), have described features of these faults that suggest they flatten with depth (see Moore, 1960; Mackin, 1960; Hamblin, 1965; Anderson, 1971; Wright and Troxel, 1973; and Proffett, 1977), .leading to fault blocks even thinner than 10-15 km. In addition, regional-gravity data (Eaton *et al*., 1978) indicate that the local mountain ranges generally are not compensated isostatically, suggesting, in turn, that faults serving as surfaces of adjustment do not pass through the lithosphere. All of these data tend to suggest that the depth limit of earthquakes may be the depth limit of faulting. Stewart (1978) illustrated alternative interpretations of how such faults may terminate downward.

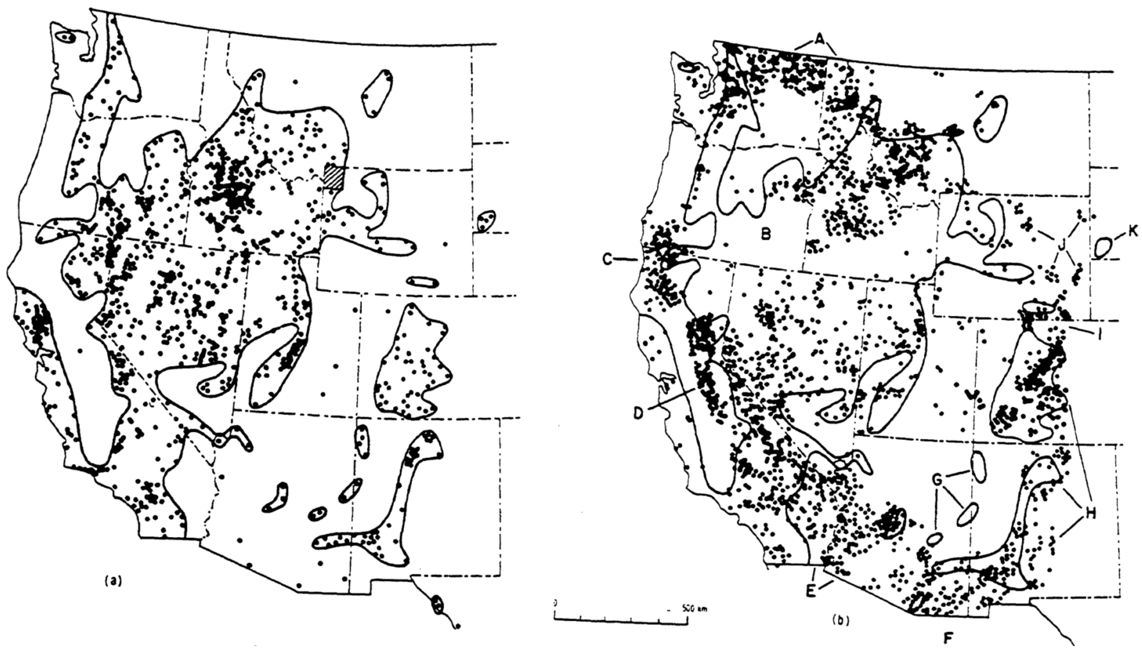


Figure 9.7
Active and fossil hydrothermal systems in the western United States. Data from Waring (1965) and Jerome and Cook (1967). (a) Active warm springs and hot springs the surface temperatures of which exceed the local mean annual air temperature by more than 8ºC. Yellowstone National Park, the site of numerous springs, is cross hatched. Boundaries enclose local and regional clusters of these springs. (b) Sites of ore deposits, largely of epigenetic origin, many of which mark the location of older hydrothermal circulation systems. Boundaries from diagram on left are repeated on right to illustrate similarities and differences in areas of occurrence. Letters identify areas of principal difference between the two maps, some of which are discussed in text.

Figure 9.10 compares the aggregate earthquake-depth distribution for the entire region with Basin and Range widths. The two histograms are generally similar in form, both having intermediate values between 0 and 5 km, peaking between 5 and 10 km, and having relative low levels of occurrence for values greater than 20 km. Of the earthquakes, 97 percent occur in the upper 15 km of the crust, which in the Great Basin is the upper *half* of the crust. Of Basin and Range block widths, 88 percent are no wider than 20 km. Theoretical and experimental analyses of the depth and spacing of fractures formed in extension (both tensile fractures and extensional *shear* fractures) suggest that the two values should be similar, at least within an order of magnitude (Lachenbruch, 1961; Sowers, 1972).

In a mechanical analysis of block gliding in which horsts and grabens formed, Voight (1973) derived an approximate equation for the width of a block formed by extensional faulting of an initially continuous slab. It is $W = 2T$ (45º-Φ/2), where W is the width of the block; T, the thickness of the faulted slab; and Φ, the coefficient of internal friction for effective stresses. Application of this equation to the Basin and Range province requires the assumption of a surface or zone of translatory sliding at the base of the fragmenting upper crust. It will be shown below that the Great Basin crust may have such a zone.

In order to solve Voight's equation we must have a value for Φ. Byerlee's (1968) experimental study of the brittle-ductile transition in rocks indicated that friction is independent of composition. Rocks under confining pressures of from 0 to 5.2 kilobars (the depth equivalent of 0 to 16.5 km) revealed a remarkably systematic variation between increasing normal stress (σ) and shear stress (τ) for friction. The limiting slopes, $\tau/\sigma = \tan \Phi$, of Byerlee's friction data curve are 36º and 46º. Substitution of these values in Voight's equation yields the following widths for fault blocks formed in this manner: for shallow crustal slabs initially 10 km thick, widths of 8.1-10.2 km; for slabs 15 km thick, 12.1-15.3 km; and for slabs 20 km thick, 16.2-20.4 km. These results are in good agreement with the observed Basin and Range block width-earthquake depth relation (Figure 9.10), hence Voight's model of extensional sliding may be judged to have potential relevance to an understanding of the mechanics of Basin-Range faulting.

Thompson (1959; 1966) was the first to suggest that extension in the deeper basin-range crust takes place via plastic stretching or injection of dikes. Hamilton and Myers (1966), Stewart (1971), and Proffett (1977) all accepted the first of these concepts, viewing Basin and Range structure as fragmentation of a shallow crustal slab riding on a plastically extending substratum. Lateral dilation of the lithosphere by magma from below is implicit in the thermomechanical model of Lachenbruch and Sass (1978). Wright and Troxel (1973) called upon both mechanisms (plastic stretching *and* intrusion) to extend the deeper crust beneath the fault-fragmenting surface slab of the western Great Basin.

NORMAL FAULTING AND BASAL SLIDING

Laboratory-scale models of extensional faulting that use sand or dry mortar as the deforming media (Hubbert, 1951; Stewart, 1971) yield structures similar to single, simple grabens and distributed arrays of alternating grabens and horsts. Stewart's model, which was designed specifically to resemble Great Basin structure, had a significant feature—a constructed surface of translatory sliding at its base. In order to predestine the spacing and plan of individual horsts and grabens, Stewart placed segmented sheets of paper beneath dry mortar. These sheets constituted a basal dislocation between the locally fragmenting mortar above and a sheet of uniformly extending rubber (the model's analog of a plastic substrate) beneath. Translatory sliding took place between the paper and the rubber sheet. In Hubbert's (1951) model, horizontal sliding took place at the base of the sand section, and it is not difficult to imagine striations developing on the floor of the deformation box parallel to the direction of extension after repeated runs.

In a real earth, such a dislocation could take one of two forms: (1) a simple surface of sliding or (2) a thin zone of

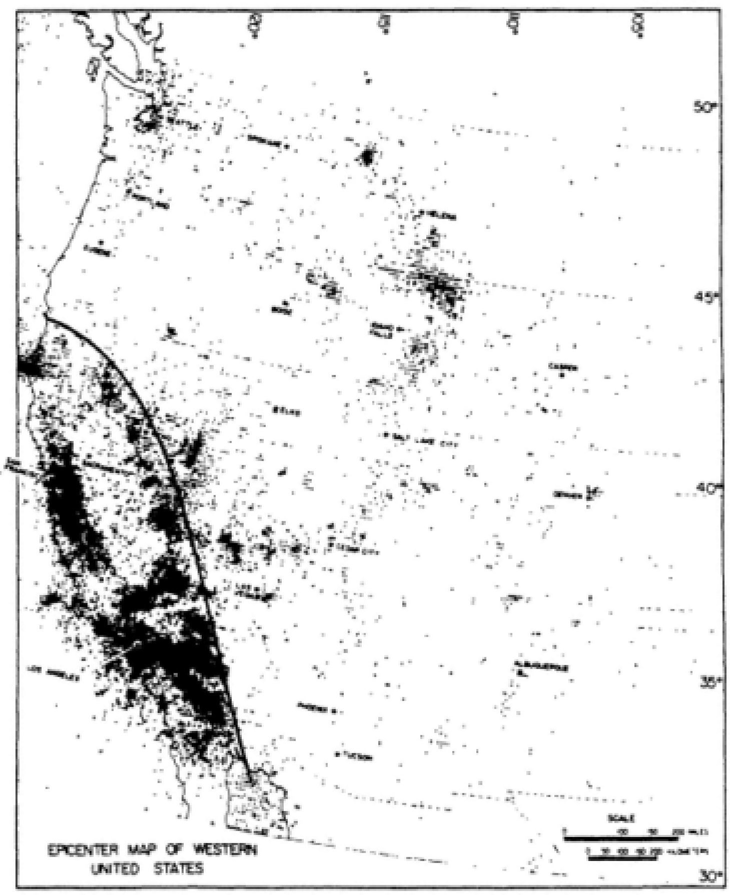


Figure 9.8
Seismicity of the western United States (from Smith, 1978). Lower threshold magnitudes were used in plotting California data. Heavy line, based on seismic data, marks inboard limit of highest earthquake event frequency and areal density in California, high cumulative seismic-strain energy release, and major strike-slip faulting related to dextral shear of the western plate boundary in Holocene time. Major faults within 150 km east of this line show oblique slip, with active strike-slip components, but farther east, the dominant mechanism is simple extension. (Reprinted from Geological Society of America Memoir 152, with permission.)

distributed shear or décollement. A subhorizontal striated surface, a thin zone of mylonite, or a thick layer of dynamothermally metamorphosed rock might thus be anticipated, depending on depth, effective pressure, temperature, composition, and shear stress.

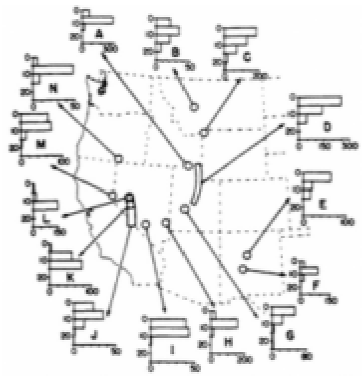


Figure 9.9
Histograms of earthquake focal depths (in kilometers) for 14 areas within the region of active extension in the western United States. Data from Gumper and Scholz (1971); Jaksha et al. (1977); Ryall and Savage (1969); Shuleski et al. (1977); and Smith (1978).

The Turnagain Heights translatory slide in Alaska may be cited as an example of such deformation. It was described and illustrated by Hansen (1965) and mechanically analyzed by Voight (1973). It mimics the Basin and Range province in structure. The original thickness of the fragmenting slab was only 20 m, hence there was little tendency for normal faults to flatten appreciably with depth. Extension and sliding of the originally intact surface slab was possible because of an absence of lateral support on one side and a perceptible (2.2º) slope of the basal surface. Genetically, the structure was a gravity slide. It resulted from a loss in strength of material in the vicinity of the sliding surface as a result of excitation by a major earthquake. Owing to the backward retreat of its headwall, as more and more of the fragmenting surface slab slid away laterally, Voight termed the feature a "retrogressive block-glide." I do not suggest that the driving force of Basin-Range faulting is the same, only that kinematics and resultant structures are grossly similar and, for this reason, instructive.

Evidence suggests that the Great Basin may be growing laterally, i.e., that it may be consuming neighboring regions on both west and east (Smith *et al*., 1976; Eaton *et al*., 1978). Both margins show transitional regions that have geophysical anomalies characteristic of the Great Basin extending tens of kilometers into the neighboring provinces (see Ryall and Stuart, 1963; Shuey *et al*., 1973; Smith and Sbar, 1974; Keller *et al*., 1975; and Eaton *et al*., 1978). The eastern margin of the Basin and Range province also has geological characteristics suggestive of transition (Best and Hamblin, 1978; Howard *et al*., 1978; Luedke and Smith, 1978). This state may relate, at least superficially, to the retrogressive aspect of the Turnagain Heights slide. If so, the Sierran and Wasatch fronts (opposed headwalls) may be retreating from each other as a result of plastic stretching at depth, thus effecting a growth of the province at the expense of adjoining regions. This could account for the observed outward restriction of magmatism with time. Uplift in the regions of these headwalls is probably a thermal phenomenon that is essentially contemporaneous with the extensional faulting itself, just as it is in the oceans. Although heat flow in the Sierra Nevada is anomalously low, it must, in part, reflect the appreciable thermal time constant of the crust, for the mass deficiency characteristic of the Great Basin continues beneath the Sierra Nevada (Eaton *et al*., 1978).

The Great Basin has a well-developed bilateral symmetry in certain aspects of its geology, but far more obviously in its geophysical fields (Proffett, 1977; Eaton *et al*., 1978). In Proffett's model the western half of the shallow, fragmenting, crustal slab translates eastward relative to the extending substrate beneath (the middle and lower crust), and that of the eastern half, westward.

According to Voight (1973) retrogression cannot take place if the glide blocks are rigid (as opposed to internally deformable) unless fluid pressures within fractures are sufficiently high to perform the function of plastic wedges. Basin and Range blocks are sufficiently fractured at the surface to suggest internal deformation. The significant deformational model thus appears to be lateral spreading with deformable block gliding.

GEOPHYSICALLY ANOMALOUS LAYERS IN THE SHALLOW CRUST

The possibility of a surface of sliding or a zone of ductile flow (mylonite or other metamorphic rocks) beneath the fault-fragmented surface slab of the Basin and Range province raises the question of their detectability by geophysical means. We examine this issue only briefly, but in the last section of the chapter offer a tentative crustal model, based primarily on geology, heat flow, and earthquake data, to which the other geophysical observations are fit by hypothesis. The hypothesis needs specific testing.

In the past decade and a half an increasing number of reports of a seismic low-velocity layer in the shallow crust have been published. One example is in the eastern Great Basin, near its boundary with the Colorado Plateaus (Mueller and Landisman, 1971; Landisman *et al*., 1971; Braile *et al*., 1974; Keller *et al*., 1975; Smith *et al*., 1975;

and Braile, 1977). Because such a feature is also associated with the Rhine graben (Landisman *et al*., 1971) it is tempting to conclude that it is a feature characteristic of extensional regimes.

Shurbet and Cebull (1971) suggested that the crustal low-velocity layer in the Great Basin is a zone of decreased rigidity that provides a means of absorbing the displacements of Basin-Range faults. According to them, the top of this zone is at levels of 5 km or so, and the base, at 8-9 km. Braile *et al*. (1974) suggested that surface extension by normal faulting at the surface is absorbed in a soft, plastically extending region of lowered seismic velocity.

Braile (1977) later published the following additional information: (1) the layer has a compressional wave velocity perhaps as low as 5.5 km sec^{-1} (compared with velocities above and below of 6.0 and 6.5 km sec^{-1}, respectively); and (2) it has a heightened Poisson's ratio; (3) an anomalously low *Q* (quality factor) for the transmission of compressional seismic waves; and (4) a top at 9.5 km and base at 15 km. Although the depth and thickness of this layer are different in the Shurbet and Cebull model, the values were derived in very different ways. The figures of Braile (1977) are preferred. Both sets of authors agree on the existence and gross mechanical properties of the layer, and on its role in accommodating extension at the surface. Neat as this picture is, it is marred by the fact that crustal low-velocity layers are not peculiar to extensional regimes.

Two collections of papers on the physical properties and conditions of the continental crust (Heacock, 1971, 1977) reveal that the phenomenon is widespread, found in areas of young extension as well as in stable Precambrian shield areas (Berry and Mair, 1977). Such features have been observed in the crust of all the continents except Antarctica. Mueller (1977) has incorporated it as a key element in a generalized model of the continental crust. According to him, it is found fairly consistently at depths of 5-15 km. Some investigators, however, place it as deep as 20 kin, e.g., Landisman and Chaipayungpun (1977). Its origin has been ascribed to high temperatures (Smith *et al*., 1975), to the presence of a zone of granitic intrusions (Mueller, 1977), to high pore-fluid pressures (Berry and Mair, 1977), or to some combination of any or all of these factors (Mueller, 1977).

Laboratory experiments (Nur and Simmons, 1969; Todd and Simmons, 1972; and Brace, 1972b) demonstrate that as pore pressures rise toward lithostatic values (thereby reducing effective pressure) seismic velocities fall toward those observed at exceedingly shallow levels in the crust. If a consensus as to the origin of the crustal seismic low-velocity layer is emerging, it is high pore pressure. In the Great Basin, high regional heat flow, which implies high crustal temperatures, probably plays an important supportive role.

Pore water at high pressures in a closed system is capable of lowering seismic velocity, *Q* values, and rock strength, conditions that appear to occur in the Great Basin crust. As Berry and Mair (1977) point out, however,

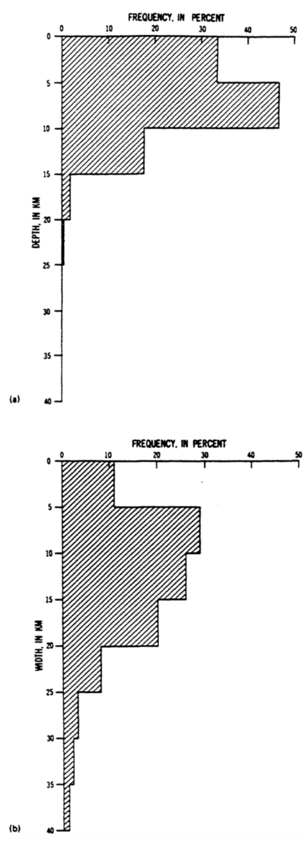


Figure 9.10
Earthquake focal depths and widths of Basin and Range blocks. (a) Histogram of focal depths of 2,475 earthquakes in the region of extension; (b) widths of individual basins and ranges in the Great Basin scaled from the map of King and Beikman (1974). Both characteristic dimensions (depth and width) show modal values in the range 5-10 kin, with most values (>85 percent) in the range 0-20 km.

this explanation requires that rocks in the layer in question have finite porosities at depths of 5-15 km, while those in the zone immediately above it must be free of permeability because the hydraulically pressured layer must have some sort of impermeable cap. This cap in the Great Basin may be a layer of rock extended by ductile flow, the one whose upper surface may mark the base of the region of brittle faulting. If so, the seismic low-velocity zone may not coincide with this layer but is perhaps *beneath* it. Depths to surfaces or zones of Tertiary sliding in the Great Basin have been estimated by Armstrong (1972) to be at least 8 km on geological grounds, but the thick, ductile zones must be deeper still, because in many places the glide faults do not rest directly on ductile rocks. Because depths of somewhat more than 8 km are in reasonable agreement with Braile's (1977) seismic estimate of 9.5 km to the *top* of the seismic low-velocity layer, I tentatively regard the ductile layer (if it truly exists) as a possible cap. Pore fluids could be trapped at high pressure in fractures in rock immediately beneath such a ductile layer. Internal displacements or structural adjustments within rocks at this level would take place by stable sliding (Brace, 1972a; 1972b), and earthquakes would not be common at such depths. As we have already seen, they are not.

Pore water in closed-rock systems will also lower electrical resistivity, as will increasing temperature, which elevates the ionic mobility of such fluids. At very high temperatures the onset of partial melting could do much the same thing; the melting temperature of the rocks is lowered by. the presence of water. For these reasons one might anticipate the presence of electrical conductors in the Great Basin crust, and, in fact, they are observed.

Some investigators (e.g., Landisman and Chaipayungpun, 1977; Lienert and Bennett, 1977) have equated the crustal low-velocity layer with a low resistivity layer in tectonically active or high heat-flow areas. Much remains to be done to substantiate this equivalence, and also to establish equivalence between a subhorizontal low-velocity layer in the crust and a porous zone capped by a stratum of ductile impermeable rock. The data in hand are permissive, but thus far hardly conclusive. The electrical data are reviewed briefly to provide an idea of what is currently known. I am indebted to my friend and colleague, J. N. Towle, for providing the information summary that follows.

Schmucker's (1970) geomagnetic variometer investigations in California indicated the presence of an electrical conductor in the western Cordillera at the eastern base of the Sierra Nevada, at and near its boundary with the Great Basin. Stanley *et al*. (1976b) identified a shallow (2-7 km), highly conductive layer in the crust beneath the Carson sink in western Nevada by means of magnetotelluric soundings. Stanley *et al*. (1976a) also studied the electrical structure of the Long Valley geothermal system in the western Great Basin by means of direct current and electromagnetic techniques, concluding that hydrothermal activity is reflected in discrete conductive zones in the crust, which are controlled, in turn, by regional faulting. Lienert and Bennett (1977) have identified a crustal conductor in the western Great Basin at a depth of 20 km using controlled-source geomagnetic variometry.

Reitzel *et al*. (1970) and Porath and Gough (1971) observed generally reduced vertical geomagnetic field variations in the eastern Great Basin that they interpreted as reflecting a shoaling of the mantle. Ambiguities in their interpretation of crustal thickness will doubtless be resolved as the evidence mounts both for a shallow, strongly conducting, crustal layer in the Great Basin as a whole and for local shoaling of the asthenosphere.

Studies by W. D. Stanley and colleagues at the U.S. Geological Survey (personal communication, 1978) have revealed the presence of a conductive crustal layer near the boundary between the Great Basin and Snake giver Plain on the north. Depths to its top range from 2 to 10 kin; its thickness may be as great as 10 km. Stanley *et al*. (1977) have also identified a conductor beneath the Snake giver Plain region at depths of only 5 km in the Yellowstone caldera, but deepening to 20 km on the southwest. In the vicinity of the Raft River geothermal area, in the northeastern Great Basin, it is 7 km deep. This conductor may be related directly to the presence of magma, at least in the Yellowstone area, and, therefore, may or may not be directly related to the crustal low-velocity layer under discussion.

On the basis of these limited data it appears that an electrically conductive layer is a common feature of the shallow Great Basin crust. Depth estimates place its top between 2 and 20 km, and in several areas, at less than 10 km. Possibly this conductor coincides with the crustal low-velocity layer, but too little is known about it to be certain. Coincidence might be anticipated simply because some of the factors that lower seismic velocity also raise electrical conductivity (high temperature, high porosity, the presence of a pore fluid, or the presence of a silicate melt). An electrically conductive layer by itself does not require abnormally high pore pressures and, hence, does not require the presence of an impermeable cap to keep the system closed. The presence of conductive minerals such as metallic sulfides or those having high ion-exchange capacity, like clays or zeolites, can also lower rock resistivities without affecting seismic velocity. The low-resistivity layer in the crust could just as well be *above* the low-velocity layer, reflecting some combination of high porosity, temperature, pore-fluid salinity, or hydrothermal alteration in the lower part of the shallow crust.

CRUSTAL MODEL FOR THE BASIN AND RANGE PROVINCE: A SUMMARY AND INTERPRETATION

The crust of the Great Basin section of the Basin and Range province (and its immediate environs) is higher in elevation, thinner, warmer, more highly fractured, and more well endowed with hot springs than that of surrounding regions, excluding the area immediately north of the Snake River Plain. The fractures (mostly faults) extend a third of the way to halfway through the crust. They are loci of abundant shallow earthquakes and vigorous hydrothermal circulation. Crustal extension takes place by faulting near the surface, but probably takes place by other modes at depth, most likely by dike intrusion and stretching of the lower crust and lithospheric mantle, and by ductile shear flow (distributed décollement) in a relatively thin layer at some intermediate level.

Repeated magmatic invasions of the crust have taken place during the past 100 million years. Some of these magmas broke the surface, but some have come to rest within the crust, giving up their heat there. Shallow magmatic systems serve to drive hydrothermal convection in the shallow crust, as does high regional heat flow from the deeper crust. The phenomena have been long lived.

At present, extension is taking place in an east-west or west-northwest direction; earlier, it was directed southwest or west-southwest (Eaton *et al*., 1978; Zoback and Thompson, 1978). The Sonoran Desert was included in the initial episode of extension but is not included in the present one. Because the Sonoran Desert is deeply eroded, it exposes the effects of crustal extension at deeper levels. A large part of the Sonoran Desert in southwestern Arizona reveals evidence of subhorizontal, unidirectional plastic strain of middle Tertiary age (Davis, 1977; Davis *et al*., 1977; Davis, see Chapter 8; Rehrig and Reynolds, 1977) that initially developed before block faulting began but that may have served as the base of the faulted, shallow slab. The mylonitization and metamorphism probably took place at lithostatic pressures of several kilobars.

Armstrong (1972) reviewed evidence of translatory displacements in the Basin and Range province and argued that some of the subhorizontal surfaces of sliding in the Great Basin are certainly Tertiary in age, that many of them *may* be Tertiary, and that they are more likely related to Basin and Range faulting than to the Sevier orogenic event, the youngest episode of pre-extension thrusting. Most of these dislocations place younger strata over older. He noted that some of the structures are of relatively deep-seated origin (at least 8 km). It is possible that these subhorizontal zones of sliding first developed as thrust soles during crustal compression, later to evolve into extensional décollements. These observations and speculations lend themselves to the interpretation that prolonged thermal conditioning of the crust plus horizontal shearing simply may have continued earlier initiated dynamothermal metamorphism of rocks that now serve as a boundary layer between parts of the lithosphere extending by fundamentally different mechanisms. As young normal faults developed near the surface in the regime of crustal extension, they became listric to (they came to sole on) older thrust zones.

The mechanical model of Kehle (1970), in which a décollement is distributed through the middle (relatively more ductile) layer of a crustal or lithospheric triad. may be applicable. Shearing in such a layer would lead to the mechanical generation of heat. Its magnitude would be controlled by the rate of shearing, which, judging from rates of extension measured at the surface, should be lower than that generated along the San Andreas Fault (see Lachenbruch and Sass, 1973). Some part of the high heat loss in the Great Basin could be due to shearing, however. The greater part has been ascribed to penetrative convection of the lithosphere by basaltic magma (Lachenbruch and Sass, 1978). Part is ascribable to convective groundwater circulation in the shallow crust and locally, to young, hot, volcanic systems residing in the upper crust. If this model, based largely on geological, heat flow, and earthquake data, is generally correct, it has implications for the surface patterns of deformation, the regional distribution of geological resources, and some of the effects of earthquakes. The model is shown in schematic form in Figure 9.11. To summarize its implications:

1. The location and extent of the Basin and Range province may have been largely predetermined by the location and extent of early Tertiary and Mesozoic magmatism that preheated (thermally weakened) the crust and augmented a regime of compressional thrusting in which subhorizontal dislocation surfaces or ductile zones (distributed décollements) first developed at middle to shallow crustal levels. Such zones could be used later as basal dislocations for normal faulting and might also serve (at depth) as crustal membranes impermeable to the deeper circulation of groundwaters but allowing the upward passage of magma by intrusion. Normal faults at the surface probably are listric to the deepest of these zones, inasmuch as the shallowest ones are exposed in the uplifted fault blocks themselves.
2. Zones of translatory, ductile shear would constitute near-horizontal surfaces of mechanical decoupling or inefficient coupling within the crust. As a result, deformations and kinematic motions in the lower crust would not always be clearly or faithfully reproduced at the surface.
3. The regional maintenance of long-continued high temperatures and high permeability assures the continuation of vigorous hydrothermal circulation and attendant epigenetic deposition of minerals through both compressional and extensional regimes. They may, on the other hand, be responsible for what appears to be a regional scarcity of oil and gas in the Basin and Range province. Where unfavorably situated, such fluids could be driven to the surface, where they could escape, except for local conditions of entrapment. For the same reason, consideration of Great Basin sites for the isolation and storage of radioactive wastes carries with it the requirement of a critical evaluation of the local hydrologic and seismotectonic regime.
4. Seismic energy traversing the crust of the extended region is probably absorbed to a somewhat greater degree than it is in the relatively less intensely fractured, shallow crust of the central and eastern United States, hence the

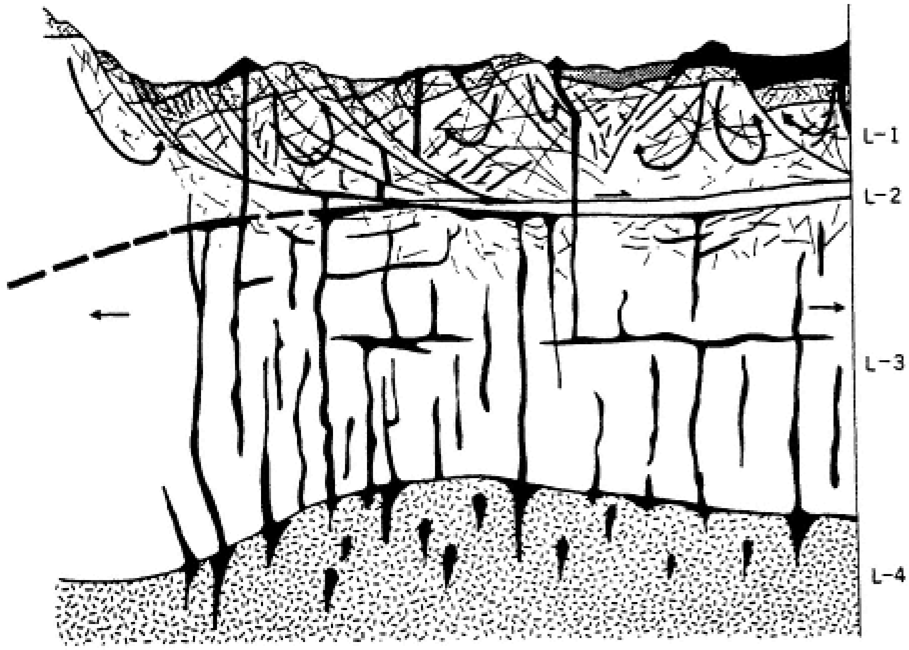


Figure 9.11
Interpretive model of possible crustal structure of the Great Basin (simplified, schematic, and not to scale); based on surface geology, heat flow, and earthquake distribution. (See Stewart, 1978, for alternative interpretations of the near-surface structure.) The crust is composed of three layers (L-l, L-2, L-3) having different lithologies and physical properties. Each fails or yields in extension by a different physical mode. Erosion in the Sonoran Desert region generally has cut down to the level of L-2. L-4 is lithospheric mantle. Characteristics of these layers are as follows: L-l, Fault-fragmented, surface layer, 8-15 km thick, composed of rocks of a great variety or origins, compositions, and ages, all exposed at the surface somewhere in the region; diagram shows Cenozoic continental sedimentary and volcanic rocks at the surface (patterns of dense stippling and solid black, respectively) overlying older sedimentary rocks (open stippling) and granitic and metamorphic basement rocks (plain white). Although the diagram does not show it, stratified rocks extend well into L-2 and probably into L-3 (as in Arizona). All of these rocks are highly fractured, as indicated by the plexus of fine, irregular lines. The layer fails in semibrittle fashion by normal faulting, fault-block rotation, pervasive fracturing, and slumping. The deformation creates high fracture deeper crust and by local, young intrusions (black, dike-like bodies). The upper part of the crustal low-resistivity layer may coincide with the lower part of this layer. The base of the layer generally marks the maximum depth of earthquakes. L-2 ductile intermediate layer, 0-3 km thick, composed of pervasively sheared, mylonitized, and/or dynamothermally metamorphosed Miocene and older rocks of a wide variety of original compositions (medium stippling). At one extreme, the layer is a vanishingly thin stratum of mylonite, 1-10 mm thick; at the other, a layer of granitic augen gneiss, schist, or amphibolite, 1-3 km thick. This layer is locally or regionally lineated and extends by laminar plastic flow. It is generally impermeable to groundwater circulation except where later uplifted and fractured in the brittle regime, but at depth it may be cut by dikes of Tertiary igneous rocks (solid block). It developed first as a regional thrust sole (heavy dashed line) during earlier crustal compression. L-3, lower crustal layer, 10-20 km thick, composed near its top of igneous and metamorphic basement rocks like those of layer L-1 but grading downward into increasing proportions of old granites, migmatites, gneisses, amphibolites, and felsic to mafic granulites, in approximately that order. This layer extends by a combination of diking (by basalts from the asthenosphere, solid black) and solid-state convection (stretching and underplating). It is rigid at relatively high and intermediate anomalously high heat flow observed in the province, The seismic low-velocity zone may coincide with the uppermost part of this layer mantle, 25-35 km thick, composed of ultramafic rock devoid of finely disseminated melt. This layer, like the lower crustal layer above it, extends by diking (perhaps via rising, bleb-like bodies) and solid-state convection. It is immediately underlain by asthenospheric mantle.

geographical extent of isoseismal boundaries for an earthquake of given magnitude is generally less in the West than it is in the Midwest and East.

ACKNOWLEDGMENTS

Preparation of this review was aided substantially by discussions with, and/or data contributions from, M. D. Crittenden, Jr., G. A. Davis, Mel Friedman, W. Hamilton, A. H. Lachenbruch, P. W. Lipman, E. H. McKee, S. S. Oriel, A. R. Sanford, R. B. Smith, T. A. Steven, and J. N. Towle. The manuscript benefited from constructive reviews by K. A. Howard and H. M. Iyer. None of the above should be regarded as subscribing to the interpretive crustal model presented here, however, nor to its stated implications. I am indebted to members of the technical staff of the Hawaiian Volcano Observatory, Ron Hanatani, Rick Hazlett, Jenny Nakata, and Maurice Sako, as well as summer student aide Kevin Cuff for graphical compilation of many of the regional geological and geophysical data presented here.

References

Anderson, R. E. (1971). Thin skin distension in Tertiary rocks of southeastern Nevada, *Geol. Soc. Am. Bull. 82*, 43-58.

Armstrong, R. L. (1972). Low-angle (denudation) faults, hinterland of the Sevier orogenic belt, eastern Nevada and western Utah, *Geol. Soc. Am. Bull. 83*, 1729-1754.

Atwater, T. (1970). Implications of plate tectonics for the Cenozic tectonic evolution of western North America, *Geol. Soc. Am. Bull. 81*, 3513-3536.

Barazangi, M., and J. Dorman (1969). World seismicity maps compiled from ESSA, Coast and Geodetic Survey, epicenter data, 1961-1967, *Bull. Seismol. Soc. Am. 59*, 369-380.

Berry, M. J., and J. A. Mair (1977). The nature of the earth's crust in Canada, in *The Earth's Crust*, J. G. Heacock, ed., Am. Get-phys. Union Geophys. Monogr. 20, pp. 319-348.

Best, M. G., and W. K. Hamblin (1978). Origin of the northern Basin and Range province: implications from the geology of its eastern boundary, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. R. Eaton, eds., *Geol. Soc. Am. Mere. 152*, 313-340.

Brace, W. F. (1972a). Laboratory studies of stick-slip and their application to earthquakes, *Tectonophysics 14*, 189-200.

Brace, W. F. (1972b). Pore pressure in geophysics, in *Flow and Fracture of Rocks*, H. C. Heard, I. Y. Borg, N. L. Carter, and C. B. Raleigh, eds., Am. Getphys. Union, pp. 265-273.

Brace, W. F., and J. D. Byerlee (1968). Stick slip, stable sliding, and earthquakes—effect and rock type, pressure, strain rate, and stiffness, *J. Geophys. Res. 73*, 6031-6037.

Braile, L. W. (1977). Interpretation of crustal velocity gradients and Q structure using amplitude-corrected seismic refraction profiles, in *The Earth's Crust*, J. G. Heacock, ed., Am. Get-phys. Union Geophys. Monogr. 20, pp. 427-439.

Braile, L. W. R. B. Smith, G. R. Keller, R. M. Welch, and R. P. Meyer (1974). Crustal structure across the Wasatch front from detailed seismic refraction studies, *J. Geophys. Res. 79*, 2669-2677.

Byerlee, J. D. (1968). Brittle-ductile transition in rocks, *J. Geo-phys. Res. 73*, 4741-4750.

Chapin, C. E., and W. R. Seager (1975). Evolution of the Rio Grande Rift in the Socorro and Las Cruces area, in *New Mexico Geological Society Guidebook, 26th Field Conference, Las Cruces County*, pp. 297-321.

Christiansen, R. L., and P. W. Lipman (1972). Cenozoic volcanism and plate-tectonic evolution of the western United States, II, Late Cenozoic, *Phil. Trans. Roy. Soc. London A271*, 249-284.

Christiansen, R. L., and E. H. McKee (1978). Late Cenozoic volcanic and tectonic evolution of the Great Basin and Columbia intermountain region, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. 152*, 283-312.

Cohee, G. V. (compiler) (1961). Tectonic map of the United States, exclusive of Alaska and Hawaii, U.S. Geol. Surv. and Am. Assoc. Petrol. Geol., scale, 1:2,500,000.

Couch, R. W., G. Thrasher, and K. Keeling (1976). The Deschutes Valley earthquake of April 12, 1976, *Ore Bin 38*, 151-161.

Crampin, S., C. J. Fyfe, D. P. Rickmore, and R. H. W. Linton (1976). Atlas of seismic activity 1909 to 1968, *Inst. Geol. Sci. Seismol. Bull. 5*, 1-29.

Davis, G. A., K. V. Evans, E. G. Frost, S. H. Lingrey, and T. J. Shackelford (1977). Enigmatic Miocene low angle faulting, southeastern California and west-central Arizona: suprastructural tectonics?, *Geol. Soc. Am. Abstr. Programs 9*, 943.

Davis, G. H. (1977). Characteristics of metamorphic core complexes, southern Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 944.

Eaton, G. P., R. L. Christiansen, H. M. Iyer, A. M. Pitt, D. R. Mabey, H. R. Blank, Jr., I. Zietz, and M. E. Gettings (1975). Magma beneath Yellowstone National Park, *Science 188*, 787-796.

Eaton, G. P., R. R. Wahl, H. J. Prostka, D. R. Mabey, and M. D. Kleinkopf (1978). Regional gravity. and tectonic patterns: their relation to late Cenozoic epeirogeny and lateral spreading in the western Cordillera, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mere. 152*, 51-92.

Eberly, L. D., and T. B. Stanley, Jr. (1978). Cenozoic stratigraphy and geologic history of southwestern Arizona, *Geol. Soc. Am. Bull. 89*, 921-940.

Fenneman, N. M. (1931). *Physiography of Western United States*, McGraw-Hill, New York, 534 pp.

Gumper, F. J., and C. Scholz (1971). Microseismicity and tectonics of the Nevada seismic zone, *Bull. Seismol. Soc. Am. 61*, 1413-1432.

Haimson, B. C. (1977). Crustal stress in the continental United States as derived from hydrofracturing tests, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 576-592.

Hamblin, W. K. (1965). Origin of "reverse drag" on the down-thrown side of normal faults, *Geol. Soc. Am. Bull. 76*, 1145-1164.

Hamilton, W, and W. B. Myers (1966). Cenozoic tectonics of the western United States, *Rev. Geophys. 4*, 509-549.

Hansen, W. R. (1965). Effects of the earthquake of March 27, 1964, at Anchorage, Alaska, in The Alaska Earthquake, March 27, 1964—Effects on Communities, *U.S. Geol. Surv. Prof. Paper 542-A*, 68 pp.

Heacock, J. G., ed. (1971). *The Structure and Physical Properties of the Earth's Crust*, Am. Geophys. Union Geophys. Monogr. 14, 348 pp.

Heacock, J. G., ed. (1977). *The Earth's Crust, Its Nature and Physical Properties*, Am. Geophys. Union Geophys. Monogr. 20, 754 pp.
Howard, K. A., J. M. Aaron, E. E. Brabb, M. R. Brock, H. D. Gower, S. J. Hunt, D. J. Milton, W. R. Muehlberger, J. K. Nakata, G. Plafker, D. C. Prowell, R. E. Wallace, and I. J. Witkind (1978). Preliminary map of young faults in the United States as a guide to possible fault activity, *U.S. Geol. Surv. Map MF-916*, 1:5,000,000.
Hubbert, M. K. (1951). Mechanical basis for certain familiar geologic structures, *Geol. Soc. Am. Bull. 62*, 355-372.
Hunt, C. B. (1967). *Physiography of the United States*, W. H. Freeman, San Francisco, 480 pp.
Jaksha, L. H., J. Locke, J. B. Thompson, and A. Garcia (1977). Albuquerque basin seismic network, *U.S. Geol. Surv. Open-File Rep. 77-865*.
Jerome, S. E., and D. R. Cook (1967). Relation of some metal mining districts in the western United States to regional tectonic environments and igneous activity, *Nevada Bur. Mines Bull. 69*, 1-35.
Karig, D. E. (1974). Evolution of are systems in the western Pacific, *Ann. Rev. Earth Planet. Sci. 2*, 51-75.
Kehle, R. O. (1970). Analysis of gravity sliding and orogenic translation, *Geol. Soc. Am. Bull. 81*, 1641-1663.
Keller, G. R., R. B. Smith, and L. W. Braile (1975). Crustal structure along the Great Basin-Colorado plateau transition from seismic refraction studies, *J. Geophys. Res. 80*, 1093-1098.
King, P. B., compiler (1969). Tectonic map of North America, *U.S. Geol. Surv. Map*, scale 1:5,000,000.
King, P. B., and H. M. Beikman (1974). Geologic map of the United States, *U.S. Geol. Surv. Map*, scale 1:2,500,000.
Lachenbruch, A. H. (1961). Depth and spacing of tension cracks, *J. Geophys. Res. 66*, 4273-4292.
Lachenbruch, A. H., and J. H. Sass (1973). Thenno-mechanical aspects of the San Andreas Fault systems, in *Proceedings of the Conference on the Tectonic Problems of the San Andreas Fault*, Stanford U. Press, pp. 192-205.
Lachenbruch, A. H., and J. H. Sass (1977). Heat flow in the United States and the thermal regime of the crust, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geo-phys. Monogr. 20, pp. 626-675.
Lachenbruch, A. H., and J. H. Sass (1978). Models of an extending lithosphere and heat flow in the Basin and Range province, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. 152*, 209-250.
Landisman, M., and W. Chaipayungpun (1977). First results from electrical and seismic studies of low resistivity, low velocity material beneath eastern Colorado, *Geophysics 42*, 804-810.
Landisman, M., S. Mueller, and B. J. Mitchell (1971). Review of evidencefor velocity inversions in the continental crust , in *The Structure and Physical Properties of the Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 14, pp. 11-34.
Lienert, B. R. and D. J. Bennett (1977). High electrical conductivities in the lower crust of the northwestern Basin and Range: an application of inverse theory to a controlled-source deep-magnetic-sounding experiment, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 531-552.
Lipman, P. W., H. J. Prostka, and R. L. Christiansen (1972). Cenozoic volcanism and plate-tectonic evolution of the western United States, I. Early arid middle Cenozoic, *Phil. Trans. Roy. Soc. London A271*, 217-248.
Lobeck, A. K. (1939). *Geomorphology, an Introduction to the Study of Landscapes*, McGraw-Hill, New York, 731 pp.
Longwell, C. R. (1933). Rotated faults in the Desert Range, southern Nevada (abstr.), *Geol. Soc. Am. Bull. 44*, 93.
Longwell, C. R. (1945). Long-angle normal faults in the Basin and Range province, *Eos Trans. Am. Geophys. Union 26*, 107-118.
Luedke, R. G., and R. L. Smith (1978). Map showing distribution, composition, and age of late Cenozoic volcanic centers in Arizona and New Mexico, *U.S. Geol. Surv. Map I-1091A*, scale 1:1,000,000.
Mackin, J. H. (1960). Eruptive tectonic hypothesis for origin of Basin-Range structure (abstr.), *Geol. Soc. Am. Bull. 71*, 1921.
Malone, S. D., G. H. Rothe, and S. W. Smith (1975). Details of microearthquake swarms in the Columbia Basin, Washington, *Bull. Seismol. Soc. Am. 65*, 855-864.
McKenzie, D. P. (1969). The relation between fault plane solutions for earthquakes and the directions of the principal stresses, *Bull. Seismol. Soc. Am. 59*, 591-601.
Moore, G. W. (1973). Westward tidal lag as the driving force of plate tectonics, *Geology 1*, 99-100.
Moore, J. G. (1960). Curvature of normal faults in the Basin and Range province of the western United States, *U.S. Geol. Surv. Prof. Paper 400-B*, B409-B411.
Mott, R. P., Jr. (1976). The relationship of microearthquake activity to structural geology for the region around Socorro, New Mexico, unpubl. M.S. thesis, New Mexico Institute of Mining and Technology, 64 pp.
Mueller, S. (1977). A new model of the continental crust, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geo-phys. Monogr. 20, pp. 289-317.
Mueller, S., and M. Landisman (1971). An example of the unified method of interpretation for crustal seismic data, *Geophys. J. Roy. Astron. Soc. 23*, 365-371.
Nut, A., and G. Simmons (1969). The effect of saturation on velocity in low porosity rocks, *Earth Planet. Sci. Lett. 7*, 183-193.
Porath, H., and D. I. Gough (1971). Mantle conductive structures in the western United States from magnetometer array studies, *Geophys. J. Roy. Astron. Soc. 22*, 261-275.
Prodehl, C. (1970). Seismic refraction study of crustal structure in the western United States, *Geol. Soc. Am. Bull. 81*, 2629-2646.
Proffett, J. M., Jr. (1977). Cenozoic geology of the Yerington district, Nevada, and implications for the nature and origin of Basin and Range faulting, *Geol. Soc. Am, Bull. 88*, 247-266.
Ranalli, G., and E. I. Tanczyk (1975). Meridional orientation of grabens and its bearing on geodynamics, *J. Geol. 83*, 526-531.
Rehrig, W. A., and S. J. Reynolds (1977). A northwest zone of metamorphic core complexes in Arizona, *Geol. Soc. Am. Abstr. Programs 9*, 1139.
Reiter, M., C. L. Edwards, H. Hartman, and C. Weidman (1975). Terrestrial heat flow along the Rio Grande rift, New Mexico and Southern Colorado, *Geol. Soc. Am. Bull.* 86, 811-818.
Reitzel, J. S., D. I. Gough, H. Porath, and C. W. Anderson Ill (1970). Geomagnetic deep sounding and upper mantle structure in the western United States, *Geophys. J. Roy. Astron. Soc. 19*, 213-235.
Rogers, G. C. (1977). Two recent Georgia Strait earthquakes (abstr.), *Geol. Assoc. Can. Ann. Mtg. Program Abstr. 2*, 45.
Ryall, A., and W. U. Savage (1969). A comparison of seismological effects for the Nevada underground test BOXCAR with natural earthquakes in the Nevada region, *J. Geophys. Res. 74*, 4281-4289.
Ryall, A., and D. J. Stuart (1963). Travel times and amplitudes from nuclear explosions, Nevada Test Site to Ordway, Colorado, *J. Geophys. Res. 68*, 5821-5835.

Ryall, A., D. B. Slemmons, and L. D. Gedney (1966). Seismicity, tectonism, and surface faulting in the western United States during historic time, *Bull. Seismol. Soc. Am. 56*, 1105-1135.
Sanford, A. R., and R. P. Mott, Jr., P. J. Shuleski, E. J. Rinehart, F. J. Caravella, R. M. Ward, and T. C. Wallace (1977). Geophysical evidence for a magma body in the crust in the vicinity of Socorro, New Mexico, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 385-403.
Sbar, M. L., and L. R. Sykes (1973). Contemporary compressive stress and seismicity. in eastern North America: an example of intraplate tectonics. *Geol. Soc. Am. Bull. 84*, 1861-1882.
Sbar, M. L., and L. R. Sykes (1977). Seismicity and lithospheric stress in New York and adjacent areas, *J. Geophys. Res. 82*, 5771-5786.
Schmucker, U. (1970). Anomalies of geomagnetic variations in the southwestern United States, *Bull. Scripps Inst. Oceanogr. 13*, 1-165.
Shuey, R. T., D. K. Schellinger, E. H. Johnson, anal L. B. Alley (1973). Aeromagnetics and the transition between the Colorado Plateau and Basin Range province, *Geology 1*, 107-110.
Shuleski, P. J., F. J. Caravella, E. J. Rinehart, A. R. Sanford, T. C. Wallace, and R. M. Ward (1977). Seismic studies of shallow magma bodies beneath the Rio Grande rift in the vicinity of Socorro, New Mexico, *Geosci. Dep. Open-File Rep. N. Mex. Inst. Min. Tech. 13*, 1-8.
Shurbet, D. H., and S. E. Cebull (1971). Crustal low-velocity layer and regional extension in the Basin and Range province, *Geol. Soc. Am. Bull. 82*, 3241-3244.
Slemmons, D. B. (1967). Pliocene and Quaternary crustal movements of the Basin and Range province, U.S.A., *J. Geosci., Osaka City U.*, 91-103.
Smith, R. B. (1978). Seismicity, crustal structure, and intraplate tectonics of the interior of the western Cordillera, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. 152*, 111-144.
Smith, R. B., and A. G. Lindh (1978). A compilation of fault plane solutions of the western United States, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. 152*, 107-110.
Smith, R. B., and M. L. Sbar (1974). Contemporary tectonics and seismicity of the western United States with emphasis on the Intermountain Seismic Belt, *Geol. Soc. Am. Bull. 85*, 1205-1218.
Smith, R. B., L. W. Braile, and G. R. Keller (1975). Upper crustal low-velocity layers: possible effects of high temperatures over a mantle upwarp at the Basin Range-Colorado Plateau transition, *Earth Planet. Sci. Lett. 28*, 197-204.
Smith, R. B., D. R. Mabey, and G. P. Eaton (1976). Regional geophysics and tectonics of the Intermountain West, *Geology 4*, 437-438.
Smith, R. L., and H. R. Shaw (1975). Igneous-related geothermal systems, in Assessment of Geothermal Resources of the United States, D. F. White and D. L. Williams, eds., *U.S. Geol. Surv. Circ. 726*, 58-83.
Snyder, W. S., W. R. Dickinson, and M. L. Silberman (1976). Tectonicimplications of space-time patterns of Cenozoic magmatism in the western United States , *Earth Planet. Sci. Lett. 32*, 91-106.
Sowers. G. M. (1972). Theory. of spacing of extension fractures, in *Geological Factors in Rapid Excavation*. H. Pincus. ed. Geol. Soc. Am., Eng. Geol. Case History No. 9, pp. 27-53.
Stanley, W. D., D. B. Jackson, and A. A. R. Zohdy (1976a). Deep electrical investigations in the Long Valley geothermal area, California, *J. Geophys. Res. 81*, 810-820.
Stanley. W. D., R. R. Wahl, and J. G. Rosenbaum (1976b). A magnetotelluric study of the Stillwater-Soda Lakes. Nevada geothermal area, *U.S. Geol. Surv. Open-File Rep 76-80*, 38 pp.
Stanley, W. D., J. E. Boehl. F. X. Bostick, and H. W. Smith (1977). Geothermal significance of magnetotelluric sounding in the eastern Snake River Plain-Yellowstone Region, *J. Geophys. Res. 82*, 2501-2514.
Stewart, J. H. (1971). Basin and Range structure: A system of horsts and grabens produced by deep-seated extension, *Geol. Soc. Am. Bull. 82*, 1019-1044.
Stewart, J. H. (1978). Basin-range structure in western North America: a review, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera. R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mem. I52*, 1-13.
Strange, W. E., and G. P. Woollard (1964). The use of geologic and geophysical parameters in the evaluation, interpolation, and prediction of gravity, Aeronautical Chart and Information Center, USAF, St. Louis, Missouri, Contact AF 23 (601)-3879 (Phase 1), 225 pp.
Thompson, G. A. (1959). Gravity measurements between Hazen and Austin, Nevada: a study of basin-range structure, *J. Geo-phys. Res. 64*, 217-229.
Thompson, G. A. (1966). The riff system of the western United States in The World Rift System, T. N. Irvine, ed., *Geol. Surv. Can. Paper 66-14*, 280-290.
Thompson, G. A., and D. B. Burke (1974). Regional geophysics of the Basin and Range province, *Ann. Rev. Earth Planet. Sci. 2*, 213-238.
Todd, T., and G. Simmons (1972). Effect of pore pressure on the velocity of compressional waves in low-porosity rocks, *J. Geo-phys. Res. 77,* 3731-3743.
Uyeda, S., and H. Kanamori (1979). Back-arc opening and the modeof subduction , *J. Geophys. Res. 84*, 1049-1061.
Voight, B. (1973). The mechanics of retrogressive block gliding, with emphasis on the evolution of the Turnagain Heights landslide, Anchorage, Alaska, in *Gravity and Tectonics*, K. A. De Jong and R. Scholten, eds., John Wiley, New York, pp. 97-121.
Wallace, R. E. (1977). Profiles and ages of young fault scarps, north-central Nevada, *Geol. Soc. Am. Bull. 88*, 1267-1281.
Waring, G. A. (1965). Thermal springs of the United States and other countries of the world—A summary, *U.S. Geol. Surv. Prof. Paper 492,* 1-383.
Warren, D. H., and J. H. Healy (1973). Structure of the crust in the conterminous United States, *Tectonophysics 20*, 203-213.
Wright, L. A., and B. W. Troxel (1973). Shallow-fault interpretation of Basin and Range structure, southwestern Great Basin, in *Gravity and Tectonics*, K. A. De Jong and R. Scholten, eds., John Wiley, New York, pp. 397-407.
Zoback, M. L., and G. A. Thompson (1978). Basin and Range rifting in northern Nevada: clues from a mid-Miocene rift and its subsequent offsets, *Geology 6*, 111-116.

IV

CHARACTERIZATION OF CONTINENTAL CRUST

10

Seismic Exploration of the Continental Basement: Trends for the 1980's

Jack E. Oliver
Cornell University

INTRODUCTION

In scientific research it is helpful to view one's activities occasionally from a fresh and different perspective. Imagine, for example, an astronaut-scientist visiting the earth from an advanced civilization on another planet and having the task of reporting the state of earthly science to his leaders. I think his report would be mixed in tone. He would report favorably about some of our efforts to investigate our surroundings. On one scale, for example, spacecraft leave the earth to explore the solar system and beyond in an effort that strains our technology. At another extreme of scale, huge sophisticated devices cause tiny subatomic particles to collide at high velocity in an effort to learn more and more about smaller and smaller entities. Once again our technology is strained. The astronaut would probably be favorably impressed by progress in certain methods for exploring the earth—sophisticated laboratory devices, the techniques of the petroleum industry, perhaps deep-sea drilling vessels. But I think he would be surprised and dismayed to find that a society of four billion people confined to the surface of the earth has been content to know so little about the rocks a few hundreds or thousands of meters below it, and from which it derives much of its livelihood —the rocks of the continental basement. In making this statement, I do not mean to be critical of those scientists who work very capably and professionally on this topic but rather of the magnitude of the total effort directed toward study of this part of the earth, an effort that seems too small in relation to the need for an inventory of its resources by an expanding and ravenous society. I hope new advances within the next decade will ease the mind of the astronaut on this point.

It is not that methods and tools for exploration of the continental basement are unavailable. We have them; many of them will be discussed in this paper. The seismological methods that I shall discuss are but a part of our overall capability. Others include further mapping, improved and extended field and laboratory studies of surface rocks, drilling for informational purposes, study of crustal xenoliths, and a variety of geophysical techniques. The problem is one of focusing scientific interest on the topic and of devoting an appropriate portion of our efforts to study of this region, for sound economic as well as scientific reasons.

Seisinology has been, of course, a major producer of information on the earth's interior—*the* major producer of certain kinds of information including structure and certain mechanical properties. Therefore, I wish to consider the potential of further seismic exploration of the continental basement. By continental basement I mean the entire continental crust below the sediments and the uppermost mantle.

From the broadest perspective and for several reasons, it is clear that the potential of the seismic method for exploration of the earth, and particularly the continental basement, is by no means exhausted. First, to do so would require sources and receivers scattered over and through the earth at the Nyquist spacing for the shortest waves that can be detected after propagation through the deep region of interest. We are certainly far from achieving such a level of observations at this time and from predicting what we would observe if we did. Second, we lag in analysis; we are not able to understand and to make useful conclusions from all the information we obtain now. Third, the current rate of discovery of new features continues to be high. From these three points, one can deduce confidently that a much improved understanding of the earth's interior remains to be revealed by seismology. The challenge of seismology is to approach this ideal in an optimum manner given economic and other practical constraints. In a sense, the various kinds of seismological studies represent various routes toward this goal. Let us consider some of them, more or less in order of decreasing wavelength.

SPECTRUM OF SEISMIC WAVELENGTHS

Since 1960, seismologists have been observing and studying waves of very long periods (up to almost an hour) and hence very long wavelengths that may be thought of as corresponding to *free oscillations of the earth*. Most studies of free oscillations treat the earth as almost spherically symmetrical (they include flattening and rotation). The resolution of differences between continents and oceans is very limited and the differences are averaged out in most cases. If some of the very-high-mode, i.e., short-wavelength, free oscillations can be observed, identified, and resolved in sufficient detail, information on gross structural differences between continents and oceans including associated mantle structure may be provided.

Many such higher-mode oscillations are akin to their traveling counterparts, *seismic surface waves*. Traveling surface waves offer the opportunity for determining earth structure based on one pass of the waves, as opposed to many passes with consequent averaging with other effects in the case of free oscillations. Surface waves have been used regularly in recent years to determine crustal and upper-mantle structure, including regional variations of such structure. Many of the measurements of depth to the top of the low-velocity zone in the mantle, which is presumably though not necessarily near or related to the base of the lithosphere, are derived from surface waves. Although the results are sometimes ambiguous and very. dependent on the validity of certain assumptions, there is probably a good deal more information about the earth, including the continental basement, to be obtained from surface waves. To utilize surface-wave data for continental structure will require, as a minimum, more complete and more closely spaced observations of the phenomenon, further attention to focusing and multipathing, integration of surface-wave observations with those of other phenomena, and further development of techniques for using models that depart from flat-lying layered structures.

In fact, the inadequacy of spherical or flat-layered models that portray the earth as lacking in lateral heterogeneity is growing in importance and has probably reached the crisis stage in the case of the continental basement. Our level of understanding has reached the point where refinement of such models may be more misleading than informative. As a first approximation, geophysicists have utilized simplified layered models, not only in the ease of seismic surface waves but also in many other kinds of seismic studies and in other branches of geophysics as well. This approach is not without good reason. Gravity is an important factor in geology, and hence there is indeed a strong tendency for spherical layering. Furthermore, layered models or one-dimensional models are relatively easy to handle from the theoretical point of view. Such simple models are no longer an adequate representation of the continental crust, where a bewildering variety of structures and rock types is found in almost any large outcrop (Figure 10.1). Figure 10.1, taken from a paper by Smithson *et al*. (1977), but in turn borrowed from Berthelsen (1960), shows pyroxene granulite layers surrounded by granitic gneiss. The important point in the present context has to do with the complex three-dimensional structure of this feature. Resolution of such complex buried contortions by seismic methods, or any methods, will be difficult, but on the other hand to apply flat-lying layered approximations to such structures is nonsense. We must develop models and observational and analytical techniques that will provide information on structures of complexity greater than that of simple layers and to the level of complexity shown in Figure 10.1, if possible. Such structures must be probed at depth within the crust, and perhaps the upper mantle.

Seismologists have been moving away gradually from simple-layered models for some time. Plate tectonics was a step in this direction. Perhaps more than anything, study of lateral variations and complex structures will characterize the application of seismology to the study of the continental basement during the 1980's.

Continuing through the spectrum to shorter wavelengths, consider the body waves generated by earthquakes, typically in the range of a few seconds per cycle to a few cycles per second. The *body-wave travel-time method*, which has produced so much of our knowledge of the earth's deep interior, has been applied widely on a crustal scale for study of the continents. However, the traditional, so-called near-earthquake studies have not produced much new information recently, partly because superior precision and flexibility may be had by use of artificial sources and partly because of the limitations of the flat-layered models usually used in interpretation of near-earthquake studies.

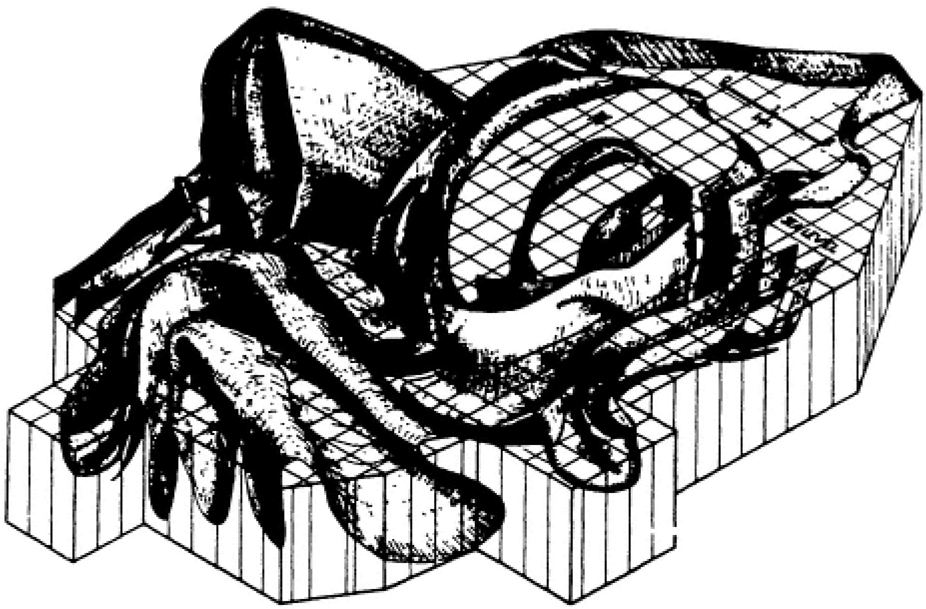


Figure 10.1
A three-dimensional structure in the continental basement. To approximate such structures by flat-layered models is of very. limited value. From Smithson et al. (1977) (taken from Berthelsen, 1960).

In recent years there have been some significant departures from the well-worn path, however. Tightly spaced detection networks have produced more precise locations of sources in time and space and more reliable travel-time data. Models involving lateral variations of structure such as faults can be considered using powerful computer-based methods. Techniques involving differential travel times at networks of stations encompassing a particular feature have led to investigation of features of unusual shape, the work, for example, of Aki *et al*. (1977) and the group at the U.S. Geological Survey (USGS) (Iyer, 1973; Ellsworth and Koyangi, 1977).

On the basis of the clear signs of life in current study of body waves, and the renewed emphasis that is likely to result from new efforts to predict earthquakes and reduce the earthquake hazard and to detect and identify nuclear explosions, one can foresee new results, particularly those emphasizing lateral variations, over the next decade at least. The keys for application of three-dimensional inversion of travel times are tightly spaced, probably movable, networks and developments in methods for using not only travel-time differences but also wave character resulting from attenuation, focusing, conversion, or other effects.

NATURAL SEISMIC SOURCES

In the case of natural seismic sources, perhaps as much information can be obtained from the sources as from the wave propagation. Historically, with each increase in precision of location of hypocenters of earthquakes, our understanding of tectonics has grown. For example, in the 1950's and earlier, hypocenters of earthquakes in the southern hemisphere were frequently mislocated by more than 100 km. At that stage, an earthquake could at best be associated with a major regional feature such as an island are. In the 1960's, with the advent of the World Wide Standardized Seismic Network (WWSSN) and other advances, it became possible to locate hypocenters with a precision of less than 10 or 20 km. Earthquakes could then be associated with tectonic features of smaller scale—a rift valley, a deep earthquake zone, the outer wall of a trench. Now, with the closely spaced observing networks that are available in a few areas, earthquake hypocenters can be located with a precision of less than a kilometer, much less in some cases. Hence we can now confidently associate earthquakes with particular faults, and also associate properties of the quake with properties of that fault. Fault-plane solutions tell us of the orientation of the fault and the movement; other focal mechanism data tell us of the change in stress and scale of the movement.

Further improvements in precision of hypocentral location may tell us about movements along a particular part of a fault, progression in fault activity, and complexities such as asperities and slices. Figure 10.2 shows how an interpretation of tectonics may be strongly dependent on the precision of hypocentral data. This figure is taken from a paper by Barazangi and Isacks (in press). A cross section of the seismicity through central and northern peru is shown twice. In the upper half of the figure only hypocenters of very high precision are plotted. In the lower half of the figure other hypocenters located during the same time period but with lower precision are shown. There is clearly a great contrast between the structure defined by the well-located hypocenters and what might be deduced from the less well-located hypocenters. A similar effect may be anticipated at smaller scale.

Other modem techniques are telling us of fault move ments that are slower than the abrupt displacements of typical earthquakes. Such slow movements may fail to generate short-period seismic waves, may generate only very long-period seismic waves, or may generate no detectable seismic wave and may be thought of as episodes of fault creep. Such studies enhance greatly our understanding of fault motion. The study of seismicity and sources is an area of vast potential and one in which integration of seismology and geology is likely to be crucial.

One may confidently state that interaction of seismology, or geophysics in general, and geology is likely to increase markedly in the next decade. A recent report by the NRC Committee on Seismology (1976) notes that much of the progress in understanding the problem of earthquake hazards in recent years has come from geological, not purely seismological, evidence. Plans to drill deeply into the San Andreas Fault are afoot. We can anticipate growing interaction between these disciplines, with mutual benefit as our understanding of the continental basement grows.

ARTIFICIAL SEISMIC SOURCES

In general, artificial sources have the advantage of precise timing, simplicity and control of source function, and flexibility of location and the disadvantage of lower energy except in the special case of nuclear explosions. Thus, nonnuclear artificial sources are currently of no value for study of the deep parts of the earth's interior, but for the continental crust and the upper mantle they can indeed provide information, and in fact the information with the best resolution at the level of detail that we now are seeking.

Seismic studies using artificial sources fall loosely and somewhat arbitrarily into three classes: (1) refraction, (2) deep seismic sounding (DSS) or refraction and wide-angle reflection, and (3) reflection profiling.

The *refraction method* is responsible for much of what is known about the structure of the deep crust; it provides the depth to the Mohorovicic* discontinuity and typically a simple model of crustal structure consisting of layers of different velocities. Seismic refraction studies have been carried out in the United States by various university and private groups and the USGS. However, almost all of this work was done in the 1960's and before. Figure 10.3, from Prodehl (1977), shows a summary of crustal models deduced from refraction data for various parts of the United States. Nearly flat-lying layered models prevail for each area, and the differences from one area to another are illustrated at the bottom of the figure. There is considerable similarity in all the crustal models, and, in fact, the rather uniform depth of the mantle throughout the continent is a rather striking feature. There are also substantial and measurable variations from one province to another on this gross scale. This figure is a summary of current knowledge of U.S. crustal structure based on refraction data. Surprisingly, for about the last decade there has been relatively little activity of this type directed toward study of the continental basement in the United States.

Abroad, however, the story is different. The Soviet Union has operated a program of crustal exploration at a very high level of activity since World War II and has developed *the DSS method*, in essence the use of refracted and wide-angle reflected waves observed with closely spaced stations, to determine crustal structure with resolution better than that obtained by refraction measurements alone. Furthermore and after a somewhat slower start, groups of European seismologists have been using similar techniques and obtaining detailed and abundant results. Figure 10.4 shows a typical set of seismic refraction data from Europe and a crustal model deduced from the data. Detailed velocity-depth function and a great deal of other seismic information are obtained that is not explained by the simple model. The work in the Soviet Union and in Europe is typically characterized by much more thorough observation through very closely spaced detecting stations than is the case for the older studies of this type in the United States. The papers in Giese *et al*. (1976) provide a comprehensive survey of explosion seismology in central Europe. Further seismic refraction studies in the United States will surely provide new information on the crust. The extent to which the new information will be coupled with other geological and geophysical information to provide enhanced understanding of the continents will depend on the further development of methods of interpretation to produce more realistic crustal models than the present simple layered configurations. The models should present a more realistic geological picture.

The newcomer on the scene of seismic exploration of the continental basement is *seismic reflection profiling*. In various countries, within the last decade, this technique has been applied with some modification to study of basement to depths as much as 50 or so kilometers, primarily by the Consortium for Continental Reflection Profiling (COCORP) project in the United States (see Oliver *et al*., 1976, for a review). The method uses closely spaced vibratory or explosive sources, listening arrays of thousands of geophones, and sophisticated computer analysis of the data. Typically, tens of millions of rays penetrating the earth are sampled and utilized at each site. The potential resolution of structural features of the deep crust through reflection profiling is greater than that of any other method, but it is no small operation. Figure 10.5 shows a field party in action. Of particular interest is that the sources of the seismic waves are not explosives but large truck-mounted vibrators. The VIBROESIS method (registered trademark of the Continental Oil Company) has been used in all the COCORP studies to date. The results from seismic reflection profiling so far clearly indicate vast potential for this method in future studies of the continental basement.

Figure 10.6 shows a seismic reflection profile across the Rio Grande rift. The abscissa is distance in decameters;

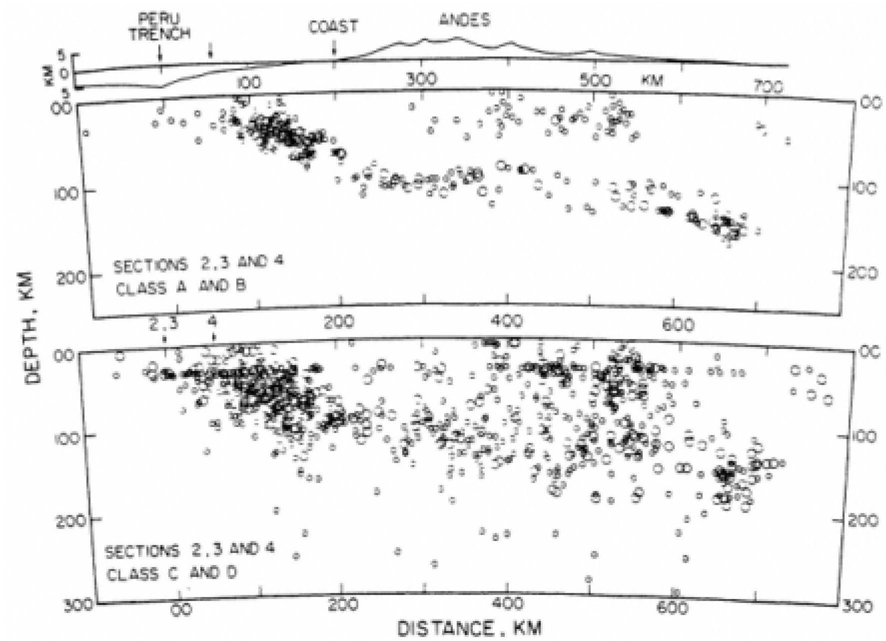


Figure 10.2
A generalized section through Peru showing by hypocentral locations of high precision in upper view and other hypocentral locations of lower precision in lower view, Combining these two sets of data can result in an interpretation different from one based only on the better data. From Barazangi and Isacks (in press).

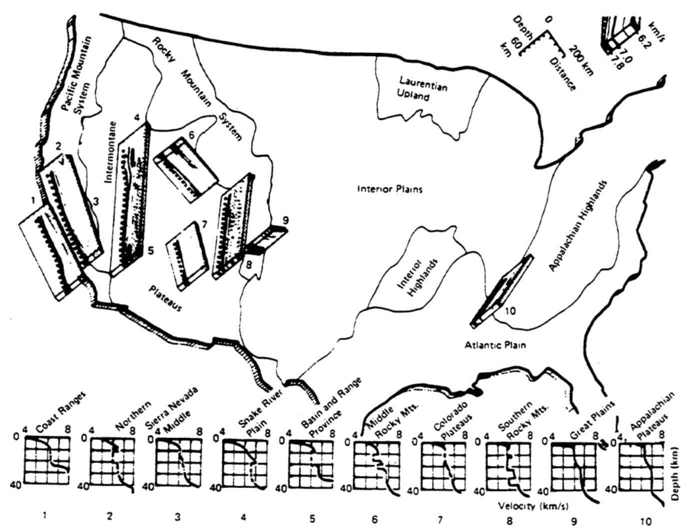


Figure 10.3
A summary by region of seismic refraction studies of the U.S. continental crust. Note (1) overall similarity, (2) variations from one province to another, (3) use of layered models. From Prodehl (1977, copyrighted by American Geophysical Union).

the ordinate is two-way travel time in seconds. To convert time in seconds to depths in kilometers approximately, multiply by 3. In a gross sense such a section may be thought of as a section of the earth, but a detailed interpretation of the data requires knowledge of wave propagation and the data processing involved, so that it is invalid to assume that the details of a time section such as the one shown are necessarily or precisely features of the deep earth. This section, for example, has not yet been subjected to migration to position the reflections in their proper spatial locations. Even so, certain gross features can be seen in the data that are illustrative of the capability of the method. The sedimentary sections and the sedimentary basement boundary are clear near the top of the section. There is an intragraben horst of substantial size and other more subtle evidence for faulting of the sediments. At the time of about 7-8 see there is a rather strong reflector that corresponds to the top of the magma body first proposed for this area by Sanford *et al*. (1977) on the basis of microearthquake data. Although this particular section does not show much information at a time corresponding to the crust-mantle boundary, other sections do.

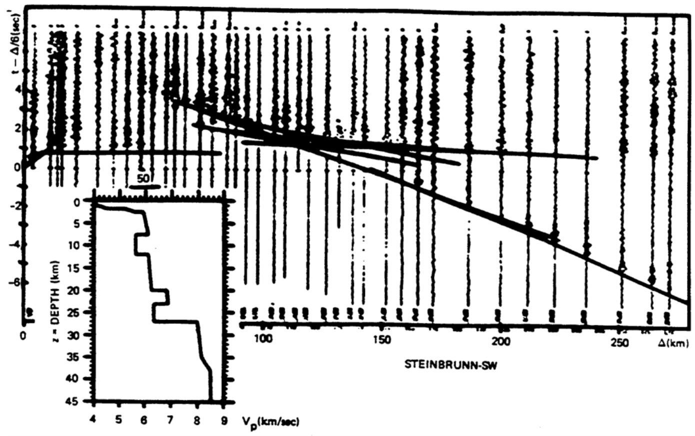


Figure 10.4
Typical record section, travel-time curves, and flat-layered model based on refraction study. From Mueller (1977, copyrighted by American Geophysical Union).

Figure 10.7 shows a line drawing of the data from the section in Figure 10.6 and another section to the east. The two sections together span the rift valley in the vicinity of Belch, New Mexico. The top of the basement and various sedimentary features are clear, as are certain other features within the basement itself. The magma body shows as a strong reflector in parts of both sections. At a time of about 12 see in the easternmost section there is an arrival that may be associated with the crust-mantle boundary, although it is not continuous for a very long distance. Other profiles in the Rio Grande riff area show a stratified pattern to the reflectors in the vicinity of this boundary. Much more detail can be found throughout the section in the original data.

Figure 10.8 shows a line drawing of a short section taken as a test of the method in the vicinity of the San

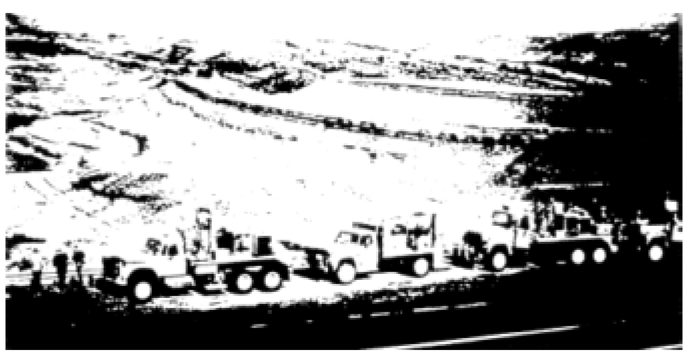


Figure 10.5
Seismic reflection profiling party in the field in Wyoming.

Andreas Fault. There is great contrast in the data from one side of the fault to the other. In the fault zone, to a depth of about 10 or 12. km, diffraction hyperbolas, associated with discontinuities marking the fault zone, are seen. Below that zone, however, there is a region of little information, suggesting that the zone was not penetrated by the seismic waves, which would be surprising in view of the information from much greater depths on both sides, or that the zone is so distorted through flow that no coherent reflected energy was obtained. The latter seems more probable. At a depth near the crust-mantle boundary, particularly in the western part of the section, there are many reflectors, suggesting a complex feature for this boundary.

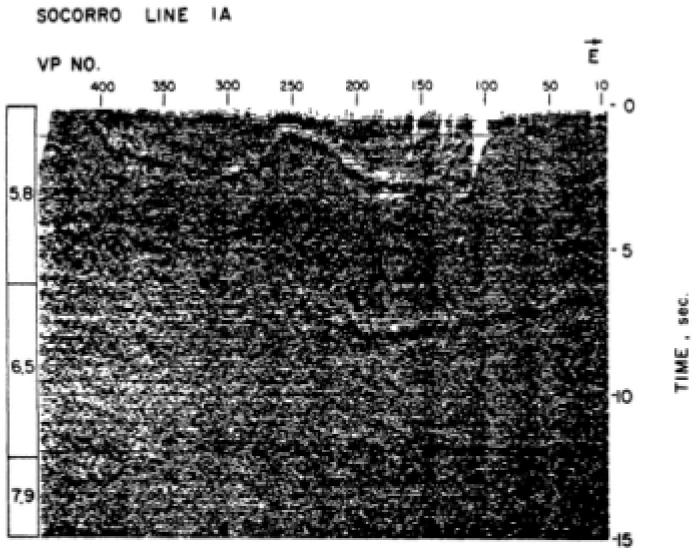


Figure 10.6
COCORP seismic reflection profile (unmigrated) across western part of Rio Grande rift north of Socorro. Abscissa is vibration point number or distance in decameters; ordinate is two-way travel time in seconds. Velocity structure from a nearby refraction study shown in column on left. Courtesy Larry Brown, Cornell University.

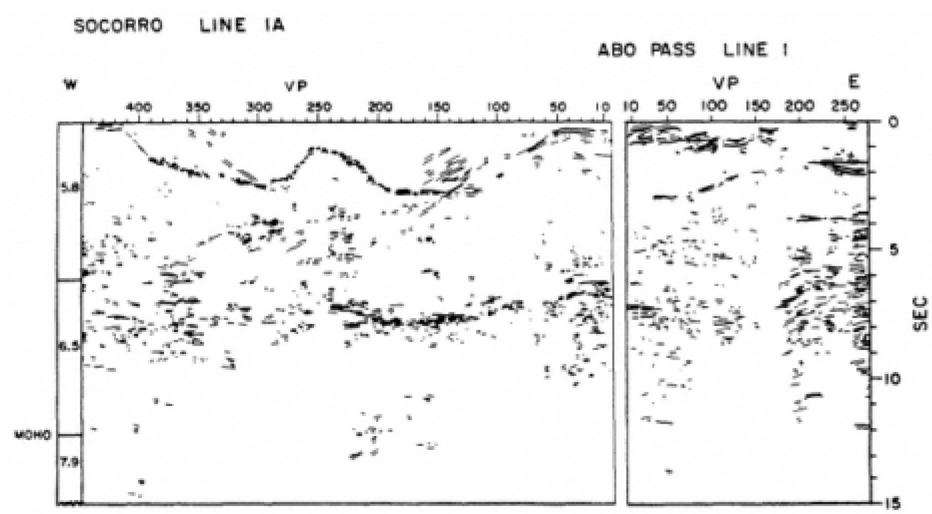


Figure 10.7
Line drawing based on data of Figure 10.6. See text for discussion. Column on left shows velocity structure determined from a nearby seismic refraction study.

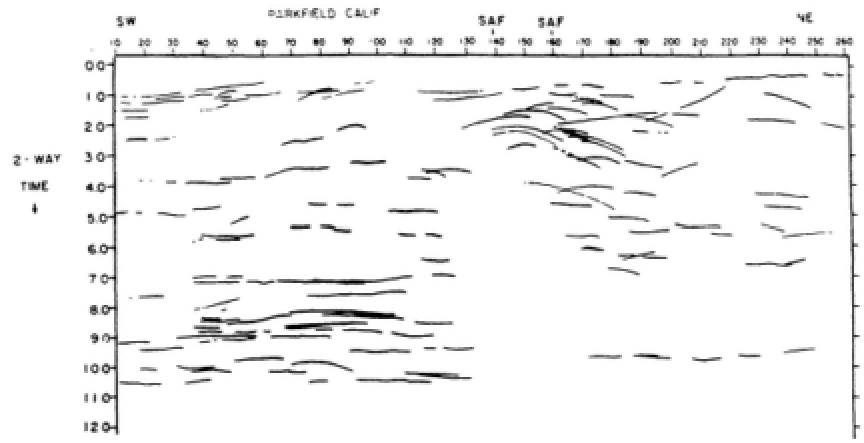


Figure 10.8

Line drawing of short test section crossing San Andreas Fault near Parkfield. Distance in decameters. Two-way travel time in seconds. S.A.F. indicates boundary of San Andreas Fault zone. See text for discussion.

Figure 10.9 shows a map of the southeasternmost portion of the Wind River range and indicates the line of a seismic reflection profile made in this area. Line drawings of parts of this section are indicated in Figure 10.10. There is a great deal of other information of considerable variety. The most prominent feature is the eastward extension of the thrust fault on the western boundary of the Wind Rivers Mountains. This fault extends to a depth of at least 25 km and more likely to 40 km without much change in dip. These data, then, seem to resolve the long-standing geological problem of the origin of the Wind Rivers in the sense that compressional forces seem to predominate as opposed to those producing vertical uplift. Substantial shortening of the crust is clear.

CONCLUSION

With the possible exception of study of the lower modes of free oscillations, all the seismic methods described above have something substantial to contribute to understanding of the continental basement. Furthermore, there are other useful methods and techniques that I will not discuss here, and in any case the dividing lines between the various methods are somewhat arbitrary. *My purposes are (1) to draw attention to the large unrealized potential for understanding the continental basement through application of present-day seismic methods and (2) to point to means for improving that potential through continuing evolution and improvement of seismological methods*. Suppose, for example, that a method could be found for artificially generating shear waves of sufficient amplitude so that the deep crust and upper mantle could be explored using such waves, or that movable, tightly spaced seismic arrays could be deployed so as eventually to cover the continent. Suppose the deep-sea floor, the continental shelves, and various remote areas were no longer areas of essentially no seismic data on earthquakes. Suppose that seismic profiling were conducted along long, closely spaced lines spanning the continents. Although this might seem like an ambitious undertaking to some, one should remember that it is only within about the last 30 years that there has been significant seismic exploration of the deep-sea floor and only about 20 years since reflection profiling of the seafloor began. The ocean basins, which, of course, occupy a much greater portion of the earth's surface than do the continents, have in that short time been crisscrossed innumerable times by seismic reflection profiling. A comparable achievement in the form of seismic profiling of the continents is within the grasp of the present generation of geophysicists. Suppose microprocessors and other electronic developments result in more sensitive and selective seismographs and new ways of managing large complex sets of data. Suppose the seismic reflection method is generalized so that three-dimensional images rather than two-dimensional sections are produced. Suppose comparable advances are made concurrently in related branches of geophysics and geology. Every one of these suppositions is technically within our reach today.

If even some of these suppositions are fulfilled, I think we can look forward to a new understanding of the crust, one in which deep subsurface structural features become as familiar to continental scientists as the midocean ridges, the trenches, the seamounts, and the submarine canyons have become to ocean scientists over the last few decades. When they are, surely our understanding of the evolution and the genesis of continents will become more

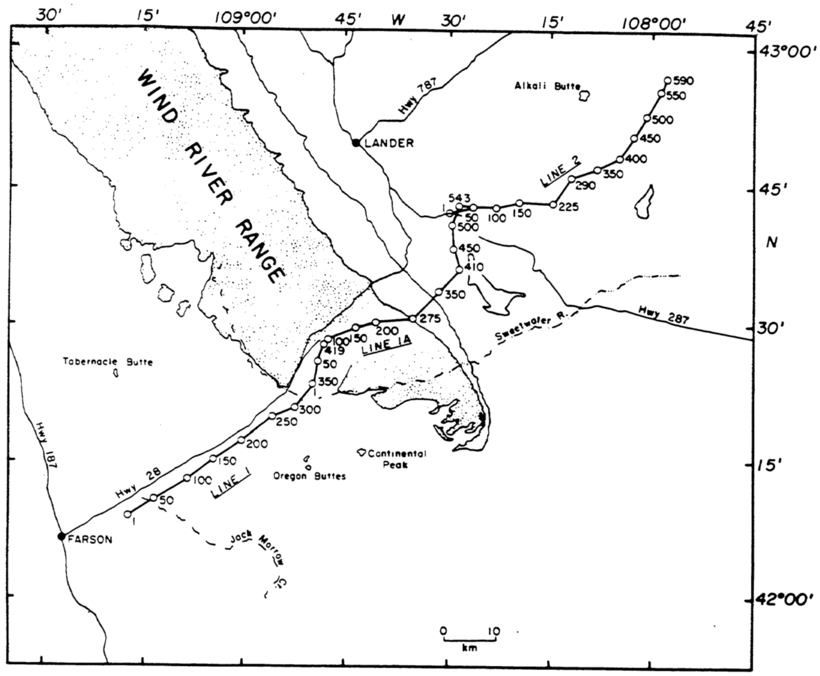


Figure 10.9
Map of southeastern Wind River Mountains in Wyoming showing location of COCORP seismic reflection line.

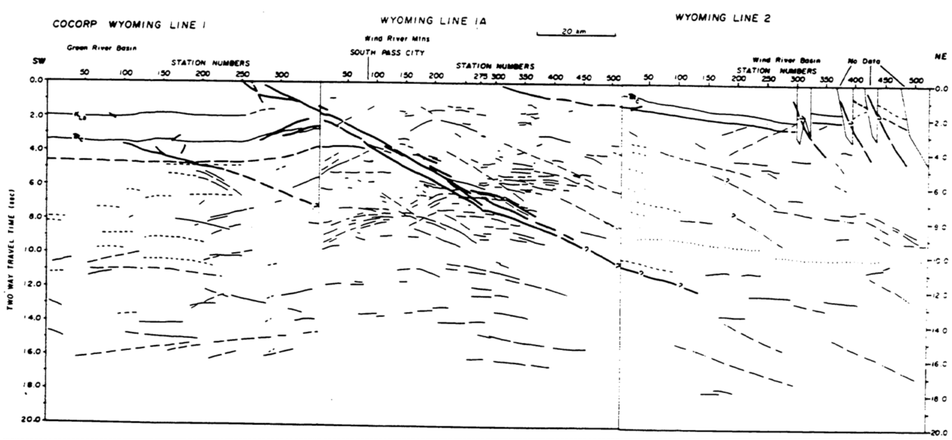


Figure 10.10
Line drawing of section along line of Figure 10.9. Note especially major thrust fault bounding the Wind River uplift on west. See text for discussion. Courtesy Jonathan Brewer and Scott Smithson, University of Wyoming.

profound, just as our understanding of the origin of the ocean basins became more thorough. Are these thoughts idle dreams never to be fulfilled? Somehow I feel the visiting astronaut would not see them that way but instead would view them as imminent and inevitable steps in man's progress toward a better scientific understanding of his planet—and a better life for its inhabitants.

ACKNOWLEDGMENTS

The author wishes to thank Muawia Barazangi, Bryan Isacks, Larry Brown, Jonathan Brewer, and Scott Smithson for permission to use figures from papers in preparation and various colleagues for constructive comments. The COCORP research is supported by National Science Foundation grants EAR 77-13653 and EAR 77-14674.

References

Aki, K., A. Christofferssen, and E. S. Husbye (1977). Determination of the three-dimensional seismic structure of the lithosphere, *J. Geophys. Res. 82*, 277-295.

Barazangi, M., and B. Isacks (in press). Subduction of the Nazca plate beneath Peru: evidence from spatial distribution of earthquakes, *Geophys. J.*

Berthelsen, A. (1960). Structural studies in the Precambrian of western Greenland, *Medd. Groenl. 123*, 1-222.

Ellsworth, W. L., and R. Y. Koyanagi (1977). Three-dimensional crust and mantle structure of Kilauea Volcano, Hawaii, *J. Geophys. Res. 82*, 5379-5394.

Giese, P., C. Prodehl, and A. Stein, eds. (1976). *Explosion Seismology in Central Europe, Data and Results*, Springer-Verlag, New York, 429 pp.

Iyer, H. M. (1973). Anomalous delays of teleseismic *P* waves in Yellowstone National Park, *Nature 253*, 425-427.

Mueller, S. (1977). A new model of the continental crust, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 289-317.

NRC Committee on Seismology (1976). *Predicting Earthquakes*, National Academy of Sciences, Washington, D.C., 62 pp.

Oliver, J, M. Dobrin, S. Kaufman, R. Meyer, and R. Phinney (1976). Continuous seismic reflection profiling of the deep basement, Hardeman County, Texas, *Bull. Geol. Soc. Am. 87*, 1537-1546.

Prodehl, C. (1977). The structure of the crust-mantle boundary beneath North America and Europe as derived from explosion seismology, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 349-369.

Sanford, A. R., R. P. Mort, Jr., P. J. Shuleski, E. J. Rinehart, F. J. Caravella, R. M. Ward, and T. C. Wallace (1977). Geophysical evidence for a magma body in the crust in the vicinity of Socorro, New Mexico, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 385-403.

Smithson, S. B., P. N. Shive, and S. K. Brown (1977). Seismic velocity, reflections, and structure of the crystalline crust, in *The Earth's Crust*, J. G. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 254-270.

11

Exploration of the Continental Crust Using Aeromagnetic Data

Isidore Zeitz
U.S. Geological Survey

INTRODUCTION

It is significant that many countries have undertaken aeromagnetic mapping programs to obtain complete coverage of their nations in a reasonable period of time. The Soviet Union, for example (Solov'yeva, 1968), had complete coverage by 1967 using a flight-spacing of 2 km. The Canadian Geological Survey had initiated a program to map the entire Canadian Shield at a flight separation of 1/2 mile (804 m). Almost three quarters of the shield has now been flown, and the maps have been made available to the public (Hood, 1974). Yet, at this writing, there is no federal program to map the United States aeromagnetically. To date, published aeromagnetic maps are available for only 20 percent of the conterminous United States. Although at least half of the country has been adequately surveyed by private industry, the results have not been made available to the public. Most of the available coverage in the United States has been provided by the U.S. Geological Survey, the funds often coming from other organizations, such as state geological surveys. For the past few years, the U.S. Geological Survey has adopted a policy of promptly releasing aeromagnetic data to the public via the open-file route. These recent publications are mostly at a scale of 1:250,000.

The obvious advantages of the airborne method over ground measurements are the speed and economy of the survey, the coverage of areas that are inaccessible on the ground, and the nature of the data—recorded continuously along a profile rather than as discrete points.

An advantage of the magnetic method is that it permits the mapping of the basement crystalline rocks underneath a cover of sedimentary rocks, for the latter are nonmagnetic. The specifications of an airborne magnetic survey are dictated by the nature of the problem. For detailed surveys, the detector should be close to the ground and the separation between flight lines small. For regional and crustal geological studies, the flight elevation should be higher, and the flight spacing larger.

Aeromagnetic data may reflect rock distributions anywhere from the surface to depths at which the Curie point of the magnetic rocks is reached. In the continental crust, this may vary anywhere from 15-50 km, depending on heat flow and the thermal properties of the rocks. The

Figure 11.1
Aeromagnetic map of Nevada and Utah (from Stewart et al., 1977, courtesy the Geological Society of America).

depth of penetration is not so great as those reached by other geophysical methods, such as seismic, gravity, or magnetotelluric methods. However, the magnetic method, which "sees" the shallow rocks, in combination with the aforementioned geophysical methods, which "see" the deep rocks, provides a powerful tool in evaluating the total crust.

In this paper, I will discuss selected aeromagnetic surveys and their geological interpretations for areas in the western United States, the midcontinent, and the eastern United States. I selected these three provinces because they reflect three distinctly different phases of crustal evolution—the tectonically active area in the west, the old and stable crust of the midcontinent, and the Appalachian province in the east.

WESTERN UNITED STATES

The work of Blake *et al*. (1978) is particularly significant in that it attempts, by examining available aeromagnetic data, to correlate a number of ophiolite belts in California and to relate these belts to former plate boundaries. This is possible because the serpentinized ultramafic rock, which makes up a large volume of the California ophiolites, has unusually high magnetic susceptibilities. Blake *et al*. (1978) suggest that the ophiolite belts for the entire western margin of both North and South America could be mapped, provided adequate aeromagnetic coverage were available. In addition, the authors suggest that the Great Valley anomaly may be related to the emplacement of ultramafic and mafic rock in a marginal basin behind a Late Jurassic island arc.

Griscom (1973) deduces some important tectonic relationships from aeromagnetic data at the junction of the San Andreas Fault and Mendocino fracture zone. "Straight magnetic lineaments on the continental shelf west of the San Andreas fault between Point Arena and Cape Mendocino have northwest trends, and are interpreted to be caused by ophiolite belts. An east-west magnetic anomaly associated with the Mendocino fracture zone extends from the deep ocean inland at Punta Gorda, a distance of 20 km. During the time this anomaly has existed, the San Andreas fault cannot have extended north of Cape Mendocino, and lateral movement between the oceanic and continental plates cannot have taken place near Cape Mendocino."

A compilation of all available aeromagnetic data in Nevada and Utah (Stewart *et al*., 1977) clearly shows east-west trends (Figure 11.1, A-A', B-B', C-C'), in spite of the fact that the basin and range structures trend north-south. Geological investigations subsequent to this discovery by aeromagnetic mapping showed that these east-west trends are correlatable with Cenozoic igneous rocks and that significant mineral deposits are aligned along these same belts in eastern Nevada and western Utah.

In a regional sense, the aeromagnetic map over the Basin and Range province in Nevada and Utah is especially "quiet" when compared with the magnetic data elsewhere over the continental United States. In eastern Utah, the amplitude of the anomalies is an order of magnitude greater than those to the west. If these anomalies are related to Precambrian rocks, the magnetic data might indicate their western limit (Figure 11.1, D-D').

The major volcanic areas of the Columbia Plateau, Snake River Plain, Yellowstone Park, High Cascades, and the Tertiary basalt areas of coastal Oregon and Washington all show characteristic magnetic patterns and anomalies.

A high-altitude (15,000-ft barometric) aeromagnetic survey with widely spaced flight lines (5 miles) for a strip across the northwestern United States (Zietz *et al*., 1971) proved useful in studying the gross features of the earth's crust. The strip is bounded approximately by latitudes 45º30' N and 47º00' N, and extends from the Pacific Ocean approximately 92 km from west of the coast, eastward to and including the Great Plains in Montana. The results are spectacular in that the magnetic map is marked by conspicuous northeast and northwest anomaly trends, lineaments, and breaks in the anomaly pattern. Their regional distribution, overall magnetic character, and geological evidence suggest that they are major structural features in the basement rocks. The close correspondence of structural and geological features in younger rocks with these basement magnetic and structural trends suggests that basement trends controlled, or at least greatly influenced, intrusion, deposition, and structural history of younger rocks. In some cases, evidence suggests that basement structures have been reactivated during later tectonic activity. Perhaps even more striking than the northeast-and northwest-trending features are large east-west magnetic discontinuities that, in some cases, extend completely across the strip to the edge of the shelf and that, in some cases, can be correlated with large-scale discontinuities dating back to the Precambrian.

MIDWESTERN UNITED STATES

A compilation of magnetic data (Plate 11.1) for a large area in the northern part of the midwestern United States shows the characteristic magnetic patterns typical of the Precambrian stable craton of North America. The map was prepared by reducing large-scale aeromagnetic maps and removing the earth's main magnetic field (International Geomagnetic Reference Field). The residual map was colored at 200-gamma intervals, the magnetic intensities varying in a rough way with the colors of the optical spectrum. Black-and-white presentations of aeromagnetic maps are important in structural geological interpretations but are less helpful in evaluating lithologies. The colored map adds another dimension to the original data and is indispensable in the discrimination of lithologic units. In addition, the colored map acts as a filter as it shows broad regional anomalies that would be indiscernible in black-and-white presentations. The amplitudes of the anomalies within the area are very high, in marked contrast to the smaller amplitudes of anomalies shown in the aeromagnetic map of Nevada and Utah.

One of the more prominent features on the map extends from Lake Superior, passing through the entire states of Minnesota and Iowa and part of Nebraska. It corresponds to the midcontinent gravity high (Figure 11.2) originally described by Woollard (1943), probably the most outstanding feature on the gravity map of the United States. A more detailed aeromagnetic map of this feature, using a contour interval of 100 gammas, is shown in Figure 11.3. The combined geophysical data, surface outcrops in the Lake Superior district, and drilling data suggest the existence of a thick sequence of basalts filling a trough approximately 40 miles wide and 4 miles deep, extending from the Lake Superior district (where extensive areas of basalt of Keweenawan age are known to crop out) to Kansas. Basalt crops out in only a very small area in the northeast part of the map in northwestern Wisconsin. The rest of the area is covered by a sedimentary se

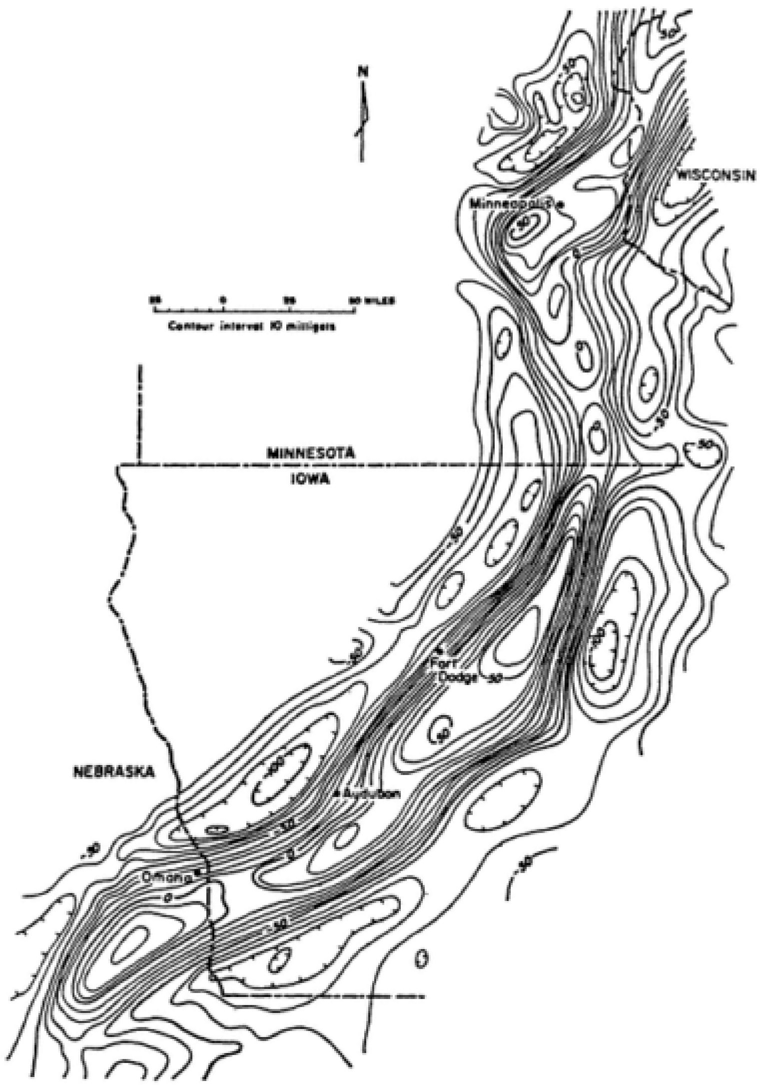


Figure 11.2
Midcontinent gravity high (from Zietz, 1969, copyrighted by American Geophysical Union).

quence that gradually thickens in a southerly direction to a maximum thickness of approximately 5000 feet in Nebraska. The presence of basalt in the Precambrian is verified by 11 drill holes in Minnesota and Iowa, all of which bottom in basalt. The flows are bounded by high-angle faults on both sides and are flanked by a series of elongated basins containing red elastic sandstones 1 or 2 miles thick (King and Zietz, 1971).

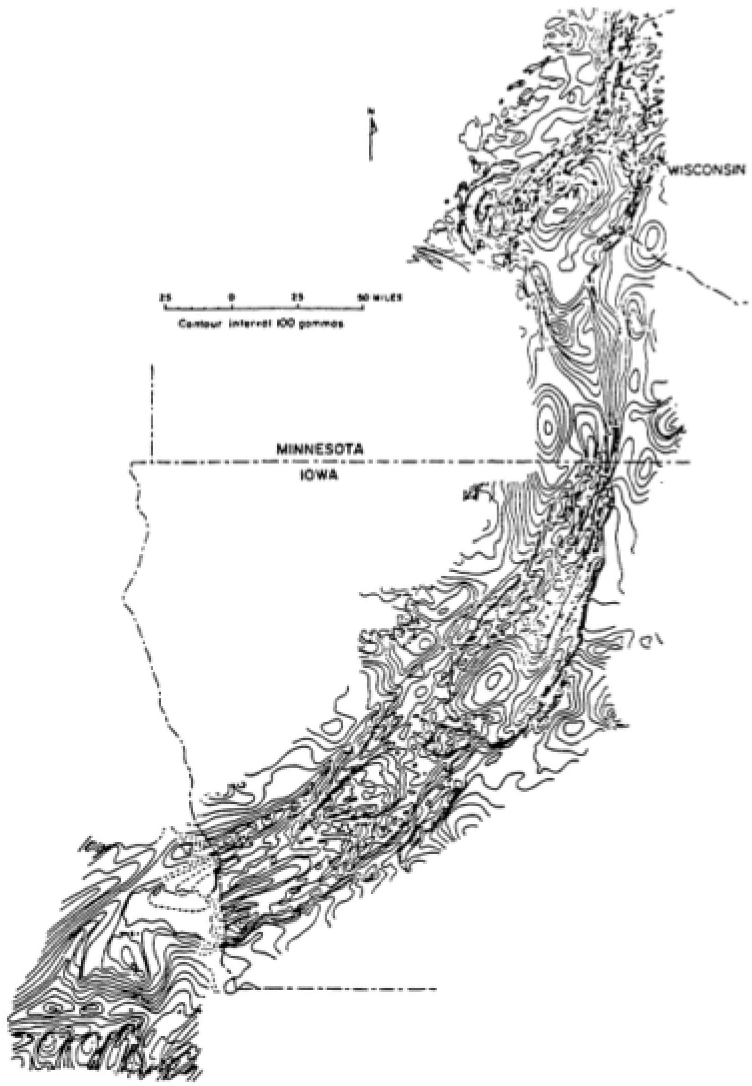


Figure 11.3
Aeromagnetic map of midcontinent gravity high (from Zietz, 1969, copyrighted by the American Geophysical Union).

The geophysical data suggest that the basalt flows were subsequently faulted and folded. The trough probably represents a major rift in the older Precambrian crust and appears to be discordant with the older structures. This midcontinent riff may well have been part of a Keweenawan global rift system, with initial offsets consisting of transform faults along pre-existing fractures, but apparently it never fully developed laterally into an ocean basin (see Chapter 4), and the upwelling mafic material was localized along a relatively narrow belt.

Much geological information can be extracted by the combined use of the aeromagnetic and gravity data to

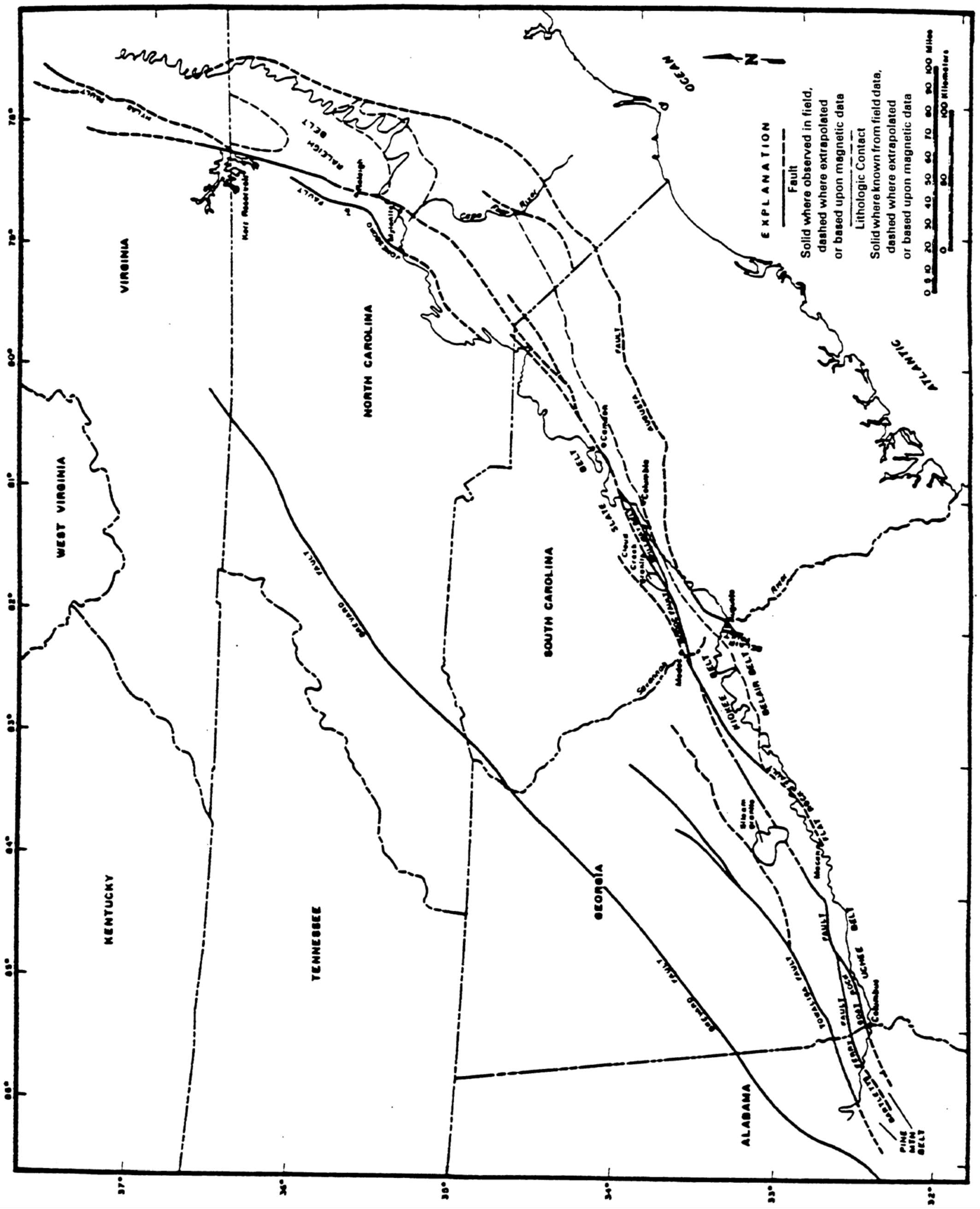


Figure 11.4
Distril, in of Eastern Piedmont fault system (from Hatcher et al., 1977, reprinted from Geology, with permission).

make a geological map of the buried Precambrian surface and to construct geological cross sections at right angles to the strike of the flows.

In northern Minnesota and eastern South Dakota, northeast-trending broad-wavelength magnetic anomalies, on the order of 30-40 km, are clearly discernible (Plate 11.1). These alternating bands of highs and lows can be traced to the northeast part of the state, where they can be correlated with rocks that crop out at the surface. Surprisingly, the positive magnetic anomalies reflect granite gneisses, and the negative anomalies are identified with a greenstone terrane. This correlation is consistent with geophysical interpretations published for the Canadian Shield but inconsistent with the commonly accepted generalization that felsic rocks produce magnetic lows, and mafic rocks magnetic highs. A detailed gravity survey of the entire State of Minnesota corroborated our predicted results. The outlines and shapes of the gravity and magnetic anomalies match splendidly but have an inverse correlation, i.e., gravity high with magnetic low and gravity low with magnetic high. We do not yet understand why certain major metamorphic belts are magnetic, but the answer is probably a combination of original composition and the chemical potential of oxygen (i.e., the chemical conditions in effect during metamorphism). We do know, however, that the granitic gneisses, being less dense than the host rocks, should give gravity lows, whereas the greenstones, which are more dense than the host rocks, should produce gravity highs. The use of this geophysical approach (combined magnetics and gravity) has broad ramifications in other areas of the midwestern United States. It should be possible to produce lithologic and possibly structural maps of the Precambrian basement surface by the proper analysis of the geophysical data when used in conjunction with all available drill-hole data on basement cores.

In the southwestern part of the map (Plate 11.1), a broad-wavelength magnetic trough, indicated by the deep blue area, covers all of Iowa. I suspect that it is caused by anomalous conditions deep in the crust. It may be that the Curie point is elevated in this area. Additional geophysical investigations to probe into the deep crust, such as deep resistivity or magnetotelluric soundings, would be helpful.

Since the compilation of this map, a detailed aeromagnetic survey was conducted of the entire Precambrian shield area in northern Wisconsin. The flight spacing was 1/2 mile, and the flight elevation was 1000 feet barometric. The aeromagnetic maps (1:250,000 scale), both in black and white and in color, were placed on open rile by the U.S. Geological Survey (uses), and thus made available to the public. At the time of this survey, geological mapping of this area of Wisconsin was sketchy, except for local areas (see, for example, the lithologic map of Precambrian rocks in Wisconsin produced by Dutton and Bradley, 1970). Through the use of the aeromagnetic map and a gravity map based on widely spaced observation points, a new geological map (1:250,000 scale) was produced by Sims *et al.* (1978). It is significant that, to produce the geological map, no additional field mapping was necessary. This new map is an order of magnitude more detailed than the previous one and is summarized by the abstracts of Sims and Peterman (1978) and Cannon (1978), presented at the 24th annual meeting of the Institute on Lake Superior Geology in May 1978.

EASTERN UNITED STATES

Except for the State of Maine, the entire exposed Appalachian crystalline belt from Alabama to Newfoundland has been flown aeromagnetically, with a line spacing of one mile or closer. In addition, most of the Coastal Plain of the eastern United States has been surveyed aeromagnetically at one-mile spacings. The survey covers the area from east of the Fall Line to the shore line and from the Potomac River to the approximate latitude of Ocala, Florida.

The magnetic anomalies over the crystalline rocks are distinctly linear, probably reflecting tectonic trends. These linear trends may be caused by belts of magnetic plutons (either mafic or granite), belts of certain amphibolites and magnetite-bearing schists, or belts of mafic and ultramafic rocks that may represent rifts, sutures, or former subduction zones.

Because of the linearity of the aeromagnetic trends, aeromagnetic maps have proven useful as an aid to field mapping over the Appalachian crystalline rocks. They are especially useful in the southeast United States, where outcrops are sparse and much of the area is covered by sapprolite.

The USGS has initiated a program to publish aeromagnetic maps over the entire crystalline belt in black and white and in color, at a scale of 1:250,000. Each of these maps will be accompanied by a geological interpretation. South of New York City, new geological maps will be based primarily on the aeromagnetic data but will obviously include all existing geological information. Radioactivity maps, when used in conjunction with the aeromagnetic data, are particularly useful in surficial mapping. Whereas the felsic rocks are generally less magnetic than the mafic rocks, the reverse is true for radioactivity, the granitic rocks being more radioactive.

For the Appalachian orogen, the literature is becoming inundated with papers dealing with plate tectonics, that is, those related to the opening and closing of the Atlantic Ocean. Any such model would have to be continental in scope. Unfortunately, reliable field investigations are limited to small areas, on the order of a few 15' quadrangles, and each author can usually accommodate his mapping to fit the plate-tectonic model of his choice. These 1:250,000-scale aeromagnetic maps and the accompanying regional geology covering very large areas, should rill an important gap in arriving at a more meaningful model.

An important paper by Hatcher *et al.* (1977) is typical of the use one can make of aeromagnetic data in the southeast. To quote from the abstract:

> Geologic mapping, interpretation, and field checking of recent aeromagnetic data suggest the existence of a closely associated series of faults and splays extending from Alabama to Virginia, herein termed the Eastern Piedmont fault system. Characteristic magnetic anomalies were found to be associated with known faults, and were used to trace them through covered intervals. The fault system extends northeastward from the Goat Rock fault of Alabama and west-central Georgia, crossing the lower Piedmont of South Carolina, passes beneath a segment of the Coastal Plains in the Carolinas, and then flanks the Raleigh belt in North Carolina and continues into Virginia. From east-central Georgia to Virginia, cataclastic rocks along the faults of the system are bounded to the northwest and southeast by rocks of the Carolina slate belt, forming perhaps the most extensive fault system in eastern North America.

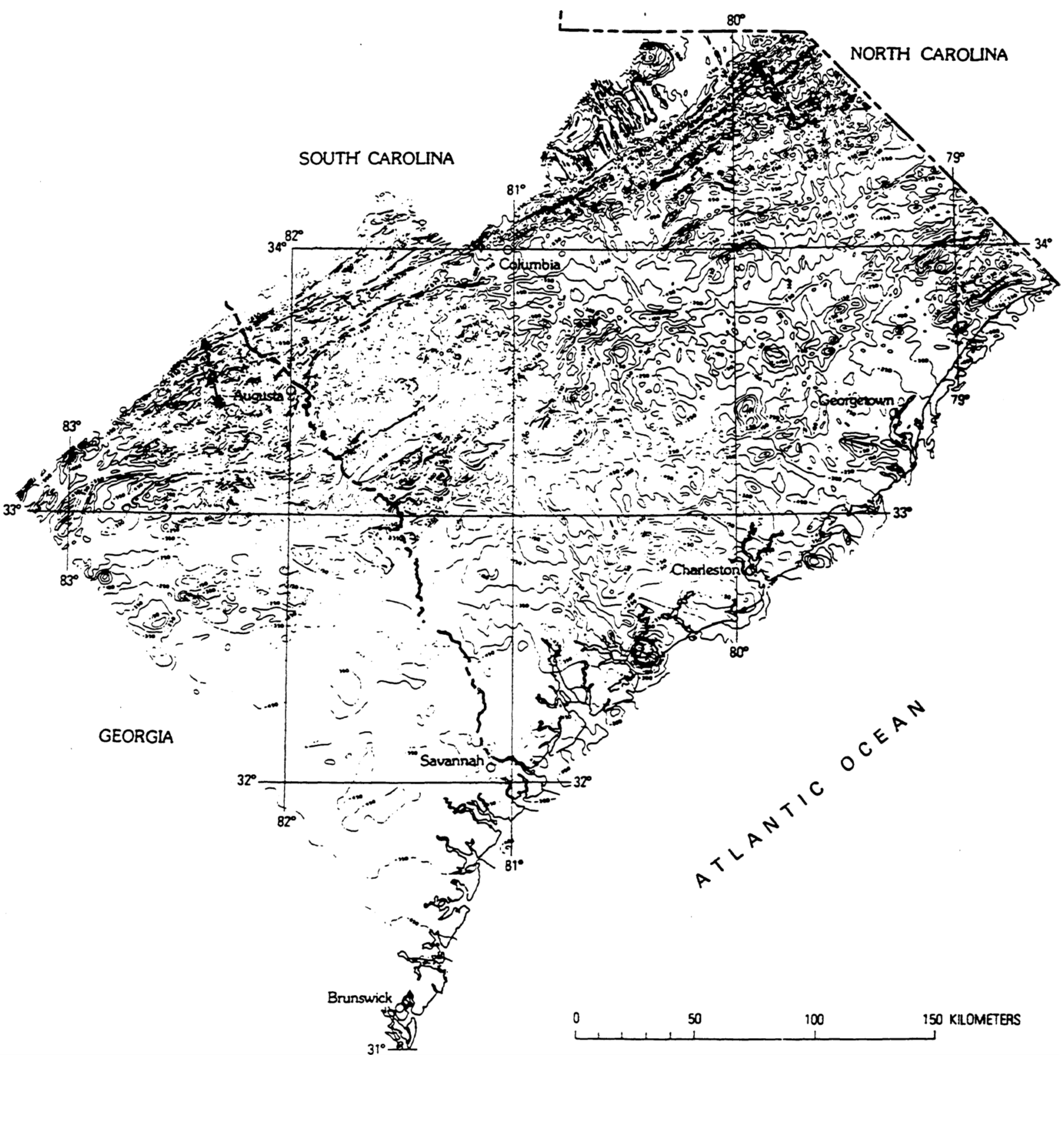


Figure 11.5
Generalized aeromagnetic map of southeastern South Carolina and eastern Georgia. Contour interval is 100 gammas (from Popenoe and Zietz, 1977).

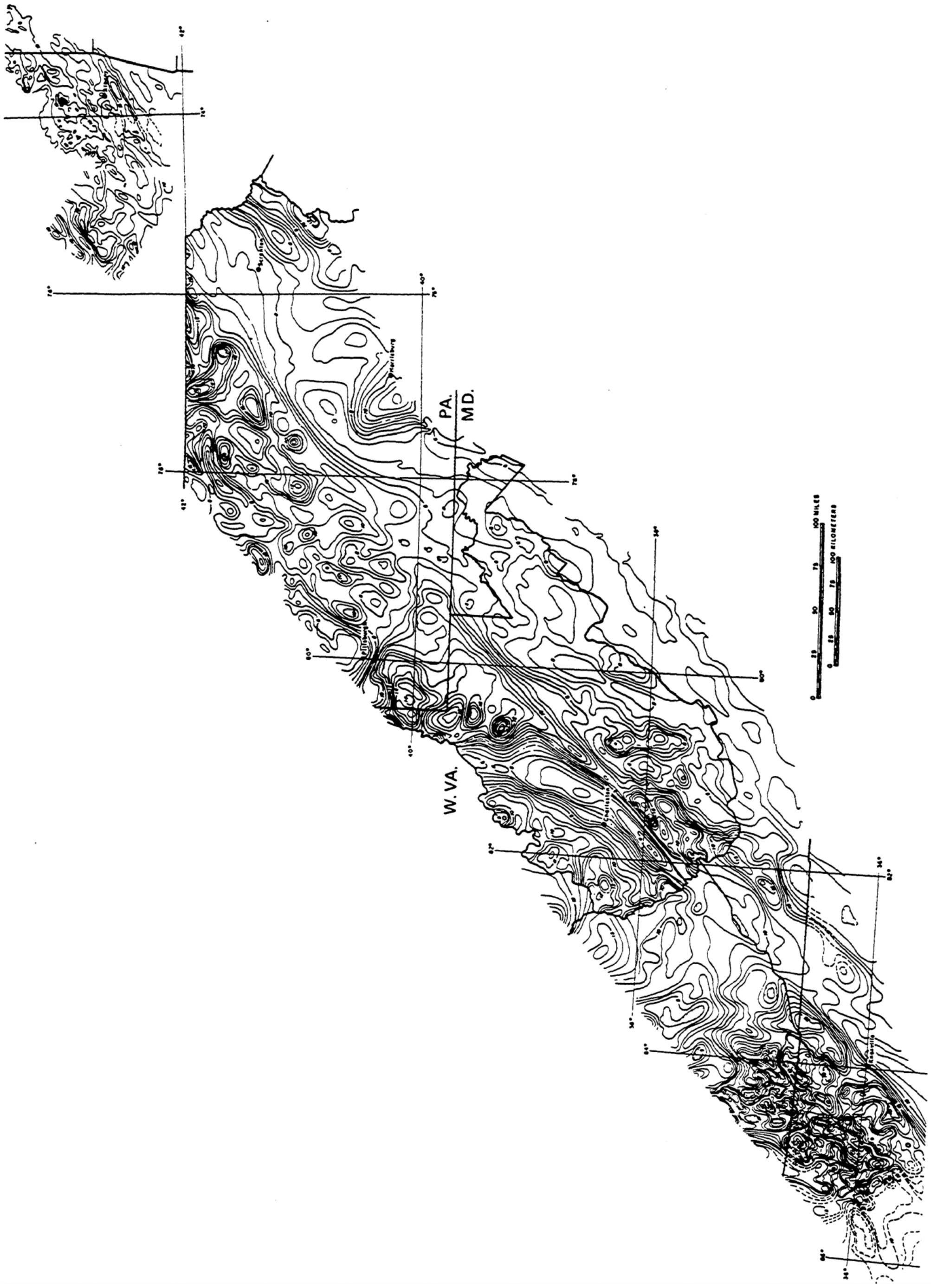


Figure 11.6
Composite aeromagnetic map of the New York-Alabama lineament (from King and Zeitz, 1978, courtesy of the Geological Society of America).

The Eastern Piedmont fault system is shown in Figure 11.4. The aeromagnetic data on which the fault system is based shows anomalies of unusual linearity and small amplitude (tens of gammas). Aeromagnetic maps for a small area of the system are shown in Figure 11.5 (A-A') (Popenoe and Zietz, 1977).

One of the more significant magnetic features offshore is the East Coast anomaly, which is a prominent linear magnetic high that follows the scarp at the eastern North American Continental Shelf edge. Off the Georgia coast, the anomaly turns westward, crosses the Shelf, and comes ashore near Brunswick, Georgia. This segment is called the Brunswick anomaly.

The magnetic data combined with gravity, heat flow, and drill-hole data show quite clearly that the Brunswick anomaly is axial to the Southeast Georgia Embayment, one of the largest Atlantic Coastal Plain sedimentary basins, with a history of subsidence that was active throughout the Cenozoic and is still taking place.

Seismic profiles and computed profiles and models from the aeromagnetic and gravity maps indicate that the anomaly-causing body is a mafic intrusion at or just below the top of the basement beneath the Coastal Plain. The intrusive body and the volcanic rocks form what is probably an anomalous crest separating the continental rocks of Georgia from those of northern Florida. The intrusive body probably marks an old rift zone, and thermal contraction of the anomalous crest has caused subsidence of the Southeast Georgia Embayment. The Brunswick and East Coast anomalies probably mark the old edge of the North American continent.

One of the longest magnetic lineaments in the North American continent (Figure 11.6) is prominently displayed along the length of the Appalachian basin (King and Zietz, 1978). It doubtless reflects a discontinuity in the basement, trending in a northeasterly direction for 1600 km from the Mississippi Embayment to the Green Mountains. King and Zietz have called this magnetic feature the New York-Alabama lineament. It is marked by a series of linear magnetic gradients that bound areas of magnetic rocks in the basement, implying strike-slip displacement along a profound crustal break. The lineament tends to coincide with the west side of the regional Appalachian gravity low (Figure 11.7) and separates a province of dominantly north-trending gravity anomalies on the northwest from a province of prevailing northeast trends on the southeast. The lineament seems to mark the southeast edge of a stable crustal block that acted as a buttress for the strong deformation of the Appalachian fold belt, so that the arcuate salients in Pennsylvania and Tennessee are tangential to it. Present-day seismic activity shows a correlation of the location of the lineament (Figure 11.8) with an active area to the southeast in the eastern part of the Appalachian basin and a seismically inactive block along the northwest side of the lineament. A parallel active zone to the west extends from New Madrid, Missouri (see Chapter 7) to the valley of the St. Lawrence River.

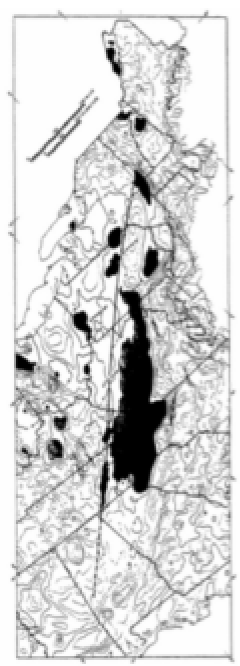


Figure 11.7
Bouguer gravity map of area near lineament (from King and Zeitz, 1978, courtesy of the Geological Society of America). Contour interval 10 mgal.

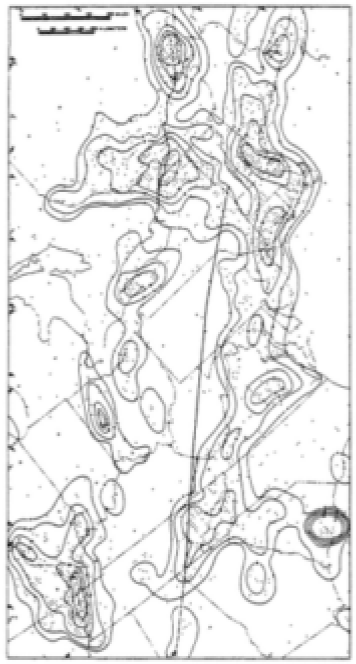


Figure 11.8
Seismotectonic map of eastern United States (from King and Zeitz, 1978, courtesy of the Geological Society of America).

The lineament is in the subsurface extension of the Grenville province, which may mark a region of a former continental collision analogous to the India-Asia region. Here, strike-slip faults of great length and linearity and large horizontal displacement, such as the Altyn Tagh Fault of Tibet, typify the region north of India and may have their counterpart in the New York-Alabama lineament, which records probable strike-slip displacement of great magnitude in the basement rocks under the Appalachian basin.

CONCLUSIONS

With just a few selected examples in the United States, this paper attempts to show the significance of the aeromagnetic method in evaluating the upper part of the earth's crust. Because crystalline rocks usually have a magnetic signature associated with them, the magnetic method can be used in geological mapping for both lithology and structure. In order to integrate on a regional scale detailed mapping of local areas, often done by different people, using vastly different approaches, aeromagnetic maps provide a common denominator by relating stratigraphic units defined using different criteria and tracing units through unmapped areas.

References

Blake, M. C., Jr., I. Zietz, and L. Daniels (1978). Aeromagnetic and generalized geologic map of parts of central California, U.S. Geol. Surv. Inv. Map GP-918.

Cannon, W. F. (1978). A middle and late Precambrian fault system in northern Wisconsin and northern Michigan (abstr.), *Proceedings of the 24th Annual Meeting,* Institute on Lake Superior Geology, May 1978.

Dutton, C. E., and R. E. Bradley (1970). Lithologic, geophysical and mineral commodity maps of Precambrian rocks in Wisconsin, U.S. Geol. Surv. Misc. Field Inv. Map 1-631.

Griscom, A. (1973). Tectonics at the junction of the San Andreas Fault and Mendocino fracture zone from gravity and magnetic data, in Conference on Tectonic Problems of the San Andreas Fault System, Proceedings, *Stanford U. Publ. Geol. Sci. 13,* 383-390.

Hatcher, R. D., Jr., D. E. Howell, and P. Talwani (1977). Eastern Piedmont fault system: speculations on its extent, *Geology 5,* 636-640.

Hood, P. (1974). Mineral exploration, trends and development in 1973, *Can. Min. J. 95* (3), 1-20.

King, E. R., and I. Zietz (1971). Aeromagnetic study of the mid-continent gravity high of central United States, *Geol. Soc. Am. Bull. 82,* 2187-2208.

King, E. R., and I. Zietz (1978). The New York-Alabama lineament: geophysical evidence for a major crustal break in the basement beneath the Appalachian basin, *Geology 6,* 312-318.

Popenoe, P., and I. Zietz (1977). The nature of the geophysical basement beneath the Coastal Plain of South Carolina and northeastern Georgia, in Studies Related to the Charleston, South Carolina, Earthquake of 1886—A Preliminary Report, W. Rankin, ed., *US. Geol. Surv. Prof. Paper 1028,* 119-137.

Sims, P. K., and Z. E. Peterman (1978). Precambrian geologic framework of northern Wisconsin (abstr.), *Preceedings of the 24th Annual Meeting,* Institute on Lake Superior Geology, May 1978.

Sims, P. K., W. F. Cannon, and M. G. Mudrey Jr. (1978). Preliminary geologic map of Precambrian rocks in part of northern Wisconsin, *U.S. Geol. Surv. Open-File Rep. 78-318.*

Solov'yeva, N. M. (1968). Karta anomal' nago magnitnogo poly territorii, S.S.S.R. 1968 (Map of the anomalous magnetic field of the territory of the U.S.S.R.), Moscow: Bsesoyuznyy Nauchno-Issled-Ovatel'skiy Geologisheskiy Institue (Vsyegyei) (1:10,000,000).

Stewart, J. H., W. J. Moore, and I. Zietz (1977). East-west patterns of Cenozoic igneous rocks, aeromagnetic anomalies, and mineral deposits, Nevada and Utah, *Geol. Soc. Am. Bull. 88,* 67-77.

Woollard, G. P. (1943). Transcontinental gravitational and magnetic profile of North America and its relation to geologic structure, *Geol. Soc. Am. Bull. 54,* 747-790.

Zietz, I. (1969). Aeromagnetic investigations of the earth's crust in the United States, in *The Earth's Crust and Upper Mantle,* P. J. Hart, ed., Am. Geophys. Union Geophys. Monogr. 13, pp. 404-415.

Zietz, I., B. Hearn, Jr., M. W. Higgins, G. D. Robinson, and D. A. Swanson (1971). Interpretation of an aeromagnetic strip across the northwestern United States, *Geol. Soc. Am. Bull. 82,* 3347-3372.

12

Chemistry of the Lower Crust: Inferences from Magmas And Xenoliths

Robert W. Kay and Suzanne Mahlburg Kay
Cornell University

INTRODUCTION

In any discussion of the lower crust, deep crustal xenoliths and crustal-derived or crustal-contaminated magmas should receive an important share of attention. This rationale is directly analogous to the interest in peridotites and basalts as probes of the upper mantle. Studies of crustal-derived magmas and xenoliths from different regions indicate regional variations in the composition, age, and thermal history of lower crustal regions. Coordination of xenolith and geophysical studies has the potential of defining subsurface map units of regional extent. More comprehensive study of continental magmas and xenoliths should result in the development of increasingly realistic crustal models (e.g., Smithson and Brown, 1977).

The following is a review of some approaches toward models of the lower crust using magmas and xenoliths. It will become apparent that there are more questions than answers in much of this work. The last section will formulate some fundamental questions and suggest ways to get answers to them.

CONCEPTUAL FRAMEWORK

Questions about the lower crust must be asked within a spatial and temporal framework. Foremost is the basic observation that continental fission, fusion, and strike-slip motion implicit in the plate-tectonic development of the earth's surface have resulted in continents composed of crustal fragments of diverse ages and histories, separated by cryptic suture zones. Because of these processes, later events may be superimposed on juxtaposed crustal fragments having quite different early histories. A common later event may be intrusion and attendant heating of the crust by magmas in continental rift zones or convergent plate margins [e.g., Andean arc magmas intruding Precambrian crust in Peru (Dalmayrac *et al.,* 1977)]. In addition, we recognize that at present, new continental crust is being formed largely by magmatic activity along arcuate zones at convergent plate margins and within broad intraplate tensional zones, including rift zones. We expect that zero-age crust will have distinctive differences at these two localities. Lastly, the mechanisms of formation of Ar chean crust, particularly the crust under greenstone belts and anorthosites, may not be operating today and are an intrinsically more difficult problem (Hargraves, 1976).

CONTINENTAL MAGMATIC STUDIES

The study of melts derived from the crust is an indirect way of studying the crust itself. Some magmas that rise to shallow crustal levels have either formed in the crust or have been extensively modified by addition of a "granitic" crustal low melting fraction. The magmas occur in two general regions—convergent plate margins and intraplate regions with high thermal gradients. In both cases, intrusion of mantle-derived basaltic magmas into the crust has increased thermal gradients to crustal melting conditions as indicated by surface heat flow (Lachenbruch and Sass, 1977). Figure 12.1 shows the distribution of high-heat-flow areas in the United States. Figure 12.2 shows the relationship between thermal gradients and melting curves of crustal rocks.

A good way to identify crustal-derived components in magmas is to match chemical and isotopic signatures in the magmas with those of known sedimentary rock types or basement age provinces (e.g., Zartman, 1974). The work of Armstrong *et al.* (1977) shows that high $^{87}Sr/^{86}Sr$ ratios in young igneous rocks in the western United States are restricted to regions underlain (or inferred to be underlain) by Precambrian basement (Figure 12.3). Similarly, Sr and Pb isotope tracers have been used by Lipman *et al.* (1978) in a comprehensive study of volcanic rocks from a small region in southern Colorado. They recognize temporal as well as spatial differences in isotopic composition and find that both upper and lower crustal contamination can be identified. The implication of these data is that basement provinces can be outlined by isotopic analyses of recent shallow magmatic rocks that have interacted with the deep crust in an intraplate environment. As a further illustration, Turi and Taylor (1976) and Taylor and Turi (1976) found a correlation of high $^{87}Sr/^{86}Sr$ and ^{18}O values in Tuscan (Italy) silicic magmas (Figure 12.4) that demonstrates a major contamination by crustal melting. The effect continues into the Roman province to the south, where mixing of the silicic magmas with basic mantle-derived magmas apparently has occurred. Chemical trends in aluminous *S-type* (sedimentary-type) granitoids analogous to Turi and Taylor's (1976) Tuscan magmas have been used by White and Chappell (1976) to argue for a sedimentary crustal source, which they contrast to igneous-type, or *I-type,* granitoids that have an igneous (basaltic) crustal source.

Probably not all isotopic variations in continental magmas can be attributed to crustal contamination. Brooks *et al.* (1976) and Whitford *et al.* (1977) have identified isotopic variations in magmas that reflect subcontinental lithospheric mantle and subducted oceanic sediments. However, evidence looks persuasive for interaction of magmas with deep continental crust in Italy and in the western United States, reinforcing the hypothesis that magmas can provide information on the lower crust.

CONSEQUENCES OF CRUSTAL MELTING: TWO TYPES OF LOWER CRUST

Residual Crust

An important consequence of the contamination of magmas by a crustal low melting fraction is that a refractory residue complementary to the melt exists at depth. This

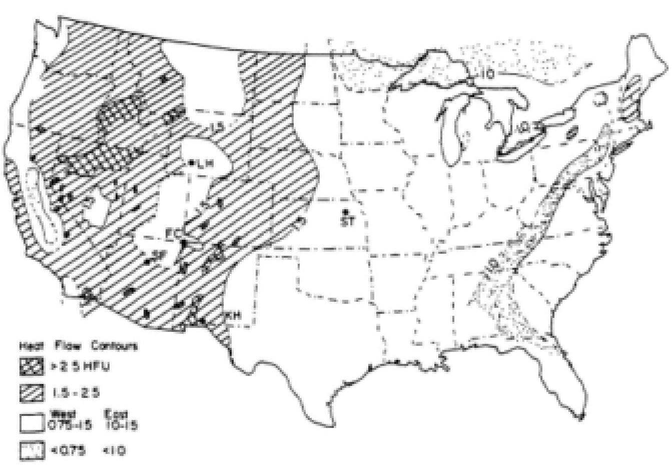


Figure 12.1
A generalized heat-flow contour map of the conterminous United States. The areas of highest heat flow coincide with the areas of Plio-Pleistocene volcanism. The Baffle Mountain high in Idaho and Nevada is especially prominent and is an interior area with high convective heat loss. Differences in heat flow in different areas probably influence the mineralogy and composition of the lower crust, creating provinces that may mimic the heat-flow map, particularly where heat flow is exceptionally high. Figure is from Lachenbruch and Sass (1977, copyrighted by the American Geophysical Union). Also shown are several xenolith localities referred to in the text: SF (San Francisco Volcanic Field), FC (Four Comers region, Colorado Plateau), LH (Leucite Hills, Wyoming), KH (Kilbourne Hole, New Mexico), ST (Stockdale, Kansas).

residue (restite) is probably at equilibrium with the melt at the depth and temperature of magma segregation with respect to mineralogy; partitioning of major, minor, and trace elements; and isotopic ratios. For example, restites in equilibrium with granitoids should be aluminous granulites in the case of pelitic staring material (Green, 1976) or amphibolite to pyroxene granulite in the case of basaltic starting material (Wyllie *et al.,* 1976). White and Chappell (1976) believe that many of these restites are now found as mafic clots and inclusions in S-type and I-type granitoids, but the restite mineralogy does not generally indicate a high-pressure origin because of re-equilibration at low pressure.

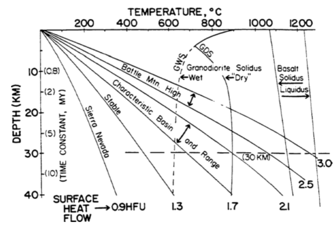
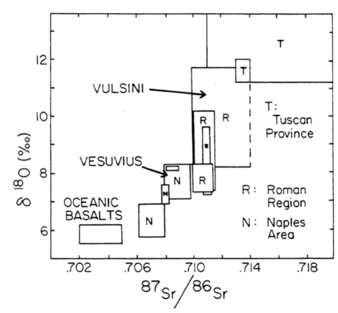


Figure 12.2 Generalized conductive temperature profiles (from left to right) for the Sierra Nevada crust, a stable reference crust, the characteristic range for the Basin and Range, and the lower limiting and typical conditions in the Battle Mountain high. All curves are drawn for a surface heat production (A_0) of 5 HGU (heat generation units) and thermal conductivity (K) of 6 mcal cm^{-1} see^{-1} $^{o}C^{-1}$. Corresponding surface heat flow is shown at the bottom of each curve. Melting relations (Wyllie, 1971) are shown for intermediate crustal rock by the curves GWS (granodiorite water saturated solidus, dashed line), and GDS (granodiorite "dry" solidus, heavy line), and for basalt by the dry basalt solidus, and dry basalt liquidus. Figure and caption modified from Lachenbruch and Sass (1977).



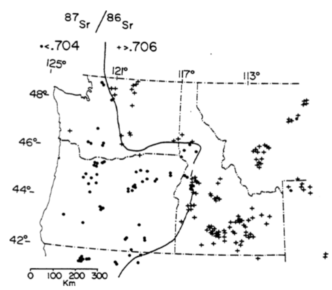


Figure 12.3 Map of the northwestern United States showing locations of Mesozoic and Cenozoic igneous rocks with $^{87}Sr/^{86}Sr$ ratios of less than 0.704 (circles) and greater than 0.706 (crosses). The heavy line is the approximate location of the 0.704 contour, and is drawn to join its trace in California defined by Kistler and Peterman (1973). References for data points are given in Armstrong et al. (1977). The strontium isotope differences may be controlled by the age and nature of the lower crust. (Figure is from Armstrong et al., 1977, reprinted from Bulletin of the Geological Society of America, with permission.)

Figure 12.4 Plot of the generalized range of whole rock $\delta^{18}0$ versus $^{87}Sr/^{86}Sr$ for various volcanic rocks in the Tuscan and Roman provinces of Italy. The correlation illustrates major contamination by crustal melting. Figure is from Turi and Taylor (1976).

Several studies have used a residual crust concept to predict the composition of the lower crust. Models of Zielinski and Lipman (1976), Taylor (1977), DeLong (1974), and Arth and Hanson (1975) illustrate the usefulness of lithophile elements, particularly rare-earth elements, in modeling mineral melt equilibria. Three points might be made in connection with these models:

1. Melts derived from continental crust are diverse, and different magmas leave different residual crusts. Thus, the residual crust concept is easily reconciled with lower crustal heterogeneity.
2. Predictive characteristics of the residual crust may have shortcomings because the trace-element partition coefficients that are the key to the calculations are not well known. This is particularly true for potentially important accessory phases (apatite, scapolite, and zircon)

and for the heat-producing elements U, Th, and K. Acidic compositions in the lower crust are difficult to model because accessory minerals may incorporate large proportions of the total trace-element content of the rock (Gromet and Silver, 1978). A further problem concerns whether minerals and melt are at equilibrium.

3. The above residual-crust models assumed that material was transferred by a melt. Alternative methods of material transport, possibly involving a fluid phase, must also be considered. For example, the role of aqueous fluid phase in upward concentration of U, Th, and K in the crust (Tarney, 1976) will have to be evaluated quantitatively. Diffusive transport (Orville, 1963) or advective transport in a hydrous fluid are alternative mechanisms. Aqueous fluids are also important in the problem of phase transportation kinetics (Ahrens and Schubert, 1975).

Solidified Basic Melts

A second consequence of generation of abundant silicic crustal melts is that a large part of the heat of fusion must come from intrusion and solidification of higher-temperature mantle-derived basic melts. Large regions of lower crust, equivalent to the volume of crustal-derived silicic magma, must be crystallized basic intrusions. This gives us a rationale for another class of lower crustal rocks that are represented in xenolith population—cumulate rocks of basic composition.

EXPOSED LOWER CRUSTAL SECTIONS

Possible lower crustal sections, now exposed at the surface by tectonic activity, that have been investigated by geophysical techniques indicate what we may expect in some regions of the lower crust. The most cited example of an exposed lower crustal section is the Ivrea zone in northern Italy (Berckhemer, 1969), where both restites of sedimentary parentage and metamorphosed equivalents of basaltic magmas have been found (Mehnert, 1975; Schmid and Wood, 1976). The lower crust in this region appears to have originally been composed of a sedimentary sequence of sandstone, shales, and limestones with mafic lavas intruded into the section particularly at the base. The grade of metamorphism varies from the upper amphibolite to the granulite facies (Schmid and Wood, 1976).

XENOLITH STUDIES

In the absence of drill holes to the relevant depths and because of questions in the interpretation of surface granulite facies rocks in crustal sections of normal thickness, xenolith data must be regarded as a primary source of information about the lower crust. The anticipated scale of lateral inhomogeneities of the lower crust far exceeds the spacing of xenolith localities. Because of the possibility that xenolith localities may represent unique or unusual conditions in the crust, a better approach might be to group the localities based on tectonic similarities: convergent plate margins, rift valleys, areas of intraplate volcanism, and Precambrian cratonic regions. Xenoliths have not been reported from all tectonic regimes, e.g., continental collision zones and greenstone belts.

Prior to a discussion of specific xenolith areas, two important points should be noted:

1. Xenoliths represent the mantle and crustal section only at the time of eruption of the conveying medium (e.g., lava or kimberlite). The section is not necessarily the same as the present one, because of possible subsequent tectonic activity or textural re-equilibration.
2. Pressure-temperature (*P-T*) conditions in the crust-mantle column may or may not be reflected by *P-T* estimates based on coexisting minerals from the xenoliths. Equilibration of xenoliths from the lower crust during periods of kimberlite eruption may not occur because of slow reaction rates at low crustal temperatures. In this case, the geothermal gradient determined from xenolith studies is probably a fossil gradient. In contrast, during volcanic episodes on a regional scale (e.g., western United States), temperatures may reach 800-900ºC in the lower crust (see Figure 12.2) causing textural re-equilibration. This statement follows from the observation that diffusion over 1 cm will occur within 10 million years (m.y.) for diffusion coefficients greater than 3×10^{-15} cm^2/sec (which occur in the range 800-900ºC in silicates, see Hofmann and Hart, 1978; Buckley, 1973). Textural equilibration will probably occur in a shorter time, perhaps 10^5 years, if a melt phase or a fluid phase is present, because diffusion coefficients are several orders of magnitude higher in liquids than in solids (Aherns and Schubert, 1975). However, intergranular transport may not always be the rate-controlling process (Loomis, 1976) because the rate of divariant reactions can depend on diffusion rates within reactants even if intergranular transport is fast.

Xenoliths Associated with Convergent Plate Boundaries

In the Japanese arc, where the tectonic setting is well known, there are two contrasting xenolith localities in young volcanic rocks. One (Ichinomegata) has a lower equilibration temperature and is more hydrous than the other (Oki-Dogo).

1. Oki-Dogo alkali basalt has abundant xenoliths ranging from granite (underlying crust) to peridotite. Takahashi (1978) has constructed a crustal column based on mineralogically derived *P-T* estimates (Figure 12.5). An important observation is that the transition from lower crust to upper mantle is within a series of layered cumulate rocks ranging from gabbros to peridotites. The xenoliths display no deformation effects, yet they are not sim

ply crustal cumulates of the Oki-Dogo magma itself; their Ar release pattern indicates ages of perhaps 30 m.y. to 40 m.y. and their $^{87}Sr/^{86}Sr$ ratios are generally higher than the host lava. Other volcanic rocks of southwest Honshu have similar assemblages. The geothermal gradients derived from the xenoliths for the crust resemble those of the Battle Mountain high of Figure 12.1.

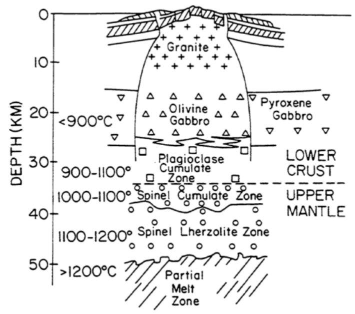


Figure 12.5
Petrological model of Oki-Dogo Island, Japan, based on xenoliths found in alkali basalt flows. Figure is from Takahashi (1978).

2. Ichinomegata andesite also has abundant xenoliths ranging from granite to peridotite. However, equilibration temperatures of the xenoliths are consistently lower than those of the Oki-Dogo xenoliths for equivalent pressures. The mafic and ultramafic assemblages contain hornblende and are deformed (sheared) rather than cumulate in texture. The regional extent of Ichinomegata-type lower crust is indicated by similar xenolith assemblages found in high-alumina basalts from the Honshu and Kurile arcs (Katsui *et al.,* 1978; Shimazu *et al.,* 1978).

In summary for Japan, the lower crustal xenolith suite includes abundant types of basaltic composition and appears to lack metasedimentary and refractory types. The xenolith suite indicates that crystallized basic melts, modified by crystal settling and shearing, form extensive regions of the lower crust in Japan. Considering the importance of island arcs in many models of lower crustal development (see Chapter 3 as an example), the Japanese data seem especially important.

Xenoliths Associated with Rift Valley Environments and Intraplate Volcanic Areas

Rio Grande Rift

Two xenolith localities (Kilbourne Hole and the Elephant Butte Reservoir area) are young volcanic maars and flows associated with the Rio Grande rift in New Mexico. Both metasedimentary and meta-igneous rocks of granulite metamorphic grade have been recognized among the abundant xenoliths at Kilbourne Hole (Padovani and Carter, 1977). Metasedimentary rock types include sillimanite-bearing garnet granulites, which have been interpreted as high-temperature residues of pelitic sediments. Meta-igneous rock types include two-pyroxene granulites that are compositionally basalts and are perhaps solidified basaltic magmas. Anorthosites and charnockites also occur in the meta-igneous suite and have led Padovani and Carter (1977) to suggest that the assemblages in the lower crust at Kilbourne Hole are similar to some high-grade metamorphic terranes exposed at the earth's surface. Two-pyroxene granulities of basaltic composition and charnockites have also been recognized in the Elephant Butte Reservoir area farther north by A. M. Kudo (University of New Mexico, personal communication, 1978).

The principal techniques used to study these xenolith suites include petrographic analyses and electron-microprobe analyses of coexisting minerals. Such studies are indispensable for any quantitative evaluation of pressure and temperature. A major result of the Kilbourne Hole study was that the geothermal gradient of 30ºC/km calculated from mineral *P-T* conditions (Figure 12.6) is roughly equivalent to the inferred gradient for the Battle

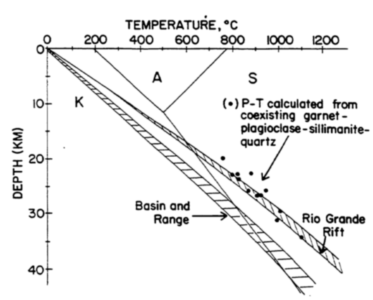


Figure 12.6
Temperature-depth plot showing distribution of values (plotted symbols) determined by Padovani and Carter (1977) from coexisting garnet and plagioclase in Kilbourne Hole garnet granulites. The aluminum silicate stability field from Holdaway (1971) is plotted for reference (K: Kyanite; A: Andalusite; S: Sillimanite). The vertically lined field represents the range of temperature with depth for surface heat fluxes of 2.4 HFU (heat flow units) in the southern Rio Grande rift (Decker and Smithson, 1975). The horizontally lined field represents the range of temperature with depth for surface heat fluxes of 2.0 HFU for the Basin and Range (Decker and Smithson, 1975). Figure is modified from Padovani and Carter (1977).

Mountain high (Figure 12.2) and the Rio Grande rift (Cook *et al.,* 1978; Reiter *et al.,* 1978).

Midcontinent Rift

The kimberlite at Stockdale, Kansas (Meyer and Brookins, 1976) is located on the midcontinent gravity high and yields only a few granulite and pyroxenite xenoliths of probable lower crustal or upper mantle origin. Most are altered, but a few have fresh cores and contain minor sillimanite and sapphirine. One chemically analyzed sample has the whole-rock composition of an olivine tholeiite and could represent magma crystallized at depth associated with rifting 1100 m.y. ago. No xenoliths of possible sedimentary origin have been reported. Granites apparently overlie the gabbros and other igneous-derived rocks of the lower ciust. If the granites are a low-temperature crustal melt fraction, the residues in the lower crust have not been found or apparently are not present in the lower crust beneath Stockdale.

Dating of the xenoliths by Rb/Sr mineral isochron techniques (Brookins and Woods, 1970) suggests a complex thermal history for the region. The postulated rift environment suggests very high heat flow in the Precambrian. The question of the timing of the alteration becomes significant. Has the lower crust in this region undergone retrograde metamorphism, and does this retrogression correspond to present heat flow values? Or was alteration associated with the later intrusion of the kimberlite, and has a fossil high-temperature geotherm been recorded by the xenoliths? Brookins and Meyer (1974) believe that the alteration accompanied kimberlite intrusion.

Massif Central, France

Xenoliths from Bournac, a Plio-Pleistocene volcanic pipe in the Massif Central of southern France, were studied by Leyreloup *et al.* (1977) and Bilal (1976). Numerous fragments of granulite facies metamorphic rocks derived from both sedimentary and igneous parents have been found. The igneous suite is made up principally of basic granulites but includes rocks with compositions ranging from ultrabasic to acidic. Leyreloup *et al.* (1977) interpret the meta-igneous suite as low-pressure crystal fractionares of a tholeiitic magma that recrystallized in the granulite metamorphic facies. Granulites of sedimentary composition include metamorphosed shales, graywackes, and arkoses. Sparse calcareous granulites have been found in other xenolith localities in the Massif Central. Intercalation of the igneous and sedimentary components is present in composite hand specimens, where it is presumed that the igneous component intruded the sedimentary component. Age dating by whole-rock Rb-Sr (Hamet *et al.,* 1978) indicates an age of 1300 m.y. for the xenoliths; zircon dating indicates heating events at 300 m.y. and 600 m.y. Outcrops of old granulitic terrane in the nearby Variscan (late Paleozoic) metamorphic rocks may be equivalent in age (Hamet and Allègre, 1976), implying that the polymetamorphism is also recorded in surface exposures in this region. Much remains to be learned of the thermal history of the deep crust from surface exposures.

Leyreloup *et al.* (1977) conclude from extrapolation of Bournac xenolith abundances and types that the regional composition of the lower crust is andesitic and that the metasedimentary rocks are not depleted in large-ion lithophile elements as in the classic depleted granulite terrane of Heier (1973).

Intraplate Volcanic Areas

Other volcanic areas may also have a lower crust similar to that in rift valleys. For example, an assemblage of coexisting charnockites and two-pyroxene granulites similar to the Kilbourne Hole suite has been recognized by Stoesser (1973) in the recently active San Francisco volcanic field in Arizona. Stoesser (1973) interprets these xenoliths as wall rock into which an ultramafic layered sequence associated with the volcanic activity has been intruded. This suggests that one component of the lower crust in volcanic areas that include rift valleys would be layered mafic intrusions formed by cumulate processes. Francis (1976) reaches the same conclusion for granulites from Nunivak, Alaska, as have Kay *et al.* (1978) for the Leucite Hills, Wyoming.

An intraplate xenolith locality that has received considerable attention is the Australian Delegate pipe (Lovering and White, 1968; Irving, 1974), where possible lower crustal xenoliths include a charnockite and various granulites of basaltic composition. Similar xenolith types have been found in alkali basalt flows in an intraplate area to the north in Queensland (Stephenson and Griffin, 1976). A summary of xenolith types in Australian localities has been compiled by Wass and Irving (1976).

The lower crust in volcanic regions, particularly in rift valleys, appears to be in the granulite metamorphic facies and appears to have two-pyroxene granulites of basaltic composition as a component. These basaltic composition rocks may represent the metamorphic equivalent of basic intrusions responsible for crustal dilation in rift valleys and uplift of the flanking crust. Low-melting granitic components present in the crust prior to intrusion may rise to upper crustal levels as contaminants of basaltic lavas or as silicic intrusives or extrusives. Other assemblages, such as residual sillimanite-garnet granulites and anorthosites may be present, but if so, they are probably related to the history of the crust prior to the high-heat-flow episode.

Xenoliths from Polymetamorphosed Precambrian Lower Crust

Two of the most highly studied suites of probable lower crustal xenoliths come from kimberlites in regions where polymetamorphism of Precambrian lower crust can be demonstrated. Similar complex histories for other lower crustal suites may be developed with further investigation.

Southern Africa

Kimberlites with abundant mantle and crustal xenoliths occur within the Precambrian Kaapvaal Craton and its surrounding younger mobile belt in Lesotho. Regional difference in the lower crust may be indicated by contrasting xenolith populations in the kimberlites (Griffin *et al.,* 1979). High-temperature anhydrous granulites are confined to the mobile belts (Figure 12.7), while the intercratonic crustal xenoliths are typically lower-temperature (amphibolite facies) gneisses that may or may not represent deep crustal material. The mobile belts are interpreted by Kröner (1977) to have been formed by reworking of older rocks in the Precambrian with little addition of new material. Work has concentrated on granulite xenoliths from the mobile belt (Griffin *et al.,* 1979; Rogers, 1977), while xenoliths from the craton have received little attention.

At one locality in the mobile belt, 50-70 percent of the meta-igneous granulite xenoliths are basaltic in composition and 30-50 percent are intermediate to acidic in composition (Bloomer and Nixon, 1973). Whole-rock chemical trends (i.e., Mg/Fe increases with SiO_2) do not resemble those resulting from crystal fractionation. Variable concentrations of incompatible elements (such as K and Rb) may reflect modification by high-grade metamorphic processes, perhaps including partial melting (Griffin *et al.,* 1979; Rogers, 1977). The rare-earth analyses for mafic garnet granulites indicate that amphibole-bearing garnet-free mineral assemblages were present during the solid-liquid fractionation. The rocks are at present garnet bearing.

The Lesotho lower crust appears to have undergone a multistage heating history—zircon ages are Proterozoic, but the last heating episode probably coincided with the widespread Karoo volcanism in Cretaceous time. Geothermal gradients were probably fairly high at the time of kimberlite intrusion as the time period of their intrusion is similar to that of the Karoo volcanism. These xenoliths may have resided in abnormally warm lower crust, although probably not so hot as the Rio Grande rift.

Colorado Plateau

Tertiary kimberlite diatremes in the Four Corners region of the Colorado Plateau contain abundant xenoliths and xenocrysts from both the crust and upper mantle. McGetchin and Silver (1972) examined xenoliths from the Mule Ear diatreme and based a lower crustal and mantle model of the Colorado Plateau on the shape, relative size, and abundance of the crystalline fragments. The lower crust in the model consists of amphibolite and granulite facies metamorphic rocks, all of which have been hydrated during retrograde metamorphism in the amphibole facies. The most common lower crustal rock type is meta-gabbroic granulite gneiss containing garnet. Metasedimentary gneisses of upper amphibolite grade (E. Padovani, Massachusetts Institute of Technology, personal communication, 1978), eclogite, chlorite schist, serpentine schist, and pyroxenite also occur.

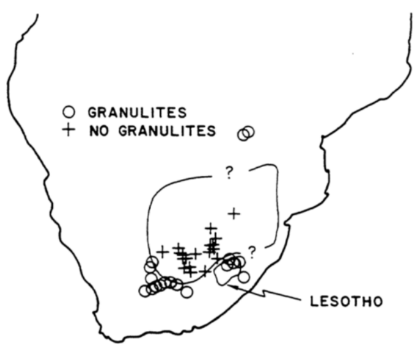


Figure 12.7
Distribution of granulites in crustal xenoliths from South African kimberlites. The crosses represent pipes where no granulites have been found; all lie within the Kaapvaal Craton (area surrounded by the solid line). The open circles represent pipes in the surrounding mobile belt where granulites have been found. Figure modified from Griffin et al. (1979).

The ubiquitous retrograde metamorphism, even in mantle xenoliths (Smith and Levy, 1976), stands in sharp contrast to xenoliths discussed from other localities. Geothermal gradients implied by the mineral assemblages are low, approximating the stable reference crust geotherm of Figure 12.2. Helmstaedt and Schulze (1977) advocate an even lower geothermal gradient, on the basis of a low-*T*, high-*P* chlorite-eclogite retrograde mineralogy of an originally high-*T* eclogite. They argue that subcontinental hydration accompanied retrograde metamorphism, which was caused by underthrusting of hydrated oceanic lithosphere. In contrast, Smith (1977) has suggested that hydration is localized along zones of lithospheric weakness that coincide with the monoclines along which the kimberlites occur.

Xenoliths from South Africa and the Colorado Plateau are compositionally diverse and have complex thermal and deformational histories. Yet several generalizations can be made. First, metasedimentary rocks are subordinate to meta-igneous rocks. Second, fossil metamorphic gradients seem to be represented in Africa, since regional heat-flow values are low and relatively uniform across the mobile belt-craton boundary. Third, despite the retrograde metamorphic and hydration effects found in the

Colorado Plateau xenoliths, the rock types are not significantly different from those found in other xenolith localities of the western United States. This finding implies that differences in the lower crust may be due to regional variations in water content and metamorphic grade as well as to heterogeneity in chemical composition.

OUTSTANDING QUESTIONS AND FUTURE DIRECTIONS

Research on continental magmas and xenoliths has a potentially unique role in crustal studies. Several studies have resulted in important observations, but the most basic questions about the lower crust remain unanswered. Continental crustal studies are probably at a stage of development equivalent to that reached in the study of oceanic crust 20 years ago. It may seem presumptuous to ask questions about a region so complex as the continental crust, but it seemed equally presumptuous then to ask equivalent questions about the oceanic crust. We will ask three questions about the crust and outline what the role of xenolith and magma studies might be in formulating answers.

What is the Origin of the Rock Suites at Sites of Continental Crustal Formation?

What is the origin of rocks that make up new crust? Are they sedimentary, or are they derived from sedimentary rocks by melting or fluid-phase transport (metasomatism), or are they igneous? Are they hydrous or anhydrous? There are a number of chemical and isotopic criteria that have a wide applicability in answering these questions, particularly the use of trace elements (lithophile elements, transition metals) and radiogenic isotopes (Sr, Pb, Nd).

Interpretation of xenoliths and magmas will benefit from comparison with a large number of existing analyses on a wide range of igneous and sedimentary rock types. Some types of analyses that have not been widely done, in particular, rare gas, volatile element, and oxygen isotope analyses, may also be valuable. For example, Phinney *et al.* (1978) found positive ^{129}Xe anomalies in CO_2-rich well gas, indicating that the gas was not atmospheric. In addition, trapped gases may prove to be good monitors of diffusive material transfer. The possible use of oxygen isotope analyses may be demonstrated by the unexpected and only partially understood variability found in eclogites (Vogel and Garlick, 1970). Shieh and Schwartz (1974) have shown that oxygen isotope analyses of exposed granulite facies rocks are similar to basalts. The fractionation of volatile elements (such as hydrogen, the halogens, and sulfur) in deep crustal xenoliths is important in view of the role of volatiles in melting, diffusive transport, electrical conductivity, and other properties of the lower crust (Goldsmith, 1976; Necut *et al.,* 1977).

The interpretation of trace-element and isotope data will remain uncertain until basic questions about the partition of trace elements between granitic-dioritic melt and residual crystals and the kinetics of equilibration are resolved. The application of radiogenic isotopic criteria (Sr, Nd, Pb) for characterization of melts from crustal reservoirs is likewise hampered by lack of knowledge of equilibration times and trace-element distributions (Ben Othman *et al.,* 1978). Studies of both experimental and natural mineral assemblages are necessary to resolve these questions.

The preliminary conclusion drawn from several xenolith studies is that many deep crustal rocks are igneous in origin and basaltic to andesitic in bulk chemistry. However, sampling bias may exist because studied suites are from volcanic and kimberlite areas, where the lithosphere may be structurally and petrologically abnormal. A second possible bias is the ability of some xenolith types to survive transportation better than others. Additional bias results in the collection of xenoliths, in which small or altered types may be underrepresented. Field studies must take into account these biases. Improved representation of tectonic environments should be a major focus of additional sampling (e.g., greenstone belts and adjacent gneiss terranes of the Precambrian and continental collision zones).

Field studies on xenoliths and magmas could also benefit greatly from geophysical studies in the same region. Thus it seems desirable to search for localities also investigated using seismic reflection techniques (the Leucite Hills and the COCORP, Wind River line, Smithson *et al.,* 1978, for example). In the future, seismic reflection lines might be planned across regions with interesting contrasts in xenolith populations (e.g., the Kaapvaal shield and mobile belt or the Four Comers region).

Field work within areas of exposed probable lower crustal rocks (e.g., the Ivrea zone, Berckhemer, 1969) is a valuable guide to the relationship of rock types and structures in xenolith suites. However, some types of lower crust that are present in xenolith suites may not be represented in surface exposures.

What are the Age Relationships Within Vertical as Well as Horizontal Crustal Sections?

At present there is a lack of knowledge of the mechanisms of continental crustal thickening. It is possible that under-plating and horizontal underthrusting continually add young crust under continents, while volcanism and shallow plutonism add material to the upper crust. The dating of polymetamorphic rocks (and examination of initial isotope ratios) may be the only way to resolve these questions. There are many ages to be sorted out, i.e., thermal events (including heating by igneous intrusions), metasomatic events, and melting events. Rb-Sr, K-Ar, U-Pb, and Nd-Sm dating methods will be useful.

The relatively new dating method, Nd-Sm, will have a broad application in the granulite facies rocks, especially those that are low in K and Rb. The fractionation of Nd and Sm between garnet (low Nd/Sm) and pyroxene (high Nd/Sm) make them ideal for dating crystallization ages of eclogites. The solubility. of rare-earth elements in metasomatic fluids is probably less than the solubility of K and Rb, so that the Nd-Sm isotopic system may be relatively immune to metasomatic events.

What is the Temperature-Strain History of Lower Crustal Regions?

Prerequisite to any statements about the temperature history of the lower crust inferred from xenolith minerals is the *P-T* calibration of mineral assemblages and coexisting mineral compositions found in xenoliths (Whitney and Stormer, 1977). Laboratory experiments and theoretical advances are needed to clarify the calibration problems. Some rocks yield satisfactory *P-T* estimates. For other assemblages, especially high-variance ones, estimates are not very good, mainly because of uncertainties of distribution coefficients with *P, T,* and composition. Eclogites (garnet-clinopyroxene rocks) are a good example. Griffin *et al.* (1979) noted a gap in Lesotho xenolith depth estimates between a lower crustal group (depths to 35 km) and a high-pressure mantle group (depths greater than 85 km). They suspect that xenoliths in the depth range of 25-80 km are present and may include eclogites.

Intercrystalline equilibrium, involving the distribution of elements between sites in a crystal lattice, is a potentially powerful method for determining *P* and *T.* Crystallographic methods are required to determine site occupancies, but distinctive solid-solution behavior, calculated from chemical analyses, is a reflection of the site behavior. For example, Figure 12.8 is a plot of two solid-solution components (jadeite and Tschermak's molecule) in clinopyroxenes from granulite xenoliths. The amount of aluminum in fourfold coordination is thought to increase with temperature, while the amount of aluminum in sixfold coordination is thought to increase with pressure. Aluminum in fourfold coordination is calculated by the amount of Tschermak's molecule, while the amount of aluminum in sixfold coordination coupled with sodium is calculated by the amount of jadeite (White, 1964). The grouping of analyses from one locality is distinctive in Figure 12.8, but interpretation hinges on further *P* and *T* calibration. Griffin *et al.* (1979) attribute the high jadeite component in the Lesotho granulites to pressures in the lower crust. Since similar jadeite components are not found in surface granulites, they postulate that aluminum has re-equilibrated upon uplift if surface granulite outcrops were formerly in the deep crust.

Determination of a lower crustal geothermal gradient has been a major concern of several xenolith studies. Some granulites yield temperatures that are compatible with present heat flow, and the geothermal gradients derived from these xenoliths may reflect the present gradients (see Padovani and Carter, 1977). Dating of texturally defined mineral assemblages would be a valuable constraint on the timing of the *P-T* conditions recorded by the mineral assemblage and the relation to present heat flow.

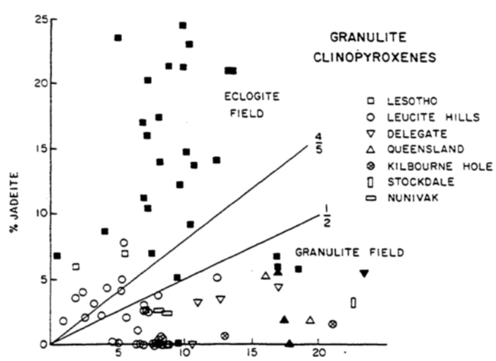


Figure 12.8
Plot of mole percent jadeite against Tschermak's molecule for granulite clinopyroxenes from xenoliths of probable lower crustal origin. Solid symbols represent garnet-bearing assemblages; open symbols represent garnet-free assemblages. When Fe_2O_3 was not reported in the analyses, it was calculated using the method of Papike et al. (1974). Jadeite and Tschermak contents were calculated using the method of Yoder and Tilley (1962) and White (1964), Lesotho analyses from Griffin et al. (1979), Delegate analyses from Lovering and White (1968) and Irving (1974), Kilbourne Hole analyses from Padovani and Carter (1977), Stockdale analyses from Meyer and Brookins (1976), and Leucite Hills and Queensland analyses by authors.

In several xenolith studies, retrograde metamorphic effects have been noted, indicating partial re-equilibration of lower crustal mineral assemblages in response to changing *P-T* conditions. A major question is: How fast do the assemblages respond to changes in geothermal gradient? At the earth's surface, high-grade metamorphic rocks react too slowly to equilibrate. Reactions at mantle (asthenospheric) depths should be complete within a hand specimen. The depth to the zone of equilibration is undoubtedly variable. In some areas this zone may reach into the mantle (Frazer and Lawless, 1978; Herzberg and Chapman, 1976) but in areas of crustal melting and a high geothermal gradient, the partially re-equilibrated zone should be in the lower crust. Mineral zoning and corona textures of rock in the zone may be used to determine temperature and time of reaction (e.g., Loomis, 1976). In some cases (e.g., Colorado Plateau) partly equilibrated xenoliths indicate both a high paleogeothermal gradient and a low present geothermal gradient, corresponding to a hydrated retrograde mineral assemblage. Exsolved phases in plagioclases and pyroxenes from deep crustal xenoliths may also yield important information on the kinetics of mineral equilibration and residence time of the xenoliths in the lower crust.

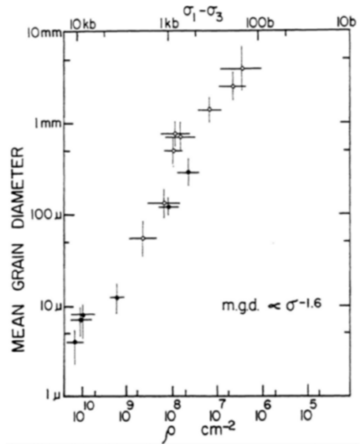


Figure 12.9
Mean grain size versus dislocation density and differential stress for oilvine. Analogous relations can be defined for plagioclase and clinopyroxene and should be useful in defining differential stress for deformed granulites. Figure is from Koldstedt et al. (1976).

An important question that follows from study of the retrograde hydration in some xenoliths is the source of the water. Stable-isotope studies (hydrogen and oxygen) may be able to distinguish alternatives such as hydrothermal circulation or dehydration of underlying subducted oceanic lithosphere. Microcrack studies may also be important in defining mechanisms for migration of fluid phases in the deep crust (Simmons and Richter, 1978).

Substantial progress on the dynamics of upper-mantle flow has been made by studies of deformation mechanisms of olivine in the laboratory and application to peridotite xenoliths (e.g., Kohlstedt *et al.,* 1976). A parallel study using lower crustal granulites deserves attention, The density of unannealed defects can be examined using the transmission electron microscope. Experimental work needs to be done to understand the deformation mechanisms of feldspar and pyroxene, probably the most abundant lower crustal minerals. In addition, the simple relationship between grain size and stress found in olivine (Figure 12.9) encourages the search for a similar relationship for feldspars and pyroxenes. One observes that grain size in basic granulites is variable from locality to locality; could this reflect a variable stress in the lower crust? If stress and temperature can be determined for a xenolith, then strain rate can be calculated, and questions of lower crustal dynamics can be addressed.

SUMMARY AND CONCLUSIONS

Xenoliths and magmas can help to answer such questions as (a) what is the parentage of the lower crust, i.e., percentage of original igneous versus sedimentary material, percentage derived directly from the mantle versus material derived from crustal processes; (b) is the lower crust hydrous or anhydrous; (c) what are the temperature-pressure regimes in the lower crust; (d) are rocks in the lower crust at equilibrium with present or past temperature-pressure conditions; (e) what are the age relations between various units of the lower crust; and (f) what is the deformation history and what is the present state of stress in the lower crust? When knowledge derived from xenoliths and magmas is combined with drill holes and geophysical information, a three-dimensional picture of the crust can be constructed. Crustal studies have barely begun: a great expansion of our ability to answer all these questions can be expected in the near future.

ACKNOWLEDGMENTS

We thank D. Smith and E. Padovani for reviews and members of the Continental Tectonics Panel for discussion and acknowledge the National Science Foundation for providing financial support under Grant EAR 77-13656.

References

Ahrens, T., and G. Schubert (1975). Gabbro-eclogite reaction rate and its geophysical significance, *Rev. Geophys. Space Phys. 13,* 383-400.
Armstrong, R. L., W. Taubeneck, and P. Hales (1977). Rb-Sr and K-At geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington and Idaho, *Geol. Soc. Am. Bull. 88,* 397-411.
Arth, J. G., and G. N. Hanson (1975). Geochemistry and origin of the early Precambrian crust of northeastern Minnesota, *Geochim. Cosmochim. Acta 39,* 325-362.
Ben Othman, D., A. Juery, and C. I-Allègre (1978). Limitations on granite formation inferred by Nd, Sr and Pb isotope systematics (abstr.), *Eos Trans. Am. Geophys. Union 59,* 392.
Berckhemer, H. (1969). Direct evidence for the composition of the lower crust and the Moho, *Tectonophysics 8,* 97-105.
Bilal, A. (1976). Les fluides carboniques dans les enclaves char-nockitiques de Bournac (Massif Central français). Implications sur les structures de la croute continentale, Thèse 3e cycle, Nancy.
Bloomer, A. G., and P. J. Nixon (1973). The geology of the Letseng-la-terae kimberlite pipes, in *Lesotho Kimberlites,* P. H. Nixon, ed., Lesotho National Development Corporation, Masern, Lesotho, pp. 20-32.

Brookins, D. G., and H. O. A. Meyer (1974). Crustal and upper mantle stratigraphy beneath eastern Kansas, *Geophys. Res. Lett. 1,* 269-272.

Brookins, D. G., and M. J. Woods (1970). Rb-Sr geochronologic investigation of basic and ultrabasic xenoliths from the Stock-dale kimberlite, Riley County, Kansas, *Kansas Geol. Surv. Bull. 199,* 3-12.

Brooks, C., D. E. James, and S. R. Hart (1976). Ancient lithosphere: its role in young continental volcanism, *Science 193,* 1086-1094.

Buckley, G. (1973). The effect of diffusion on garnet zoning, Ph.D. thesis, U. of Illinois, 112 pp.

Cook, F. A., E. R. Decker, and S. B. Smithson (1978). Preliminary transient heat flow model of the Rio Grande rift in southern New Mexico, *Earth Planet. Sci. Lett. 40,* 316-326.

Dalmayrac, B., J. Lancelot, and A. Leyreloup (1977). Two billion year granulites in the late Precambrian metamorphic basement along the southern Peruvian coast, *Science 198,* 49-51.

Decker, E. R., and S. B. Smithson (1975). Heat flow and gravity interpretation across the Rio Grande rift in southern New Mexico and west Texas, *J. Geophys. Res. 80,* 2542-2552.

DeLong, S. E. (1974). Distribution of Rb, Sr, and Ni in igneous rocks, central and western Aleutian Islands, Alaska, *Geochim. Cosmochim. Acta 38,* 245-266.

Francis, D. M. (1976). Corona-bearing pyroxene granulite xenolithsand the lower crust beneath Nunivak Island , Alaska, *Can. Mineral. 41,* 291-298.

Frazer, D., and P. Lawless (1978). Paleogeotherms: implications of disequilibrium in garnet lherzolite xenoliths, *Nature 273,* 220-222.

Goldsmith, J. R. (1976). Scapolites, granulites, and volatiles in the lower crust, *Geol. Soc. Am. Bull. 87,* 161-168.

Green, T. H. (1976). Experimental generation of cordierite or garnet-bearing granitic liquids from a pelitic composition, *Geology 4,* 85-88.

Griffin, W. L., D. A. Carswell, and P. H. Nixon (1979). Lower crustal granulites and eclogites from Lesotho, Southern Africa, in *The Mantle Sample: Inclusions in Kimberlites and Other Volcanics,* F. R. Boyd and H. O. A. Meyer, eds., American Geophysical Union, Washington, D.C., pp. 59-86.

Gromet, L. P., and L. T. Silver (1978). Implications of rare earth distribution among minerals in a granodiorite. Peninsular Ranges Batholith, Southern California (abstr.), *Eos Trans. Am. Geophys. Union 59,* 399-400.

Hamet, J., and C. J. Allègre (1976). Hercynian orogeny in the Montagne Noire (France): application of ^{87}Rb-^{87}Sr systematics, *Geol. Soc. Am. Bull. 87,* 1429-1442.

Hamet, J., A. Vitrac, and C. J. Allègre (1978). ^{87}Rb-^{87}Sr and U-Pb systematics in granulitic xenoliths in volcanic rocks and limitations on the formation of the lower crust (abstr.), *Eos Trans. Am. Geophys. Union 59,* 392.

Hargraves, R. (1976). Precambrian geologic history, *Science 193,* 363-371.

Heier, K. S. (1973). Geochemistry of granulite facies and problems of their origin, *Phil. Trans. Roy. Soc. London S273,* 429-442.

Helmstaedt, H., and D. J. Schulze (1977). Type A-type C eclogite transition in a xenolith from the Moses Rock diatreme. Further evidence for the presence of metamorphosed ophiolites beneath the Colorado Plateau, in *Extended Abstracts of the Second International Kimberlite Conference,* 1977.

Herzberg, C., and N. Chapman (1976). Clinopyroxene geothermometry of spinel-lherzolites, *Am. Mineral. 61,* 626-637.

Hofmann, A. W., and S. R. Hart (1978). An assessment of local and regional equilibrium in the mantle, *Earth Planet. Sci. Lett. 38,* 44-62.

Holdaway, M. J. (1971). Stability. of andalusite and the aluminum silicate phase diagram, *Am. J. Sci. 271,* 97-131.

Irving, A. J. (1974). Geochemical and high pressure experimental studies of garnet pyroxenite and pyroxene granulite xenoliths from the Delegate basaltic pipes, Australia, *J. Petrol. 15.* 1-40.

Katsui, Y. K. Niida, M. Yamonoto, and S. Nemoto (1978). Genesis of calc-alkalic andesite from Oshima-Oshima and Ichinomegata volcanoes, North Japan (abstr.), in *International Geodynamics Conference "Western Pacific" and "Magma Genesis," Tokyo, March 13-17, 1978,* Inter-Union Commission on Geodynamics, and Science Council of Japan, p. 266.

Kay, S. M., R, W. Kay, J. Hangas, and-T. Shedden (1978). Crustal xenoliths from potassic lavas, Leucite Hills, Wyoming, *Geol. Soc. Am. Abstr. Programs 10,* 432.

Kistler, R. W., and Z. Peterman (1973). Variations of St, Rb, K. Na and initial ^{87}Sr/^{86}Sr in Mesozoic granitic rocks and intruded wall rocks in central California, *Geol. Soc. Am. Bull. 84,* 3489-3512.

Kohlstedt, D., C. Goetze, and W. Durham (1976). Experimental deformation of olivine single crystals with applications to flow in the mantle, in *Petrophysics: The Physics and Chemistry of Minerals and Rocks,* R. Strens, ed., John Wiley, New York, pp. 35-49.

Kröner, A. (1977). The Precambrian geotectonic evolution of Africa: plate accretion versus plate destruction, *Precambrian Res. 4,* 163-213.

Lachenbruch, A., and J. Sass (1977). Heat flow in the United States and the thermal regime of the crust, in *The Earth's Crust,* J. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 626-675.

Leyreloup, A., C. Dupuy, and R. Andriambololona (1977). Catazonal xenoliths in French Neogene volcanic rocks: constitution of the lower crust, *Contrib. Mineral. Petrol. 62,* 283-300.

Lipman, P., B. Doe, C. Hedge, and T. Steven (1978). Petrologic evolution of the San Juan volcanic field, southwestern Colorado: Pb and Sr isotope evidence, *Geol. Soc. Am. Bull. 89,* 59-82.

Loomis, T. P. (1976). Kinetics of a garnet granulite reaction, *Contrib.Mineral. Petrol. 62,* 1-22.

Lovering, J. F., and A. J. R. White (1968). Granulitic and eclogitic inclusions from basic pipes at Delegate, Australia, *Contrib. Mineral. Petrol. 21,* 9-52.

McGetchin, T. R., and L. T. Silver (1972). A crustal-upper mantle model for the Colorado Plateau based on observations of crystalline rock fragments in the Moses Rock Dike, *J. Geophys. Res. 77,* 7022-7037.

Mehnert, K. R. (1975). The Irvea Zone—a model of the deep crust, *Neues Jahrb. Mineral. Abh. 125,* part 2, 156-199.

Meyer, H. O. A., and D. G. Brookins (1976). Sapphirine, sillimanite and garnet in granulite xenoliths from Stockdale kimberlite, Kansas, *Am. Mineral. 61,* 1194-1202.

Necut, A., J. Connerney, and A. F. Kuckes (1977). Deep crustal electrical conductivity: evidence of water in the lower crust, *Geophys. Res. Lett. 4.* 239-242.

Orville, P. M. (1963). Alkali ion exchange between vapor and feldspar phases, *Am. J. Sci. 261,* 678-683.

Padovani, E., and J. Carter (1977). Aspects of the deep crustal evolution beneath south central New Mexico, in *The Earth's Crust,* J. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 20, pp. 19-55.

Papike, J. J., K. L. Cameron, and K. Baldwin (1974). Amphiboles and pyroxenes: characterization of other than quadrilateral components and estimates of ferric iron from microprobe data, *Geol. Soc. Am. Abstr. Programs 6,* 1053-1054.

Phinney, D., J. Tennyson, and U. Frick (1978). Xenon in CO_2 well gas revisited, *J. Geophys. Res. 83,* 2312-2319.

Reiter, M., C. Shearer, and C. L. Edwards (1978). Geothermal anomalies along the Rio Grande rift in New Mexico, *Geology 6,* 85-105.

Rogers, N. W (1977). Granulite xenoliths from Lesotho kimberlites and the composition of the lower crust, *Nature 270,* 681-684.

Schmid, R., and B. J. Wood (1976). Phase relationships in granulitic metapelites from the Irvea-Verbano zone (northern Italy), *Contrib.Mineral. Petrol. 54,* 255-279.

Shieh, Y.-N, and H. P. Schwarcz (1974). Oxygen isotope studies of granite and migmatite, Grenville province of Ontario, Canada, *Geochim. Cosmochim. Acta 38,* 21-45.

Shimazu, M., T. Yano, and M. Tazima (1978). Gabbroic inclusions in calc-alkalic rocks of the Fossa Magna, central Japan (abstr.), in *International Geodynamics Conference "Western Pacific" and "Magma Genesis," Tokyo, March 13-17, 1978,* Inter-Union Commission on Geodynamics, and Science Council of Japan, pp. 324-325.

Simmons, G., and D. Richter (1978). Microcracks in rocks, in *The Physics and Chemistry of Minerals and Rocks,* R. Strens, ed., Wiley-Interscience, New York, pp. 105-137.

Smith, D. (1977). Hydrous minerals and carbonates in peridotite inclusions from the Green Knobs and Buell Park kimberlitic diatremes on the Colorado Plateau, in *Extended Abstracts of the Second International Kimberlite Conference,* 1977.

Smith, D., and S. Levy (1976). Petrology of the Green Knobs diatreme and implications for the upper mantle below the continental plateau, *Earth Planet. Sci. Lett. 29,* 107-125.

Smithson, S., and S. Brown (1977). A model for lower continental crust, *Earth Planet. Sci. Lett. 35,* 132-144.

Smithson, S., J. Brewer, S. Kaufman, J. Oliver, and C. Hurich (1978). Nature of the Wind River thrust, Wyoming, from COCORP deep-reflection data and from gravity data, *Geology 6,* 648-652.

Stephenson, P. J., and T. J. Griffin (1976). *Cainozoic Volcanicity North Queensland,* Guidebook Excursion No. 7A, 25th International Geologic Congress, 39 pp.

Stoesser, D. B. (1973). Mafic and ultramafic xenoliths of cumulus origin, San Francisco volcanic field, Arizona, Ph.D. dissertation, U. of Oregon.

Takahashi, E. (1978). Petrological model of the upper mantle and the lower crust of the island arc: petrology of mafic and ultramafic xenoliths in Cenozoic alkali basalts of the Oki-Dogo Island in the Japan Sea (abstr,), in *International Geodynamics Conference "Western Pacific" and "Magma Genesis," Tokyo, March 13-17, 1978,* Inter-Union Commission on Geodynamics, and Science Council of Japan, pp. 334-335.

Tarney, J. (1976). Geochemistry of Archean high-grade gneisses, with implications as to the origin and evolution of the Precambrian crust, in *The Early History of the Earth,* B. Windley, ed., Wiley-Interscience, New York, pp. 405-417.

Taylor, S. R. (1977). Island arc models and the composition of the continental crust, in *Deep Drilling Results in the Atlantic Ocean,* Am. Geophys. Union Maurice Ewing Series, Vol. 1, pp. 229-242.

Taylor, H. P., and B. Turi (1976). High^{18}O igneous rocks from the Tuscan magmatic province, Italy, *Contrib. Mineral. Petrol. 55,* 33-55.

Turi, B., and H. P. Taylor (1976). Oxygen isotope studies of potassic volcanic rocks of the Roman province, central Italy, *Contrib. Mineral. Petrol. 55,* 1-31.

Vogel, D. E., and G. D. Garlick (1970). Oxygen-isotope ratios in metamorphic eclogites, *Contrib. Mineral. Petrol. 28,* 188-191.

Wass, S., and A. J. Irving (1976). *Xenmeg: A Catalogue of Occurrences of Xenoliths and Megacrysts in Basic Volcanic Rocks of Eastern Australia,* Australian Museum, Sydney, 441 pp.

White, A. J. R. (1964). Clinopyroxenes from eclogites and basic granulites, *Am. Mineral. 49,* 883-889.

White, A. J. R., and B. W. Chappell (1976). Ultrametamorphism and granitoid genesis, *25th Int. Geol. Congr. Abstr.* 3,674-675.

Whitford, D. J., W. Compston, I. Nicholls, and M. Abbott (1977). Geochemistry of late Cenozoic lavas from eastern Indonesia: role of subducted sediments in petrogenesis, *Geology 5,* 571-575.

Whitney, J. A., and J. C. Stormer, Jr. (1977). Geothermometry and geobarometry of epizonal granitic intrusions: a comparison of iron-titanium oxides and coexisting feldspars, *Am. Mineral. 61.* 751-761.

Wyllie, P. J. (1971). Experimental limits for melting in the earth's crust and upper mantle, in *The Structure and Physical Properties of the Earth's Crust,* J. Heacock, ed., Am. Geophys. Union Geophys. Monogr. 14, pp. 279-301.

Wyllie, P. J., W.-L. Huang, C. R. Stern, and S. Maaløe (1976). Granitic magmas: possible and impossible sources, water contents and crystallization sequences, *Can. J. Earth Sci. 13,* 1007-1019.

Yoder, H. S., and C. E. Tilley (1962). Origin of basalt magmas: an experimental study of natural and synthetic rock systems, *J. Petrol. 3,* 342-532.

Zartman, R. (1974). Lead isotopic provinces in the Cordillera of the western United States and their geologic significance, *Econ.Geol. 69,* 792-805.

Zielinski, R. A., and P. W. Lipman (1976). Trace-element variations at Summer Coon volcano, San Juan Mountains, Colorado. and the origin of continental-interior andesites, *Geol. Soc. Am. Bull. 87,* 1477-1485.

13

Geochemical Evolution of the Continental Crust

Gilbert N. Hanson
State University of New York at Stony Brook

INTRODUCTION

To place limits on possible origins of the earth's continental crust it is necessary to understand how the earth's crust has evolved and how the various processes acting have modified the geochemistry of the pre-existing crust. Prior to about 3900 million years (m.y.) ago the earth as well as the moon must have undergone significant infall of very large extraterrestrial bodies (Smith, 1976). This bombardment must have played a significant role in crustal evolution. However, on the earth the record of this event has yet to be found. Thus speculations on the geochemical evolution of the continental crust based on the lithological record must start from 3800 m.y. ago, the age of the oldest terrestrial rocks found so far.

The main purpose of this chapter is to suggest isotopic and trace-element approaches useful for studies leading to a better understanding of the geochemical evolution of the earth's continental crust. There are a number of recent papers pertinent to this topic, for example, Lowman (1976); Tugarinov and Bilikova (1976); Smithson and Decker (1974); Smithson and Brown (1977); Hargraves (1976); Taylor (in press); Tarney and Windley (1977); Armstrong and Hein (1973); Jahn and Nyquist (1976); Moorbath (1977); Heier (1973); Tarney (1976); Collerson and Fryer (1978); Green (1972); Pankhurst (1977); Brooks *et al.* (1976b); O'Nions and Pankhurst (1978); Oversby (1978); Engel *et al.* (1974); Shaw (1976); and O'Nions *et al.* (1979).

Models for the evolution of the crust can be placed between two extreme schools of thought (also see Chapter 15). One is that the continental crust formed early in the history of the earth (during the Arcbean) and that only small fractions of material have been added since then. The other model is that the continental crust has grown substantially since the Archean. Both models acknowledge the more or less continuous addition of igneous rocks into or upon the upper continental crust. There are, however, two possible sources for this material, the mantle or the lower crust. Material added from the mantle will, of course, increase the mass of the continents, whereas material derived from the lower crust will not change the mass of the continental crust but only redistribute matter within it.

The continental crust makes up only 0.3 percent of the mass of the earth, but it is strongly enriched in elements such as K, U, Th, Rb, Ba, and Sr (Gast, 1960). Based on heat-flow data and the abundances of K, U, and Th, Heier (1973) suggests that the lower crust has lower abundances of these elements than the upper crust and that granulite-grade rocks of intermediate composition are reasonable candidates for the lower-crust composition. A more extensive argument for this model is presented by Smithson and Brown (1977).

One of the most important factors in any interpretation of crustal evolution is how mantle convection has changed with time. In the plate tectonic model, the crust is a passive feature riding on lithospheric plates, the motions of which are determined by convection within the asthenosphere. The igneous as well as tectonic activity within the crust is directly or indirectly related to activity in the asthenosphere. Thus, to understand crustal evolution it is essential also to understand the present convection regimes of the mantle, how these regimes may have evolved with time, and the possible interactions that various parts of the mantle may have had with the continental crust.

As a first approximation, the upper mantle may be divided into two parts: the suboceanic mantle and the subcontinental mantle. Based on isotope and trace-element ratios for basalts, there are two principal sources of magma in the suboceanic mantle: one is the source of ocean-ridge basalts, the other the source of the ocean-island basalts. Radiogenic isotope data would suggest that the sources are separate and have been isolated for some 2000 m.y. (Church and Tatsumoto, 1975; Brooks *et al.,* 1976a; Sun and Hanson, 1975). The ocean-ridge types of basalts appear to be restricted to zones of spreading either at ocean ridges or in marginal basins; basalts of the ocean-island type occur in nearly every tectonic environment in the oceans and continents (Schwarzer and Rogers, 1974). Where the ocean-island-type basalts occur on continents, there may be little reaction with the continental crust (e.g., Ross Island, Sun and Hanson, 1976). The mantle source for continental basalts (a large and geochemically variable group of basalts), however, may in some cases have a history associated with the continents (Peterman *et al.,* 1970; Leeman, 1975; Brooks *et al.,* 1976b), and the source may have interacted or mixed with crustal components (Faure *et al.,* 1972; 1974).

RADIOGENIC ISOTOPES

Some of the key data for understanding the evolutionary history of sources for igneous rocks are the initial isotope ratios of Pb, Sr, and Nd. It must be emphasized that the initial ratios alone cannot be used to tell whether the immediate source of a rock is the mantle or the crust. The isotopic ratios only allow an estimation of the U/^{204}Pb, Rb/Sr, and Sm/Nd ratios of the source and a determination of the time these ratios existed. If continental evolution involves input of significant quantities of igneous rocks derived from the mantle, it is important to understand how the subcontinental and suboceanic mantle regimes have evolved.

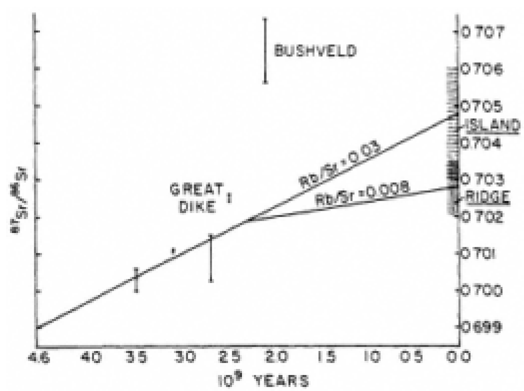


Figure 13.1
Strontium evolution diagram for mantle with data for basic and ultramafic rocks modified from Jahn and Nyquist (1976), with data for the Great Dyke and Bushveld Complex from Hamilton (1977). "Island" designates field for ocean-island basalts. "Ridge" designates field for ocean-ridge basalts.

Figure 13.1 shows some schematic mantle evolution curves for Sr. The large variation in ^{87}Sr/^{86}Sr in modem oceanic basalts indicates that the suboceanic mantle has considerable heterogeneity. This heterogeneity may have also existed in the Precambrian, but the limited number of basaltic rocks analyzed may not adequately sample the Precambrian mantle. Hamilton (1977) suggests that the initial ^{87}Sr/^{86}Sr ratios for the 2100-m.y.-old Bushveld Complex, which vary from 0.7056 to 0.7086, may reflect a heterogeneous mantle source variably enriched in Rb/Sr and is not a result of mixing with crustal components. If he is correct, prior to 2100 m.y. ago the subcontinental mantle in the vicinity of the Bushveld Complex had been variably enriched in Rb/Sr for a significant period of time.

Veizer and Compston (1976) have determined initial Sr isotope ratios on sedimentary carbonates throughout the geological record. If these values represent carbonates from oceanic environments, they should indicate the average Sr isotope ratios of the rocks supplying Sr to the oceanic environment. It can be seen in Figure 13.2 that the Sr isotope ratios in the Archean are low, typical of values assumed for the mantle. This may indicate that if the continents were extensive in the Archean, either they had low ^{87}Sr/^{86}Sr ratios and low Rb/Sr ratios or, if the continents had higher ^{87}Sr/^{86}Sr ratios, the strontium in the oceanic environment was predominantly derived from volcanic regimes and thus reflected a mantle source. After the Archean, the ^{87}Sr/^{86}Sr ratio of the carbonates increases significantly. This would suggest that the continental source is more exposed and volcanics are less of a source or that there is significant growth of the continental crust at the end of the Archean. The same evolutionary relationship can be seen in the K/Na ratio of sedimentary and volcanic rocks (Engel *et al.,* 1974) and in the rare-earth elements content of sediments (Taylor, in press).

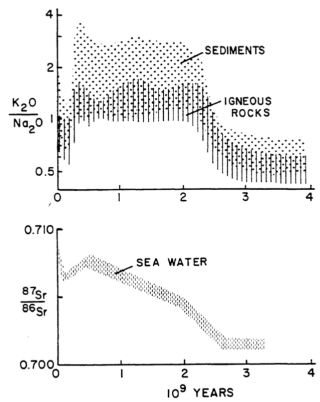


Figure 13.2
$^{87}Sr/^{86}Sr$ ratios in sedimentary carbonates (Veizer and Compston, 1976) and K_2O/Na_2O in sediments and volcanics (Engel et al., 1974) as a function of the age of the rocks.

Figure 13.3 is a single-stage mantle growth curve for Pb on a $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ plot. The data from modem oceanic basalts indicate that there is not a simple growth curve for the recent mantle. The oceanic basalt data lie along lines with slopes the ages of which are approximately 2000 m.y., suggesting that some 2000 m.y. ago their sources were variably enriched in the $^{238}U/^{204}Pb$ ratio relative to the growth curve. Some basalts from presumed subcontinental mantle show a quite different relationship. For example, the Pb isotope data for Absaroka volcanics from Wyoming (Peterman *et al.,* 1970) lie about a line with a slope of 2800 m.y. These rocks, whether derived from the mantle or the lower crust, indicate a source that has had a low $^{238}U/^{204}Pb$ ratio with respect to the mantle growth curve since 2800 m.y. ago. This age is approximately that of the basement rock in this region. Leeman (1975) found similar results for basalts from the Snake River Plain as well as from Yellowstone National Park. He suggests that the trace-and major-element composition of the basalts require their derivation from the mantle. In both studies, the lead and strontium isotopes are not correlated and cannot be explained by a simple mixing relation between crust-and mantle-derived end members. These studies suggest that in these regions the subcontinental mantle has been attached to the continental crust as a mantle keel since at least 2700 m.y. ago. The volcanics from these areas have $^{87}Sr/^{86}Sr$ ratios of 0.704-0.709, on the higher end of the oceanic basalts, suggesting that although their source was depicted in U relative to Pb it was not depleted in Rb relative to Sr. If anything, it was enriched.

Based on initial Sr, Pb, and Nd ratios, many granitic and basaltic rocks would appear to have either a source in the mantle or a source with only a short history in the crust (e.g., Moorbath, 1977; McCullogh and Wasserburg, 1978; DePaolo and Wasserburg, 1976; and O'Nions *et al.,* 1979). This suggests an episodic and continuous addition of material to the crust through time. Further geochemical study of rocks derived from crustal sources, but with essentially mantle ratios, may make it possible to place limits on how the crust evolved and the times involved. Likewise, further geochemical study of rocks derived from the mantle may allow a characterization of the scale of heterogeneities in the mantle, show how they are evolving through time, and help to distinguish parts of the mantle interacting with the continental crust at a given time. As convection models for the mantle improve, this information should allow a direct correlation between ancient tectonic regimes and convection in the mantle.

PETROGENESIS OF IGNEOUS ROCKS

Petrogenetic studies emphasizing isotope and trace-element analyses of a suite of igneous rocks are particularly pertinent for placing limits on the geodynamic factors in the mantle responsible for tectonic activity in an area at the time of formation of the suite of rocks. The purpose of a petrogenetic study of an igneous suite is to determine the chemical and mineralogical composition of the source rocks at the time of melting; the history of the sources prior to melting; the extent of melting; the temperature (T) and pressure (P) or depth conditions during

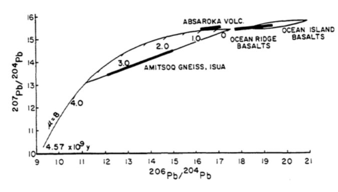


Figure 13.3
Mantle growth curve for Pb with selected rock types plotted (modified from Tatsumato, 1978). Data for Absaroka volcanics are from Peterman et al. (1970). Data for Amitsoq gneisses, Isua, W. Greenland, are from Moorbath et al. (1975).

melting; and modification of the primary melts by differentiation, assimilation, metasomatism, or late-stage fluids. Although a petrogenetic study relies heavily on major-, minor-, and trace-element analysis and isotopic ratios, it must be based on rocks for which the field, geochronological, and petrologic relations are well understood. The major-element analyses when combined with modal mineral data allow a comparison with experimental studies for estimating *T, P,* and volatiles during melting or differentiation. The isotopic data for St, Pb, and Nd allow an estimate to be made of the history of the U/Pb, Rb/Sr, and Sm/Nd ratios of the source. Modeling of trace elements allows an estimate of the trace-element composition of the source, the mineral composition of the residue at the time of removal of the melt, sequences of fractional crystallization, and the extent of these processes. Along with the initial radiogenic isotope ratios, the trace elements allow an estimate of the extent of mixing or reactions with other melts or rocks (Vollmer, 1976; Langmuir *et al.,* 1978).

To obtain the maximum information, each of the different types of analyses must be made on the same samples. There are few places where a complete study can be made in one laboratory, and it may be questioned as to how many suites of rocks require such detailed analysis. The extensive data, however, are warranted for selected suites, because they can lead to a more quantitative insight into crustal evolution. Once the data are available, the best petrogenetic interpretation may not be immediately obvious but will probably lead to new approaches and models.

As an example of tectonic application, the petrogenesis of granitic rocks in two tectonic settings will be compared. The two settings are (1) an intrusive granite-greenstone belt in northeastern Minnesota in which all the rocks dated give ages of 2700 m.y. (Arth and Hanson, 1975) and (2) a high-grade gneiss terrane in southwestern Minnesota with ages as old as 3600 m.y. (G. N. Hanson, State University of New York at Stony Brook, in preparation).

In the northeastern Minnesota greenstone belt the initial $^{87}Sr/^{86}Sr$ ratios of basic, as well as granitic, rock are all between 0.700 and 0.701, suggesting that they were derived from a mantle source or sources with high Rb/Sr ratios that existed for only a short period of time prior to melting. Dacitic and tonalitic rocks have K/Rb and Rb/Sr ratios similar to those of Archean tholeiite and strongly depleted heavy rare-earth element patterns. The model that best fits the data is that the dacites and tonalites are derived by partial melting of a tholeiitic parent, probably derived from an oceanic mantle, leaving a residue of garnet and clinopyroxene.

The quartz monzonites from the greenstone belt have lower K/Rb ratios and higher Rb/Sr ratios than the tonalites and dacites, and rare-earth element patterns similar to that of the tonalites and dacites but with higher abundances and negative Eu anomalies. The best model for the origin of the quartz monzonites is partial melting (upper amphibolite grade) of short-lived (<<50 m.y.) greywacke. The greywacke consists of dacitic and tholeiitic detritus derived from within the greenstone belt that has been enriched in K and Rb by sedimentary, processes.

In this greenstone belt all the components are thought to be derived from either the mantle or from rocks with short histories outside the mantle. The belt probably developed on an oceanic crust. If there were a continental crust underlying the greenstone belt, it was apparently not a major source for the volcanic or intrusive rocks analyzed.

In the high-grade gneiss terrane in southwestern Minnesota, the 3600-m.y.-old Morton and Montevideo gneisses were intruded by granitic rocks at 3100, 2600, and 1800 m.y. (Goldich *et al.,* 1970; S. S. Goldich, Northern Illinois University, and J. Wooden, Lockheed Electronics Company, in preparation; S. S. Goldich and C. E. Hedge, U.S. Geological Survey, in preparation). The gneisses vary from quartz diorite through quartz monzonite, and the intruding granitic rocks are granodiorite to quartz monzonite. The rare-earth element patterns for the gneisses and the later intruding granites are all very similar to one another, suggesting that they have similar sources. These patterns are quite distinctive from those of the tonalites but similar to those of the quartz monzonites from northeastern Minnesota. Based on the trace-element abundances and the geological relations, the best model is that the gneisses and the later granites are derived from melting of similar sources, presumably the lower continental crust. This model is supported by Pb isotope data (Doe and Delevaux, in press), which suggest that the later granites are derived from related sources with a significantly long history in the crust. The K content of the gneisses, mainly tonalites, is generally lower than that of the later granites, mainly granodiorites to quartz monzonites. Lower K content for high-grade metamorphic rocks as compared with those of lower grade is not unusual (Heier, 1973). This might imply that the gneisses originally formed under conditions that led to melts of lower K content or that the gneisses have lost K since the time of their origin.

These two examples of petrogenesis would indicate that although the major-element compositions of quartz monzonite and quartz diorite are similar in both the intrusive granite-greenstone belt and the ancient gneiss terrane, a more careful study of their chemistry and relations shows that the similarity is superficial and that the origins are probably quite different. The greenstone belt developed in a short period of time and consists of rocks derived principally from the mantle or from rocks with a short history outside the mantle; whereas the gneiss terrane developed over a longer period of time, and the principal source for the granite rocks appears to be the melting of pre-existing crustal sources.

CHEMISTRY OF THE LOWER CRUST

Based on heat-flow data, Heier (1973) suggested that the lower crust has lower abundances of K, U, and Th than the upper crust. If the lower crust is made up of granulite-grade rocks of intermediate composition, this could fit a model of a depleted lower crust and an enriched upper crust, because most granulite rocks are relatively depleted in K, U, and Th with normal abundances of Sr and Ba compared with similar rocks of lower grade. This results in higher K/Rb (commonly 500 or greater) and lower Sr/Ba (~10), Rb/Sr (~0.02), and U/Pb ratios in granulite rocks (Tarney and Windley, 1977). The depletion in U is reflected in the low U/Pb ratios found in some granulite-grade rocks, leading to whole-rock leads that plot along isochrons below the mantle growth curve and to the left of the geochron in Figure 13.3.

Is the relative depletion of these elements inherent in the origin of the types of rocks found in a granulite terrane, or have the rocks in a granulite terrane preferentially lost some of these elements? If the granulites have lost these elements, there are two means of transport: as melts or in aqueous or other solutions. An important difference between granulite and lower grades of metamorphism is the lower water content in the granulite facies rocks. Some of the loss of elements may thus be associated with the loss of water. One of the more surprising discoveries was that whereas fluid inclusions in rocks of amphibolite grade are rich in H_2O, fluid inclusion in granulite-grade rocks have high proportions of CO_2 (Touret, 1974), suggesting that the fluids with which they were in contact during high-grade metamorphism were CO_2 rich. Goldsmith (1976) reminded us that a very important mineral in the lower crust is scapolite and that scapolite is a mineral into which substantial fractions of CO_3, SO_4, and C1 can be placed. He suggests that much of the carbonate is em-placed in the granulite terrane directly from the mantle.

Lloyd and Bailey (1975) in studying peridotite nodules from the subcontinental mantle have found metasomatic textures, suggesting that normal lherzolite has been meta-somatized, resulting in the growth of titaniferous phlogopite, amphibole, diopside-salite, ferroaugite, titanomagnetite, sphene, perovskite, apatite, and calcite in what was originally lherzolite. It thus appears that many elements may be mobile in the mantle and are being added to the subcontinental mantle in carbonic or aqueous solutions. Wendlandt and Harrison (1978), for example, found that under mantle conditions CO_2 vapor is 3 orders of magnitude more enriched in rare-earth elements than is aqueous vapor and is also enriched in rare-earth elements relative to silicate melts.

Shieh and Schwarcz (1974) have shown that the oxygen isotope ratios in rocks of the amphibolite grade in the Grenville province are characteristic of their unmetamorphosed equivalents, whereas rocks of the highest metamorphic grade have oxygen isotope ratios more indicative of the mantle. A similar relation has been found for Archean rocks in the Superior province (Longstaffe and Schwarcz, 1977).

Although transporting material in the form of siliceous melts from the mantle to the crust or from the lower crust to the upper crust is undoubtedly important in terms of quantities of material moved, the effects of aqueous or carbonic vapors or fluids in transporting material within the mantle, from the mantle to the crust, or within the crust may be significant. Particularly, they may play an important role in separating elements that behave similarly during magmatic processes. The solubilities of elements in these fluids and mineral-fluid distribution coefficients must be determined experimentally under a variety of conditions so that a proper evaluation of these processes may be made.

MANTLE-CRUST INTERACTION

To understand the evolution of the continental crust it is necessary to understand how the mantle interacts with the continental crust. This requires characterizing variations within the mantle, determining their dimensions, and assessing how the variations are affected by mantle convection. Sun and Hanson (1975) suggested that Rb-Sr and Pb-Pb isochron ages for ocean-island basalts of about 2000 m.y. reflect a real time of separation and isolation of the mantle sources for ocean-ridge and ocean-island basalts and that they are not the result of simple mixing between a large-ion-lithophile-element-(LIL) depleted ocean-ridge source and a LIL-enriched ocean-island source. This is best shown in a plot of $^{87}Sr/^{86}Sr$ versus $^{206}Pb/^{204}Pb$, in which the ocean-ridge basalt plots away from the main trend of the data for the ocean islands and not at either end of a potential mixing curve. Although ocean-ridge basalts are only known to occur in spreading centers, whether at ocean ridges or in marginal basins, these environments encircle the globe. The ocean-island basalts are found in continental, island-arc, and oceanic terranes seemingly unrestricted in their geographic occurrence. Thus both sources appear to be ubiquitous but separated. Until we have better information regarding convection in the mantle, the simplest model to explain these observations is a stratified mantle in which the source for the ocean-ridge basalts is a convecting mantle, below which is the source for the ocean-island basalts. This lower source may also be convecting (F. Richter, University of Chicago, personal communication, 1978). Applying this model to a continental environment, there may be a continental mantle keel attached to the continental crust for hundreds to thousands of millions of years (Figure 13.4). In this model, starting from the left side of the figure and using the numbers in Figure 13.4: (1) Perturbations in the convecting mantle produce upwelling, rifting, and melting of the continental mantle with the formation of continental basalts. The wide variety of these melts may or may not be a result of reaction with or melting of the continental crust. (2) Carbonatites or kimberlites may result from melting or instability near the low-velocity zone. (3) Ocean-island-type basalts found on the continents are associated with deep-mantle plumes. (4) The addition of CO_2 to the lower crust may be a result of continued production of CO_2 over wide areas in the mantle that reacts with the granulite-grade rocks in the lower crust, or it may be episodic, associated with tectonic disturbance.

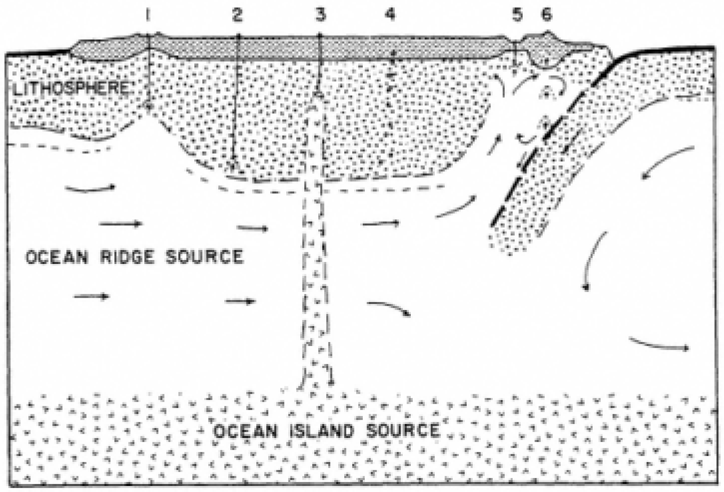


Figure 13.4
Diagrammatic representation of present-day mantle.

Figure 13.4 also depicts a subduction zone (far right) on the continental margin in which there is extensive tectonic activity, (5) the formation of a marginal basin, and (6) in the are, volcanism and the intrusion of gabbroic through granitic plutons. Below the are there may be melting of: the subducted plate to produce tonalities; the subcontinental mantle or the ocean-ridge-type mantle to produce basalts; mafic rocks near the base of the crest to produce anorthositic or gabbroic plutons and possibly an-desires; and the intermediate-composition continental crest to produce granitic intrusions. The melting is probably enhanced by the dehydration of the subducting plate. In the marginal basin the first volcanics would be derived by melting of the subcontinental mantle. As rifting proceeds and the marginal basin widens, ocean-ridge-type mantle becomes the dominant source of basalt.

Detailed petrogenetic studies of suites of modem igneous rocks should allow testing of this and other models. Similar studies on other geological time spans should allow an evaluation of the evolution of mantle regimes, mantle convection, and the interaction of the mantle with the continental crust.

ACKNOWLEDGMENTS

S. R. Hart and J. Wooden reviewed the manuscript. This report was supported by NSF Grant No. EAR 76-13354 A01 (Geochemistry).

References

Armstrong, R. L., and S. M. Hein (1973). Computer simulation of Pb and Sr isotope evaluation of the earth's crust and upper mantle, *Geochim. Cosmochim. Acta 37,* 1-18.

Arth, J. G., and G. N. Hanson (1975). Geochemistry and origin of the early Precambrian crust of northern Minnesota, *Geochim. Cosmochim. Acta 39,* 325-362.

Brooks, C. R., S. R. Hart, A. Hoffman, and D. E. James (1976a). Rb-Sr mantle isochrons from oceanic regions, *Earth Planet. Sci. Lett. 32,* 51-61.

Brooks, C., D. E. James, and S. R. Hart (1976b). Ancient lithosphere: its role in young continental volcanism, *Science 193,* 1086-1094.

Church, S. E., and M. Tatsumoto (1975). Lead isotopic relations in oceanic ridge basalts from the Juan de Fuca-Gorda Ridge area N.E. Pacific Ocean, *Contrib. Mineral. Petrol. 53,* 253-279.

Collerson, K. D., and B. J. Fryer (1978). The role of fluids in the formation and subsequent development of the early continental crust, *Contrib. Mineral. Petrol. 67,* 151-167.

DePaolo, D. J., and G. J, Wasserburg (1976). Nd isotope variations and petrogenetic models. *Geophys. Res. Lett. 3,* 249-252.

Doe, B. R., and M. H. Delevaux (in press). Lead isotope investigations in the Minnesota River Valley—I. Late-and post tectonic granites, *Geol. Soc. Am. Mem.*

Engel, A. E. J., S. P. Itson, C. G. Engel, D. M. Stickney, and E. J. Gray, Jr. (1974). Crustal evolution and global tectonics: a petrogenetic view, *Geol. Soc. Am. Bull. 85,* 843-858.

Faure, G., R. L. Hill, L. M. Jones, and D. H. Elliot (1972). Isotope composition of strontium and silica content of mesozoic basalt and dolerite from Antarctica, in *Antarctica Geology and Geophysics,* R. J. Adie, ed., Universitetaforloget, Oslo, pp. 413-418.

Faure, G., J. Bowman, D. Elliot, and L. Jones (1974). Strontium isotope composition and petrogenesis of the Kirkpatrick basalt, Queen Alexandria Range, Antarctica, *Contrib. Mineral. Petrol. 48,* 153-169.

Gast, P. W. (1960). Limitations on the composition of the upper mantle, *J. Geophys. Res. 65,* 4.

Goldich, S. S., C. E. Hedge, and T. W. Stern (1970). Age of the Morton and Montevideo gneisses and related rocks, southwestern Minnesota, *Geol. Soc. Am. Bull. 81,* 3671-3696.

Goldsmith, J. R. (1976). Scapolites, granulites and volatiles in the lower crust, *Geol. Soc. Am. Bull. 87,* 161-168.

Green, D. H. (1972). Magmatic activity. as the major process in the chemical evolution of the earth's crust and mantle, *Tectonophysics 13,* 47-71.

Hamilton, J. (1977). Sr isotope and trace element studies of the Great Dyke and Bushveld mafic phase and their relation to early proterozoic magma genesis in southern Africa, *J. Petrol. 18,* 24-52.

Hargraves, R. B. (1976). Precambrian geologic history: growth and emergence of continents from beneath the primordial sea, *Science 195,* 363-371.

Heier, K. S. (1973). A model for the composition of the deep continental crust, *Fortschr. Mineral. 50.* 174-187.

Jahn, B. M., and L. E. Nyquist (1976). Crustal evolution in the early earth-moon system: constraints from Rb-Sr studies, in *The Early History of the Earth,* B. F. Windley, ed., John Wiley, New York, pp. 55-76.

Langmuir, C. H., R. D. Vocke, Jr., G. N. Hanson, and S. R. Hart (1978). A general mixing equation with applications to Icelandic basalts, Earth Planet. Sci. Lett. 37, 380-392.

Leeman, W. P. (1975). Radiogenic tracers applied to basalt genesis in the Snake River Plain-Yellowstone National Park region—evidence for a 2.7 b.y. old upper mantle keel, *Geol. Soc. Am. Abstr. Programs 7,* 165.

Lloyd, F. E., and D. K. Bailey (1975). Light element metasomatism of the continental mantle: evidence and consequences, *Phys. Chem. Earth 9,* 389-416.

Longstaffe, F. J., and H. P. Schwartz (1977). $^{18}O/^{16}O$ of Archean elastic metasedimentary rocks: a petrogenetic indicator for Archean gneisses?, *Geochim. Cosmochim. Acta 41,* 1303-1312.

Lowman, P. D., Jr. (1976). Crustal evolution in silicate planets: implications for the origins of continents, *J. Geol. 84,* 1-26.

McCullogh, M. T., and G. J. Wasserburg (1978). Sm-Nd and Rb-Sr chronology of continental crust formation, *Science 200,* 1003-1011.

Moorbath, S. (1977). Ages, isotopes and evolution of Precambrian continental crust, *Chem. Geol. 20,* 151-187.

Moorbath, S., R. K. O'Nions, and R. J. Pankhurst (1975). The evolution of early Precambrian crustal rocks at Isua, West Greenland—geochemical and isotopic evidence, *Earth Planet. Sci. Lett. 27,* 229-239.

O'Nions, R. K., and R. J. Pankhurst (1978). Early Archean rocks and geochemical evolution of the earth's crust, *Earth Planet. Sci. Lett. 38,* 211-236.

O'Nions, R. K., S. R. Carter, N. M. Evensen, and P. J. Hamilton (1979). Geochemical and cosmochemical application of Ndisotopes, *Ann. Rev. Earth Planet. Sci. 7,* 11-38.

Oversby, V. M. (1978). Lead isotopes in Archean plutonic rocks, *Earth Planet. Sci. Lett. 38,* 237-248.

Pankhurst, R. J. (1977). Strontium isotope evidence for mantle events in the continental lithosphere, *J. Geol. Soc. 134,* 255.

Peterman, Z. E., B. R. Doe, and H. J. Prostka (1970). Lead and strontium isotopes in rocks of the Absaroka volcanic field, Wyoming, *Contrib. Mineral. Petrol. 27,* 121-130.

Schwarzer, R. R., and J. J. W. Rogers (1974). A world-wide comparison of alkali olivine basalts and their differentiation trends, *Earth Planet. Sci. Lett. 23,* 286-296.

Shaw, D. M. (1976). Development of the early continental crust, Part 2: Prearchean, Protoarchean, and later eras, in *The Early History of the Earth,* B. F. Windley, ed., John Wiley, New York, pp. 33-53.

Shieh, Y. N., and H. P. Schwarcz (1974). Oxygen isotope studies of granite and migmatite, Grenville province of Ontario, Canada, Geochim. Cosmochim. Acta 38, 21-45.

Smith, J. V. (1976). Development of the earth-moon system with implications for the geology of the early earth, in *The Early History of the Earth,* B. F. Windley, ed., John Wiley, New York, pp. 3-19.

Smithson, S. B., and S. K. Brown (1977). A model for lower continental crust, *Earth Planet. Sci. Lett. 35,* 134-144.

Smithson, S. B., and E. R. Decker (1974). A continental crustal model and its geothermal implications, *Each Planet. Sci. Lett. 22,* 215-225.

Sun, S. S., and G. N. Hanson (1975). Evolution of the mantle: geochemical evidence from alkali basalts, *Geology 3,* 297-302.

Sun, S. S., and G. N. Hanson (1976). Differentiation of McMurdo volcanics, Ross Island, Antarctica: evidence from REE, *Contrib. Mineral. Petrol. 54,* 139-155.

Tarney, J. (1976). Geochemistry of Archaean high-grade gneisses, with implications as to the origin and evolution of the Precambrian crust, in *The Early History of the Earth,* B. F. Windley, ed., John Wiley, New York, pp. 405-417.

Tarney, J., and B. F. Windley (1977). Chemistry, thermal gradients and evolution of the lower continental crust, *J. Geol. Soc. London 134,* 153-172.

Tatsumoto, M. (1978). Isotopic composition of lead in oceanic basalts and its implications to mantle evolution, *Earth Planet. Sci. Lett. 38,* 63-87.

Taylor, S. R. (in press). Chemical composition and evolution of the continental crust: the rare earth element evidence, in *The Earth: Its Origin, Structure and Evolution,* M. W. McElhinny, ed., Academic Press, New York.

Touret, J. (1974). Facies granulite et inclusions carboniques, *Ann. Soc. Geol. Belgique, Vol. Spec. P. Michot,* pp. 267-287.

Tugarinov, A. T., and Ye. V. Bilikova (1976). Evolution of the chemical composition of the earth's crust, *Geokhimiya 8,* 1151-1159.

Veizer, J., and W. Compston (1976). $^{87}Sr/^{86}Sr$ in Precambrian carbonates as an index of crustal evolution, *Geochim. Cosmochim. Acta 40,* 905-914.

Vollmer, R. (1976). Rb-Sr and U-Th-Pb systematics of alkaline rocks: the alkaline rocks from Haly, *Geochim. Cosmochim. Acta 40,* 283-295.

Wendlandt, R. F., and W. J. Harrison (1978). Rare earth element partitioning between coexisting immiscible carbonate and silicate liquids and CO_2 in the system K_2O-Al_2O_3-SiO_2-CO_2, *Geol. Soc. Am. Abstr. Programs 10,* 514.

V

CONTINENTAL EVOLUTION

14

Cenozoic Volcanism in the Western United States: Implications for Continental Tectonics

Peter W. Lipman
U.S. Geological Survey

INTRODUCTION

Volcanic rocks can provide insights into major features of continental tectonics, for which in some cases little other record may exist. Thus, study of Cenozoic volcanism in the western United States may reveal structural and compositional constraints on the lithosphere of the American plate and on the geometry of past interactions with other plates. This chapter focuses on four features of continental tectonics: (1) the composition and geometry of underlying and, as yet unexposed, cogenetic magma chambers and batholiths; (2) lateral compositional variations in the lithosphere of the American plate; (3) structural discontinuities in this lithosphere; and (4) the geometry of past interactions between the American plate and various Pacific oceanic plates.

SUBVOLCANIC INTRUSIONS

The study of Cenozoic volcanic activity in the western United States provides information on the distribution and composition of young intrusions at shallow depth. Such intrusions constitute significant upper-crustal continental-tectonic features of major potential economic significance for mineral and geothermal resources. The intrusions are also products of evolving magmatic systems that may have caused long-lived perturbations of the lower crust and upper mantle during processes of partial melting, magma migration, and fractionation.

The close relationship between continental volcanism and emplacement of shallow granitic intrusions was discussed by Hamilton and Myers (1967); they concluded that major silicic volcanism is typically the surface manifestation of batholithic intrusive activity at high crustal levels and that many large batholithic complexes are shallow, grossly lenticular bodies, roofed largely by cogenetic volcanic rocks erupted during immediately preceding stages in the rise and emplacement of the batholith. Publications by these authors and others in the past 10 years have developed generally similar interpretations for volcano-plutonic associations in the western United States, especially for the Boulder batholith of Montana (Hamilton and Myers, 1974; Klepper *et al*., 1971) and large parts of the Sierra Nevada batholith in California (Schweikert, 1976; Fiske *et al*., 1977). By analogous reasoning, the distribution and compositions of volcanic rocks, together with pertinent geophysical data, can be used to infer the geometry of subvolcanic intrusions or even magma chambers beneath young volcanic fields into which erosion has not exhumed widespread intrusive rocks.

Oligocene volcanic activity in the San Juan field, southwestern Colorado, is interpreted as recording the rise, differentiation, and consolidation of a composite batholith, largely of intermediate composition and covering an area of roughly 50 km by 100 km (Steven and Lipman, 1976; Lipman *et al*., 1978). Early eruptions that formed stratovolcanoes of intermediate composition in the San Juan field were accompanied by intrusion of small stocks into the cores of these volcanoes (Figure14.1A). No evidence exists for any shallow magma bodies of batholithic dimensions at this stage, although some of the early volcanoes are clustered, apparently marking sites of concentrated accumulations of magma. Within several of these clusters, sufficient magma subsequently accumulated at shallow depth to permit formation of silicic differentiates, eruption of ash-flow tuffs, and resulting caldera collapses (Figure 14.1B). At least 15 ash-flow calderas, mostly formed within a 2-million-year (m.y.) interval (29-27 m.y. ago), are clustered within terrane characterized by a large, steep-sided, flat-floored, Bouguer gravity low (Figure 14.2) that is interpreted as defining the final shape of the batholith, with the calderas marking isolated higher cupolas.

The dimensions of the Bouguer gravity anomaly in the San Juan Mountains (Figure 14.2) and, by inference, those of the concealed batholith are comparable with those of the Upper Cretaceous Boulder batholith of Montana. In Montana the gross stratigraphy of the Elkhorn Mountains Volcanics, erupted penecontemporaneously with emplacement of the Boulder batholith, is also similar to that of volcanics in the San Juan field; a thick sequence of intermediate-composition lavas and volcaniclastic rocks is capped by culminating more silicic ash-flow eruptions. Ash-flow-related caldera collapses must also have accom

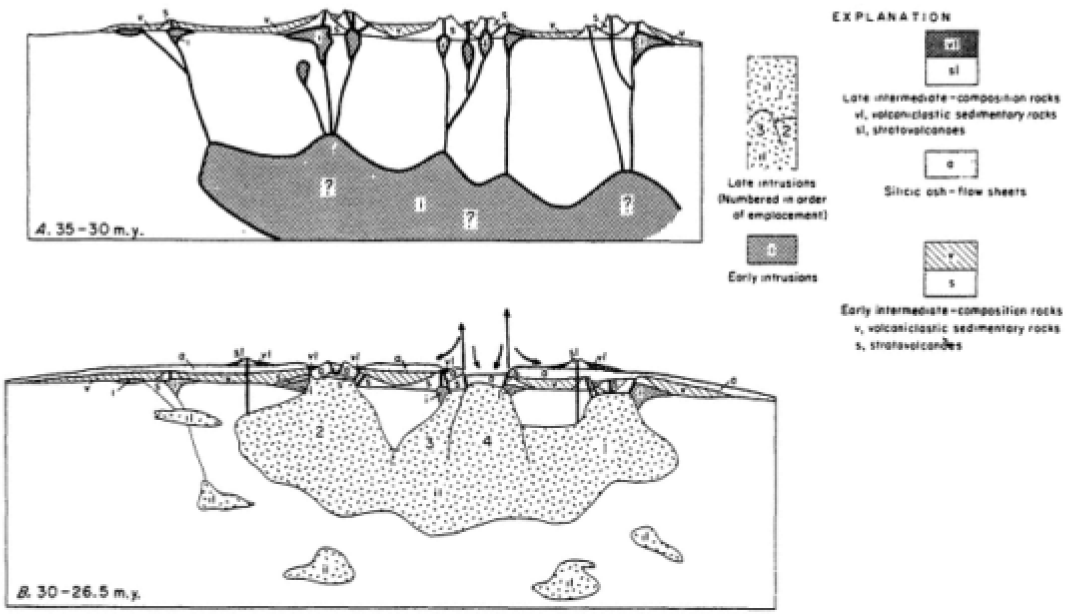


Figure 14.1
Schematic model for evolution of Oligocene subvolcanic batholith in San Juan Mountains (Lipman et al., 1978, courtesy the Geological Society of America). A, Time of early intermediate volcanoes. Clusters of intermediate-composition stratovolcanoes are surrounded by interfingering aprons of volcaniclastic debris. Small intrusions form at shallow levels in volcanic pile, but it is uncertain whether a large high-level intrusive complex has developed by this time. B, Time of ash-flow eruptions and caldera formation. Eruption of complex sequence of ash flows and associated caldera collapses is triggered by accumulation at shallow depth of batholithic-size magma bodies of intermediate to silicic composition. Many of these shallow accumulations are localized within clusters of earlier stratovolcanoes. Some calderas are composite, with younger activity nested within older collapse structures, and many caldera collapses are followed by resurgent doming, indicating renewed upwelling of silicic magma.

panied emplacement of the Boulder batholith, perhaps within the region where only intrusive rocks are currently exposed.

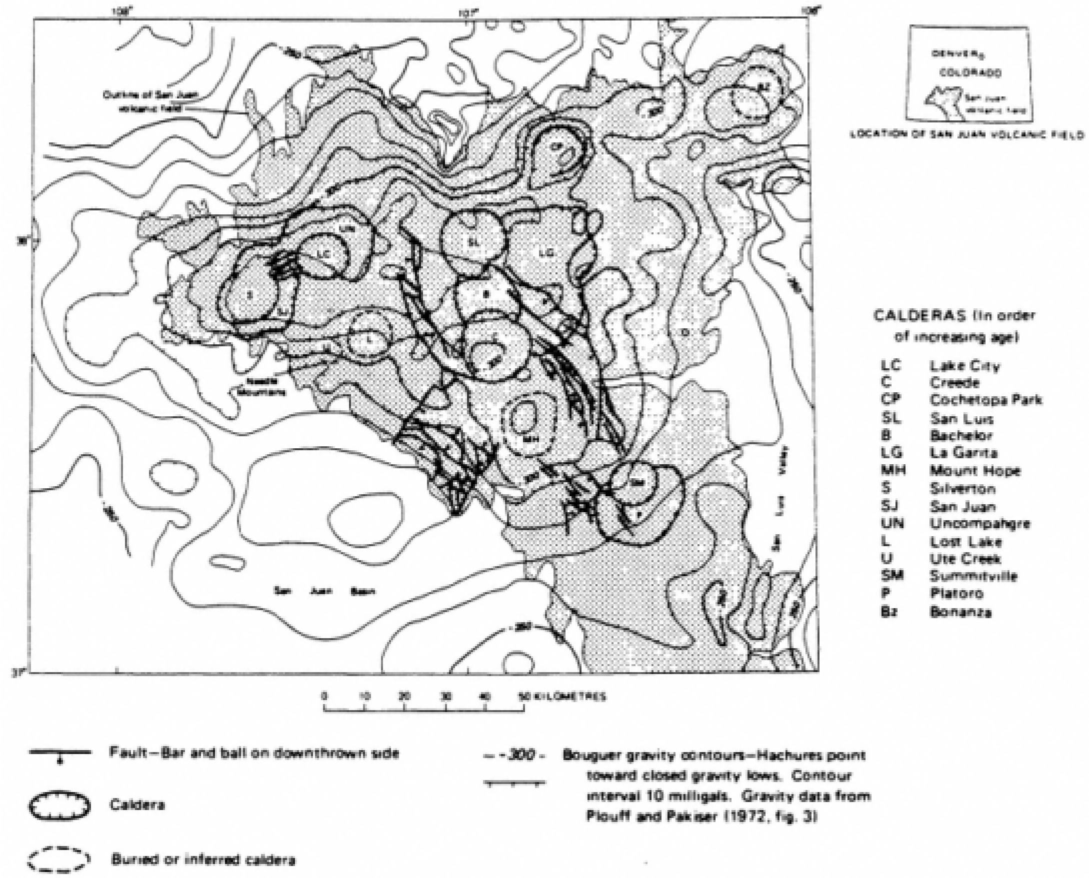


Figure 14.2
Calderas in the San Juan volcanic field in relation to Bouguer gravity field (Steven and Lipman, 1976).

At another major but much younger caldera complex, centered in Yellowstone National Park and related to large-scale ash-flow eruptions about 0.6 m.y. ago, extensive geological and geophysical evidence indicates the presence of a large shallow body of silicic magma (Eaton *et al.,* 1975). This body is thought to underlie an area more than 85 km long and 55 km wide, as indicated by surface geology and by gravity studies. Seismic-attenuation and *P*-wave-residual studies suggest that the top of the magma body is only a few kilometers below the surface and is still molten. The molten magma is interpreted to be underlain by an even larger volume of mechanically and thermally disturbed crustal and mantle rocks that contains pods of basaltic and silicic magma and extends at least 50 km into the mantle—probably through the lithosphere of the American plate (Eaton *et al*., 1975). These features have been interpreted as the geophysical expression of a "gravitational anchor" (Shaw and Jackson, 1973), which re-suited from sinking of dense residue from partial melting episodes during formation of the Yellowstone magmas (Christiansen and McKee, 1978). Here, volcanism is a surface manifestation of geophysical discontinuities that extend to great depth. Questions as yet unaddressed are whether such discontinuities are preserved under older volcanic fields and whether such discontinuities can affect later tectonic and igneous events.

Putting together fragments of the magmatic record in this way, we can determine the main loci of relatively recent plutonic activity, for which erosion has not yet provided direct exposure. Much more should be determinable about such economically important en vironments by combined geological, geochemical, and geophysical studies of transitional volcano-plutonic complexes, especially where interpretation of third-dimensional relations can be augmented by drill-hole data.

COMPOSITIONAL VARIATIONS IN THE CONTINENTAL LITHOSPHERE

Recent volcanic studies using a plate-tectonic framework have emphasized the importance of magma sources below the lithosphere—in extensional environments, from upwelling asthenosphere along spreading oceanic ridges; in convergent environments, from descending slabs along subduction systems below magmatic arcs; and in intraplate settings, from upwelling deep "mantle plumes" (Wilson, 1965; Morgan, 1972; Martin and Piwinskii, 1972). Geochemical studies, especially Sr isotopic data, have tended to emphasize the importance of mantle sources for silicic as well as for marie volcanic rocks and to rule out major contributions from upper-crustal materials (Hurley *et al*., 1962; Peterman *et al*., 1970; Kistler and Peterman, 1973). Such conclusions, initially drawn largely from studies of volcanism in ocean basins, have also been widely held to be generally applicable to continental igneous activity. However, long-recognized contrasts in the nature of volcanic activity between continental and oceanic regions (Gilluly, 1955), as well as additional arguments outlined below, indicate that continental volcanism is strongly influenced by structural and compositional features of the immediately underlying lithosphere, even though the activity may ultimately have been initiated at greater depths. In particular, continental volcanism seems sensitive to the age and composition of the immediately underlying lithosphere, as well as to the geometry of major structural flaws. Implications of the compositions of continental volcanics for compositions of the lower crust and attached lithospheric upper mantle are discussed elsewhere (see Chapter 12). Accordingly, aspects of compositional relations are outlined here only to the degree necessary to support other interpretations of relations between continental volcanism and tectonics.

Recent Pb and Sr isotopic studies of continental volcanic rocks in the western United States indicate that the isotopic composition of volcanic rocks reflects the age and composition of the underlying lithosphere, demonstrating major compositional control by the lower crust or the lithospheric upper mantle. For example, isotopic compositions of both Tertiary volcanic rocks of the Absaroka volcanic field and Quaternary volcanics of the Yellowstone Plateau in the same region (Peterman *et al*., 1970) lie along a well-defined secondary isochron that defines an apparent age of the source region of 2.8 billion years (b.y.)—the same as that of the underlying Precambrian basement (Figure 14.3). In the San Juan Mountains and adjacent parts of the Rio Grande rift in southern Colorado and northern New Mexico, Pb isotopic data for rocks ranging in age from Oligocene to Quaternary yield a secondary isochron that indicates an apparent source age of about 1.7-1.8 m.y.—again the dominant age of the underlying Precambrian basement (Lipman *et al*., 1978). In these two regions the rocks analyzed are from both compressional and extensional tectonic environments and range from basalt to rhyolite. These relations strongly suggest that, whatever the ultimate origin of the volcanism and its thermal requirements, the compositions of the rocks are dominated, at least for Pb, by contributions from the lithosphere of the American plate. In the case of the middle Tertiary rocks of intermediate composition in both the Absaroka and San Juan volcanic fields, for which complex subduction models have been proposed (Lipman *et al*., 1971; 1972), any initial compositional signature from melting of the subducted slab or the immediately overlying asthenospheric mantle has been masked by interactions of the rising magmas with the American plate. For the San Juan field, at least, detailed consideration of the isotopic data suggests a major contribution from lower crustal sources; only minor interactions with uppercrustal rocks are compatible with the Sr isotopic data. Isotopic compositions of upper Tertiary and Quaternary basaltic rocks in the Yellowstone Plateau and Rio Grande rift indicate dominant sources of Pb from mantle regions that have been a part of the American plate since formation of the associated parts of the continental craton, respectively, about 2.8 b.y. and 1.7-1.8 b.y. ago. If the volcanism of the Yellowstone Plateau-Snake River Plain trend represents the trace of a deep mantle plume, as Morgan (1972) and Suppe *et al*. (1975) conjectured, then isotopic compositional identity of the deep source has been lost.

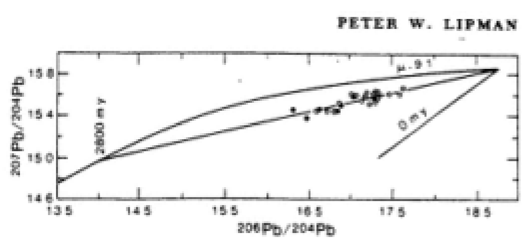


Figure 14.3
Lead isotope relations for the Tertiary Absaroka volcanic field (solid circles) and Quaternary basalts and rhyolites (open circles) from Yellowstone National Park (Peterman et al., 1970).

Variations in Pb isotopic compositions of Cenozoic igneous rocks across the western United States delimit an abrupt discontinuity between a West Coast region and the eastern Cordillera; this discontinuity is interpreted as marking the limit of Precambrian crust (Doe, 1967; Zartman, 1974). Analogous Sr isotopic studies also show strong regional gradients and boundaries that have been interpreted as reflecting major structural boundaries in the underlying continental crust (Kistler and Peterman, 1973; Armstrong *et al*., 1977). In some regions, however, interpretation of the isotopic data is age-dependent; in southern Arizona and New Mexico, mafic Oligocene volcanic rocks are relatively radiogenic ($^{87}Sr/^{86}Sr$ = 0.706-0.708), whereas Miocene and Pliocene basaltic volcanics are significantly less so ($^{87}Sr/^{88}$ Sr = 0.703-0.704). This compositional difference may reflect a change in the composition of the source region, perhaps related to upwelling of deeper mantle material in late Tertiary time, during major lithosphere extension and development of Basin-Range structure (Lipman and Mehnert, 1975).

Interpretation of this sort eventually should increase our understanding of the complex structure and composition of the continental crust and upper mantle of the western United States, where, during Cenozoic time, a plate-convergent regime was superceded by an extensional one as a result of changing plate boundaries (Atwater, 1970). In this region the compositional and structural zones, with which rising and evolving Cenozoic magmas could have interacted, include (1) sialic upper crust that ranges regionally in age from Phanerozoic to as much as 3 b.y. old and that is relatively radiogenic both in Pb and Sr; (2) more. mafic lower crust, probably of less radiogenic isotopic character; (3) lithospheric upper mantle of the American plate, having a depth of about 50-100 km between the Moho and the top of the low-velocity zone, that may have been little modified chemically since formation of the overlying craton; (4) asthenospheric mantle, below about 100 km, of relatively worldwide compositional homogeneity; and (5) at various times and places (as discussed later), a subducted plate descending slowly to the east through the asthenosphere to a depth of several hundred kilometers. This inferred subducted plate consisted of oceanic lithosphere, generated at the East Pacific Rise, and was subjected to little-understood processes and interactions with asthenosphere beneath the American plate for 10-20 m.y. before thermally equilibrating and losing its geophysical identity.

STRUCTURAL DISCONTINUITIES IN THE CONTINENTAL LITHOSPHERE

Structural discontinuities in the lithosphere also seem important in controlling the distribution and type of Cenozoic volcanism in the western United States. Especially conspicuous as controls for Cenozoic volcanism seem to be several northeast-trending zones of different ages and origins: the Snake River-Yellowstone zone, the Springerville-Raton zone, and the Colorado mineral belt (Figure 14.4). Other possible crustal flaws, not discussed here, may also be important in interpreting aspects of continental tectonics.

Proposed interpretation of the Snake River Plain-Yellowstone zone as the trace of a mantle plume or melting anomaly, mentioned earlier, is based largely on the age progression of volcanism during the last 15 m.y., from the Idaho-Nevada border to the Yellowstone caldera (Armstrong *et al*., 1975) and on agreement between this vector and the inferred absolute motion of the American plate (Minster *et al*., 1974). However, geological and geophysical studies in the Yellowstone-Snake River region cast doubt on the mantle-plume model. The Pb isotopic evidence, already mentioned, apparently requires a mantle source for the basaltic magmas that has been part of the American plate for the past 2.8 b.y. Recent geological and geophysical studies in the region have documented other complications. In addition to the northeast-trending age progression during the past 15 m.y., there is an opposing northwest-trending progression—toward Newberry caldera in Oregon (MacLeod *et al*., 1975). The Yellowstone-Snake River zone lies along a regional northeast-trending structural and aeromagnetic lineament that extends from northeastern Nevada, beyond the Yellowstone caldera, into Canada. According to Eaton *et al*. (1975), "if the Yellowstone magma body marks a contemporary deep mantle plume, this plume, in its motion relative to the American plate, would appear to be 'navigating' along a fundamental structure in the relatively shallow and brittle lithosphere overhead. The concept that a northeastward-propagating major crustal fracture controls the migration path of the major foci of volcanism is at least equally favored by existing data."

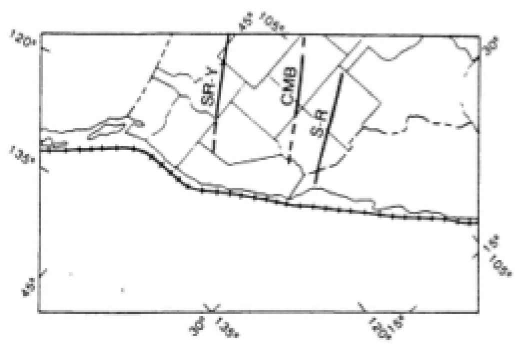


Figure 14.4
Major northeast-trending Cenozoic volcano-tectonic zones in the western United States (shown by solid lines where best constrained, dashed lines where less certain). "Railway track" indicates middle Cenozoic subduction boundary between American and Farallon plates (base map adapted from Atwater, 1970).

Another major northeast-trending igneous zone is the long-recognized Colorado mineral belt, which consists of aligned Upper Cretaceous-lower Tertiary (Laramide) intrusives and locally preserved extrusives. No age progression is evident along the Colorado mineral belt, but, as first pointed out by Tweto and Sims (1963), this belt follows structural trends of Precambrian ancestry. The Colorado mineral belt has been proposed to constitute only part of a very long northeast-trending structural zone that extends from Arizona through Colorado, across the Northem Plains, to join a major boundary between Precambrian age provinces in the Great Lakes region (Warner, 1978).

Farther to the south, another northeast-trending late

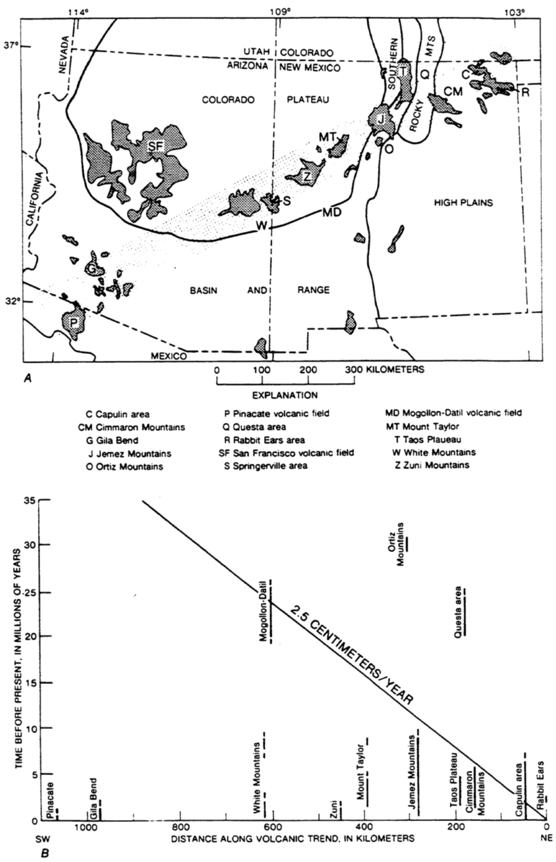


Figure 14.5
Space-time relations of Cenozoic volcanism along the Springerville-Raton zone. A, Distribution of Pliocene and Pleistocene volcanic field along the Springerville-Raton zone (stippled). B, Plot of age of volcanic activity versus position along the Springerville-Raton zone. Sloping line indicates a migration rate of 2.5 cm/year, the observed migration rate for inception of silicic activity along the Snake River Plain-Yellowstone zone; no such age-distance trend is evident for the Springerville-Raton zone.

Cenozoic volcanic feature (Figure 14.5A), referred to variously as the Springerville-Raton zone, the Jemez zone, or the Raton volcanic chain, has also been proposed as marking the trace of a mantle plume (Suppe *et al*., 1975). This major locus of late Cenozoic activity includes the Pinacate basalt field in northwest Mexico, the Springerville-White Mountain volcanic zone in eastern Arizona, the Mount Taylor and Jemez volcanic fields in New Mexico, volcanits in the Rio Grande rift on the Taos Plateau and at the Cerros del Rio, and basalts of the High Plains of northeastern New Mexico. In contrast to the Yellowstone zone, no age progression is evident, either for the upper Cenozoic basaltic rocks or for more silicic earlier Cenozoic activity along this zone (Figure 14.5B); and a mantle-plume origin seems less probable than control by a crustal flaw (Mayo, 1958; Laughlin, 1976). The

Springerville-Raton zone follows general Precambrian structural trends of the region, such as the northeast-trending basement trends of great length on the Colorado Plateau (Shoemaker *et al*., 1974), and coincides in a general way with a major boundary between Precambrian age provinces.

Analysis of patterns of Cenozoic volcanic activity suggests that these northeast-trending zones are key features in interpreting Cenozoic tectonics of the western United States—features identifiable only by the presence of volcanic activity along them.

Another aspect of tectonic control of volcanism that has implications for structure of the lithosphere is the clear pattern of recurrence of igneous (and tectonic) activity in the same places in the western United States through time. Such recurrence of activity was documented by Snyder *et al*. (1976), who delimited five major "magmatic loci" in the western United States. The distinction among several of these loci is not entirely convincing, for example, their Nevada locus is separated from the Idaho-Montana locus to the north only by younger rift-related volcanic cover along the Snake River Plain (Snyder *et al*., 1976). In contrast, the Colorado Plateau was a stable "miniplate" during late Cretaceous-early Tertiary Laramide deformation and igneous activity, during the middle Cenozoic predominantly andesitic volcanism, and during the late Cenozoic fundamentally basaltic volcanism and associated extensional block faulting. The recurrence and general confinement of Cenozoic volcanism to regions where previous episodes of magmatism and tectonism may have healed, annealed, and weakened the lithosphere seems an important consideration in interpreting the distribution of Cenozoic volcanic activity in the western United States.

PLATE-TECTONIC INTERACTIONS

Continental volcanism also provides evidence for the nature of plate interactions in the western United States during the past 100 m.y. For relatively young interactions the seafloor-spreading record is sufficiently complete to permit fairly reliable reconstructions, but for the western Cordillera of North America the plate-tectonic history is complex, the seafloor record is largely missing, and many aspects can be reconstructed only from igneous and tectonic activity recorded on the continental plate. Interpretation of this record is still very incomplete and uncertain.

A contrast exists in the western United States between earlier Cenozoic volcanic assemblages, which are predominantly andesitic in composition and are inferred to be related to convergence and subduction along the western edge of the American plate, and a younger assemblage of fundamentally basaltic volcanism, in places including bimodal basalt-rhyolite associations, that appears related to extensional tectonics within the American plate (Lip-man *et al*., 1971, 1972; Christiansen and Lipman, 1972; Snyder *et al*., 1976). The change from predominantly andesitic to fundamentally basaltic volcanism has been correlated in a general way with changing boundaries between the American and various Pacific plates (Atwater, 1970). Changes in volcanism accompanying the initial intersection of the Pacific and American plates are thought to be especially significant; these changes terminated the middle Tertiary and earlier subduction system and initiated the evolving transform boundary of the San Andreas Fault system, bounded at both ends by migrating triple junctions.

Since first application of this plate model to changing patterns of volcanism in the western United States (Lip-man *et al*., 1971, 1972; Christiansen and Lipman, 1972), new data have accumulated, and further evaluation is appropriate. Refinements of Cenozoic volcano-tectonic patterns in terms of generally similar plate models for the western United States include those of Snyder *et al*. (1976), Coney and Reynolds (1977), Cross and Pilger (1978), Dickinson and Snyder (1979). The re-examination presented here focuses on the predominantly andesitic volcanic suite, inferred to reflect subduction-related regimes. Aspects of the succeeding fundamentally basaltic volcanism and associated extensional tectonics are discussed elsewhere (see Chapter 9).

Important variables that could affect distribution and compositional variation in subduction-related andesitic suites include dip of the Benioff zone, rate of plate convergence, angle between plate boundaries and convergence direction, temperature and thickness of descending slab, warps or breaks in the descending slab, depth of magma generation, composition of andesitic magma, distance from .the trench to volcanic front, and width of the volcanic belt. Compositions of convergent volcanic suites correlate with depth to the Benioff zone, becoming more potassic and alkalic with increasing depth, and vary also from intracontinental to intraoceanic settings (Dickinson, 1975). Volcanoes occur at present above active Benioff zones over a depth range of about 75-300 km. Relatively rapid convergence favors more gently dipping Benioff zones (Luyendyk, 1970) and delayed heating of the descending slab; accordingly, such environments favor increased distances between trench and volcanic front, broader volcanic belts, and more alkalic volcanism. Oblique convergence, transitional toward transform motion, would reduce the effective convergence rate. In active convergent systems, offsets of the volcanic front correlate with segmentation of the Benioff zone, requiring transverse breaks or flexures in the descending slab (Carr *et al*., 1973; Stoiber and Carr, 1974). The behavior of the descending slab may also reflect the age of the oceanic lithosphere, which becomes cooler, thicker, and more dense with age and distance from the spreading ridge; older lithosphere should therefore sink more rapidly and preserve its physical identity to greater depths in a subduction environment than thin, hot, buoyant, young lithosphere (Molnar and Atwater, 1978).

Complex, and as yet imperfectly understood, interplay of these variables seems capable of accounting for much of the diversity of arc volcanism. Some possible effects of change in rate of plate convergence are illustrated diagrammatically in Figure 14.6. In a subcontinental environment, the structural and compositional complexities of the continental lithosphere, discussed earlier, also seem capable of further complicating the volcanic pattern, as illustrated by relations from the western United States. The following discussion is based on a series of figures that show in generalized fashion the distribution of igneous activity through Cenozoic time in 10-m.y. increments (Figure 14.7).

Through most of Mesozoic time and continuing to about 80 m.y, ago, igneous activity was confined to near the western margin of the American plate (Figure 14.7A). Granitic and cogenetic volcanic rocks as young as about 80 m.y. occur in the Sierra Nevada batholith, but Mesozoic igneous rocks are sparse east of western Nevada. The bend in the igneous trend in the Pacific Northwest is probably due largely to Cenozoic deformation (Hamilton and Myers, 1066; Simpson and Cox, 1977). Local eastwest variations in distribution and composition of granitic rocks in the central Sierra Nevada are compatible with shifting depths of magma generation along a subduction system of relatively constant geometry (Dickinson, 1970). More limited, similar data along other transects across the batholithic belt suggest possibly more complex shifts in geometry of the subduction system, especially in the Klamath Mountains, where low-potassium quartz diorites and trondhjemites tend to be concentrated on the eastern side of the plutonic belt (Hotz, 1971), and also in the Peninsular Ranges batholith to the south (Silver *et al*., 1975). Nevertheless, through late Mesozoic time, a long-lived, relatively steeply dipping, subduction system apparently was maintained, along the western margin of the American plate, concurrently with consumption of an eastern Pacific plate.

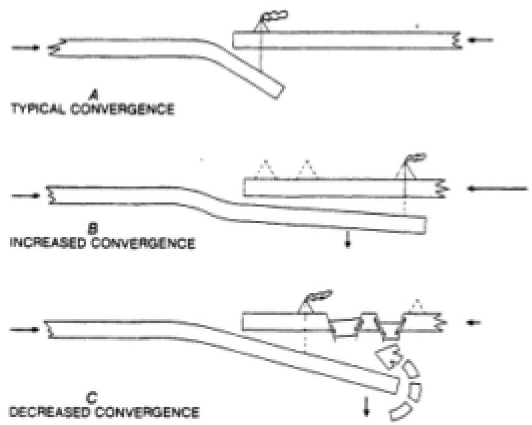


Figure 14.6
Possible effects of change in convergence rate on continental-margin subduction and related volcanism. A, Typical convergence rate: subduction zone dips about 30 degrees, and related volcanoes are in a relatively narrow are, are low in potassium content, and are located relatively close to trench. B, Increased convergence rate: continental plate overrides descending slab at a lower angle, resulting-in a broader, more diffuse volcanic arc located farther from the trench. Depending on depth to subduction zone, volcanic rocks can vary widely in composition from low to high potassium. If the dip of the subduction system becomes very shallow, volcanism may cease entirely, as has probably happened in modem southern Peru (Barazangi and Isacks, 1976). C, Decreased rate of convergence: downgoing slab sinks gravitationally or breaks and re-establishes itself in a steeper orientation, causing are volcanism to migrate toward the trench. Increased sinking of the descending slab requires counterflow of asthenospheric mantle into the region above the slab, possibly heating the base of the continental lithosphere and driving extensional rifting and basaltic volcanism within the region of earlier arc volcanism.

Starting at about 80 m.y. ago, concurrently with inferred accelerated convergence between the American and eastern Pacific plates (Coney, 1972), igneous activity migrated eastward in the western United States (Figure 14.7B), as recognized long ago by Lindgren (1915). The most dramatic eastward shift between 80 m.y. and 70 m.y. ago (Figure 14.7B) occurred in the northwest, where plutonic rocks of the Boulder batholith and associated Elk-horn Mountains volcanics were em-placed as far east as western Montana; however, igneous activity of this age also moved eastward in Nevada and Arizona.

Patterns of igneous activity continued to evolve rapidly and had changed notably by 70 m.y. to 60 m.y. ago (Figure 14.7C). The locus of intense activity associated with the Boulder batholith in western Montana extinguished abruptly about 70 m.y. ago. Many K-Ar ages in the Idaho batholith could indicate continuation of igneous activity; more likely, they reflect reheating by pervasive Eocene activity (Armstrong *et al*., 1977). To the south, major activity of Laramide age continued in southern Arizona and began to extend into southwestern New Mexico, constituting the peak formation of porphyry copper deposits in this region. A southeasterly age progression of these economically important intrusives, suggested as a mantle-plume trace (Livingston, 1973), seems more likely part of the regional eastward migration of igneous activity when considered with distributions of other intrusions of roughly similar age in the region. Between northern Idaho and southern Arizona, the only other significant locus of igneous activity at this time was the northeast-trending zone of intrusions and associated volcanics along the Colorado mineral belt. These represent the most eastern sweep of early Tertiary igneous activity in the central and southern Rocky Mountain region.

Thus, early Tertiary igneous activity in the western United States covered a broad region but was diffuse, discontinuous, and nonsynchronous in time, reaching the eastern margins of the Cordillera earlier in the northern Rockies than farther south. Even with abundant new data, these patterns reasonably seem related to effects of flattening of the subduction system during the early Tertiary (Lipman *et al*., 1971; Coney and Reynolds, 1977). Flatten

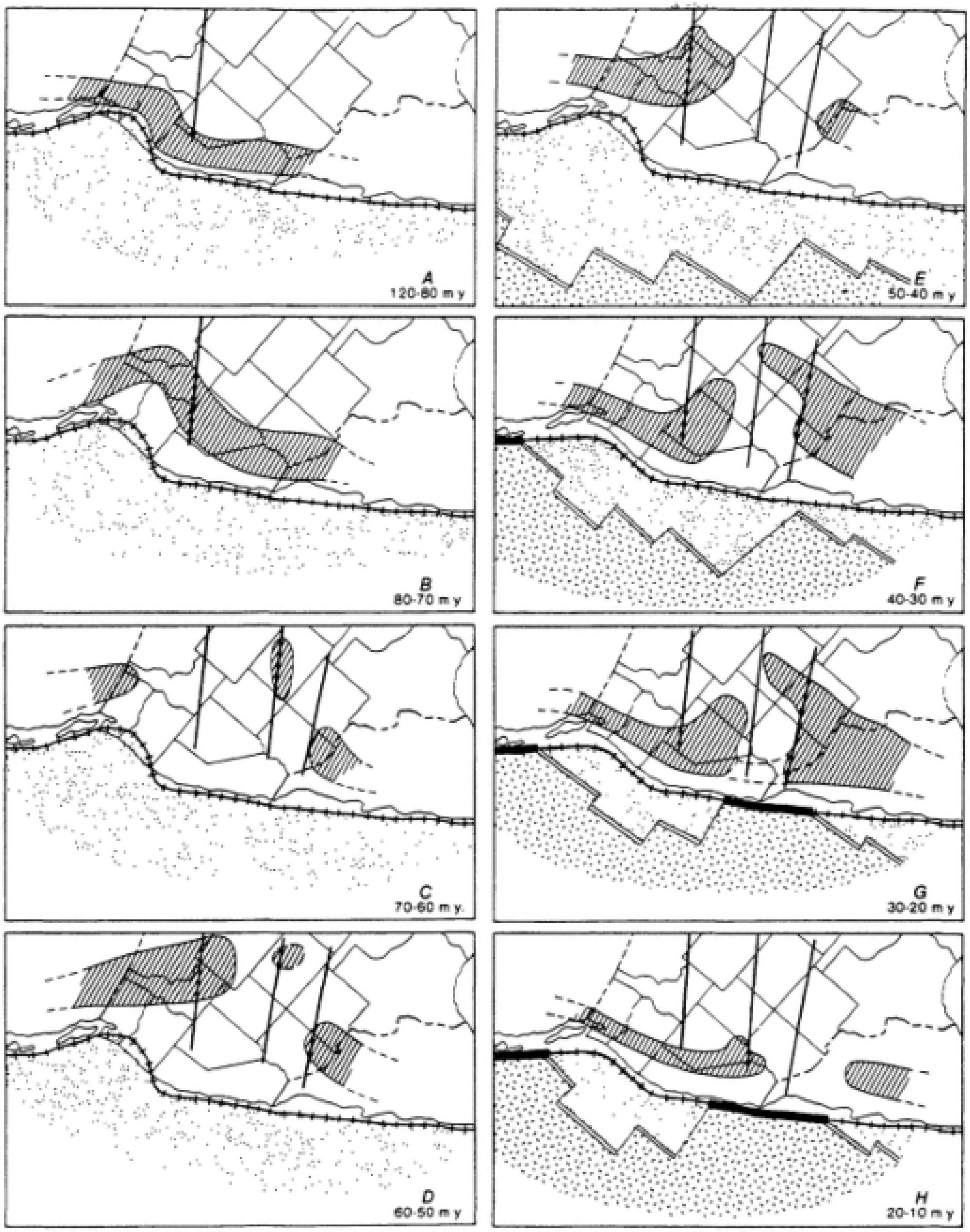


Figure 14.7
Generalized distribution in the western United States of predominantly andesitic volcanic suites, inferred to be related to subduction. Distributions are based on compilations (Lipman et al., 1972; Snyder et al., 1976; Stewart and Carlson, 1976; Armstrong et al., 1977; Cross and Pilger, 1978) and on descriptions of local areas too numerous to cite individually. The base maps and diagrammatic plate geometry are from Atwater (1970) and Atwater and Molnar (1973). No attempt has been made to remove effects of late Cenozoic extensional and rotational deformation, even though such effects are probably 1 arge (Hamilton and Myers, 1966). Northeast-trending lines mark approximate traces of the Snake River-Yellowstone zone, the Colorado mineral belt, and the Springerville-Raton zone.

ing of the subduction zone may also account for the onset of Laramide foreland deformation (Lipman *et al*., 1971; Lowell, 1974; Coney, 1976; Dickinson, 1977).

Early extinction of igneous activity in the northern Rockies—in the very region where activity first migrated eastward—may have been a consequence of further flattening of the subduction system, to such a low angle that igneous activity could no longer be sustained. A modem analog would be the southern Peruvian Andes, where a currently inactive segment of the Andean volcanic chain is underlain by a Benioff zone that dips only about 20º (Stauder, 1975; Barazangi and Isacks, 1976), although the geometry of even this modem plate boundary is currently controversial (James, 1978). The confinement of early Tertiary igneous activity in the southern Rockies to the Colorado mineral belt, despite apparently continuous subduction at the western margin of the plate, suggests the presence in this region of lithosphere that is too thick, cold, or rigid to permit penetration of subduction-related magmas, except along major structural flaws. Any igneous expression of passage of a Kula-Farallon-American plate triple junction, inferred to have swept northward in the early Tertiary (Atwater, 1970), is obscure. The distribution patterns (Figure 14.7C) do suggest, somewhat inconclusively, a possible offset in the downgoing slab in the region between the Colorado mineral belt and the Springerville-Raton zone. The significance of such offsets in controlling distribution of igneous activity in the western United States, inferred also by Cross and Pilger (1978), is more apparent for later Cenozoic volcanism. Examination of compositional data for reliably analyzed suites of Laramide rocks should be useful in evaluating the validity of this possible early Tertiary offset in igneous patterns.

For the interval 60-50 m.y. ago, the most notable changes in the igneous patterns (Figure 14.7D) are a flareup of volcanic activity in the northern Rockies, the virtual extinction of activity in the southern Rockies, and the continued slow, eastward migration of activity in Arizona and New Mexico. In the northern Rockies, volcanic activity broke out widely during this interval (mostly about 55 m.y. ago), including the Lowland Creek volcanics in western Montana, the Absaroka volcanics in northwestern Montana, and the Challis volcanics in central Idaho. Concurrent alkalic activity was also occurring in Montana. The reason for this increased activity after a 10-m.y. to 15-m.y. pause is not clear. Perhaps it reflects the presence of a more steeply dipping subduction system due to a decreased rate of plate convergence. Alternatively, a new segment of subducted slab may have been em-placed with a steeper dip. With Farallon-American plate convergence rates estimated at 10 cm/yr or more for this period (Coney, 1976), newly subducted seafloor could have arrived beneath the eastern Cordillera within 10 m.y. Thus, the geometry of the subduction system could have shifted rapidly, perhaps more rapidly than can be resolved from radiometric dates on middle Tertiary volcanic rocks, especially if changes in the rocks lag several million years after a shift in plate geometry (Christiansen and Lipman, 1972; Snyder *et al*., 1976).

Farther south, in the Colorado sector, little igneous activity seems to have occurred in the interval 60-40 m.y. ago, although a few intrusive dates suggest sparse continued activity. Along the trend of the Springerville-Raton zone, a major northeast-trending discontinuity with the subduction system to the south is evident (Figure 14.7D). In the southwest, dominant activity moved farther eastward, occurring mainly in New Mexico, and tapered off in intensity after about 55 m.y. ago. This pattern is interpreted, following Coney and Reynolds (1977), as primarily reflecting continued flattening of the subduction zone in that sector, although their analysis is simplified by assumption of a constant 150-km depth of magma generation. The Laramide and Tertiary igneous rocks in this region become more alkalic to the east, indicating generation from progressively greater depth as well (Lipman *et al*., 1971).

During the interval 50-40 m.y. ago, changes in the patterns of volcanic activity were relatively minor (Figure 14.7E). In the Northwest, continuing activity in the Challis and Absaroka fields in Idaho and Wyoming began to wind down about 45 m.y. ago, but similar activity flared up farther south in northern Nevada (Stewart *et al*., 1977) and in such parts of northern Utah as the Park City district (Bromfield *et al*., 1977). The nonlinear trend in volcanic activity in the Northwest at the end of this interval seemingly requires a flexure or discontinuity in the subducted slab, changing from relatively steeply dipping in the Northwest to more gently dipping for the Nevada-Utah-southern Idaho sector. The virtual lack of activity in the Colorado sector during this interval suggests the possible continued presence of an even more gently dipping subducted slab under this region, and the east-west-trending zone of activity in the Idaho-Utah-Nevada region may have been localized along a slab flexure (Stewart *et al.,* 1977). Initial development of this slab flexure and associated complex geometry of volcanic activity in the northwestern United States occurred near the site of the future Snake River-Yellowstone zone at about 45-40 m.y. ago, roughly coincident with changed motions of the major tectonic plates, as reflected in the Pacific Basin by the bend in the Emperor-Hawaii seamount chain. The effect of this reorganization was probably to reduce the convergence rate between the American and Farallon plates (Coney, 1972; 1976).

Major changes occurred during the interval 40-30 m.y. ago (Figure 14.7F). Andesitic activity began along the Cascades in western Washington and Oregon, and activity terminated to the east in Idaho and Wyoming. Inception of Cascade volcanism may in part be related to foundering and steepening of the preceding low-angle subduction system, but the position of the trench and subducted slab also probably shifted westward, concurrently with attachment of the Eocene submarine basaltic volcanics of the Coast Ranges of Oregon and Washington (Snavely *et al*., 1968) to the American plate. Concurrently, volcanic activ ity extended farther south in Nevada and Utah (Stewart *et al*., 1977), accentuating the L-shaped pattern of activity in the Northwest. Beginning at about 40 m.y. but only spreading widely by about 35 m.y. ago, renewed igneous activity also broke out in the southern Rocky Mountains in Colorado, New Mexico, and extending as far southeast as western Texas, as well as in the Sierra Madre Occidental in Mexico.

This complex pattern of volcanism suggests the presence of a steep, east-dipping subduction zone under the Cascades, an oblique-trending flexure in the subducting plate under northern Nevada and Utah, and a more gentle, east-dipping zone to the south. Southward migration of the oblique-trending flexure beneath Nevada and Utah in middle Tertiary time (Stewart *et al*., 1977) might have been due to oblique subduction in a southeasterly direction of a descending slab of stable geometry, although reconstructions of plate motions requiring northwesterly convergence to account for disappearance of the Kula plate (Atwater and Molnar, 1973) do not readily fit such a pattern. Also, the flexure appears to have been generated relatively abruptly about 40 m.y. ago in the Pacific Northwest. Perhaps the flexure migrated southward within the descending plate, independent of absolute plate motions, accommodating the change from low-angle to steep subduction.

Apparent offsets along all three lithospheric structural zones at this time (Figure 14.7F) may in part be deceptive, especially for the Springerville-Raton zone. There the boundary is also largely coincident with the southeast margin of the Colorado Plateau, which appears to have been nearly impenetrable to magma through Cenozoic time. An important discontinuity in volcanic patterns appears to have been virtually coincident with the Colorado mineral belt, however, separating the major activity in the southern Rocky Mountains—more than 1000 km from the plate margin—from compositionally similar volcanic fields in Nevada and Utah. This offset is interpreted as marking a major tear in the subducting plate where the change in dip was too great to be accommodated by the oblique flexure under Nevada and Utah. Such a discontinuity in the descending slab seems a preferable interpretation of the complex compositional patterns of middle Tertiary volcanism, for which duplication of subduction zones in an imbricate geometry of subduction was proposed earlier (Lipman *et al*., 1971; 1972).

The period 30-20 m.y. ago was an especially complex time in terms of evolving volcano-tectonic patterns, although the patterns of intermediate-composition volcanic activity changed only slightly from those of the previous interval (Figure 14.7G). Activity continued to spread southward in the Great Basin, and the offset along the Colorado mineral belt is even more conspicuous. This interval also includes the time of initial interaction between the Pacific and American plates, and the 10-m.y. time slice too coarse to portray effectively several significant concurrent changes in volcano-tectonic patterns on the American plate. Basaltic volcanism and associated extensional deformation (Christiansen and Lipman, 1972) were occurring nowhere in the region 30 m.y. ago but were widespread by 20 m.y. ago. Extensional faulting and associated basaltic volcanism apparently began first along the Rio Grande rift system, about 29-26 m.y. ago (Lipman and Mehnert, 1975; Chapin and Seager, 1975), near the interior of the Cordilleran belt, rather than closer to the continental side of the young, growing San Andreas transform.

In the period 29-21 m.y. ago, transitional, relatively alkalic volcanism occurred west of the Rio Grande rift in New Mexico and Arizona; this volcanism was not characterized by well-defined features of either the predominantly andesitic or the fundamentally basaltic types that were significant earlier and later in the region (Elston *et al*., 1976). Concurrently, kimberlitic and undersaturated alkalic basaltic volcanism occurred on the Colorado Plateau. These transitional volcanic rocks seem most reasonably interpreted as related to termination of subduction (Christiansen and Lipman, 1972), either by a rapid retrograde steepening of a subduction system that is no longer being actively regenerated at the plate boundary (Coney and Reynolds, 1977) or by slow foundering of a relatively gently dipping slab, perhaps weakened by the developing "slab window" to the west (Dickinson and Snyder, 1979). In either case, counterflow of asthenospheric mantle would be required to accommodate the sinking slab (Figure 14.6C), a process that could account for initiation of extension and basaltic volcanism at the eastern side of the subduction system, as observed for the Rio Grande rift. Thus, the Rio Grande rift, which also follows the trends of preceding Cenozoic volcanism and tectonism, can be considered an intracontinental analog of oceanic marginal basins, even though the cause of the extension here was termination of the subduction system. An analogous, but less well-documented, association between extensional faulting, andesitic volcanism, and steepening of subduction, may also exist for the Eocene volcanic areas in the Pacific Northwest, as recognized by Davis (see Chapter 8). As considered below, widespread younger extension in the southern Basin-Range province and in the Great Basin also is broadly associated with steepening and destruction of the subduction regimes in these regions.

By the interval 20-10 m.y. ago (Figure 14.7H), andesitic volcanism in the western United States was restricted to a narrow belt near the margin of the American plate and was represented by a southern extension into California and Nevada of the Cascade volcanic arc. During this interval, the southern limit of andesitic volcanism migrated northward, following passage of the plate-boundary triple junction and termination of active subduction (Christiansen and Lipman, 1972) but showing a delay of about 5 m.y. in the transition (Snyder *et al*., 1976). This reduced region of subduction-related volcanism, especially in Nevada and Utah, probably requires some steepening of the subduction zone in this sector. An important problem at this time, for which present plate models do not offer ready explanation, is the episodic nature of subduction-related volcanism (Gilluly, 1973); peaks of activity in the Cascade Range appear to correlate with outbursts on the Columbia Plateau, the western Snake River Plain, the Great Basin, and even with other continental-margin arc systems around the Pacific (McBirney *et al*., 1974).

After about 17 m.y. ago, a pattern of extensional tectonics accompanied by fundamentally basaltic volcanism, characterized by ENE-WSW directions of extension normal to the plate boundary, became established the length of the Great Basin and extending as far north as the Columbia Plateau (Christiansen and McKee, 1978; see Chapters 8 and 9). These tectonic patterns and major associated volcanism, including eruption of the Columbia River basalt and first volcanic activity along the Snake River-Yellowstone trend, are thought to have been initiated, largely in a back-arc environment, when the subduction system of a Juan de Fuca plate reached a critically diminished size as a result of impingement from both south and north by advancing triple junctions. Subsequent evolution of the patterns of extensional deformation and associated fundamentally basaltic volcanism, including nearly concurrent reorientation of the direction of extensional faulting (Zoback and Thompson, 1978), activation of the Snake River-Yellowstone zone (Christiansen and McKee, 1978), and voluminous eruption of the Columbia River basalt group (Swanson *et al*., 1975), probably reflect complex interplay between the enlarging transform boundaries, the diminishing Juan de Fuca plate, and structures of the American plate, discussed elsewhere in this volume (see Chapters 8 and 9).

PERSPECTIVES

Complex space-time-composition patterns of Cenozoic volcanism in the western United States offer information on the composition and structure of the continental lithosphere, as well as provide some of the best available constraints on geometry of past interactions between the American plate and various Pacific plates. These volcanic rocks are most readily interpreted in terms of two broad assemblages: an earlier Cenozoic, predominantly andesitic assemblage inferred to result from geometrically complex processes of plate convergence and a later, fundamentally basaltic assemblage associated with extensional tectonic settings both in regions behind continuing subduction systems —especially where slab dip has steepened—and in regions where subduction has terminated. Adequate age and compositional data are still lacking for many areas, and tight brackets on changes in volcanic type relative to tectonic settings are especially sparse. Many problems also remain in evaluating the primary sources of subduction-related magmas: lower crust, lithospheric upper mantle, asthenosphere; or the subducted oceanic slab.

The distribution of Cenozoic volcanic suites of intermediate composition is interpreted as due to changing geometry of subducted slabs of various Pacific oceanic plates, modified by major structural features of the overlying American plate. Possibly important variables of the subducted slab include changes in dip, physical continuity, temperature, and thickness. An important question is whether the apparent changes in dip of Tertiary subduction systems under the western United States result from deformation of an essentially continuous slab as a result of changing rates and directions of convergence, or whether changes in dip result from detachment of the downgoing slab and rapid re-establishment of the subduction system in a new orientation. Also, how significant is the apparent correlation between increased slab dip and back-arc extension?

Particularly problematic are mechanisms by which major structural features of the American plate may have influenced the distribution of subduction-related volcanism, especially the northeast-trending zones of Precambrian ancestry, "miniplates" such as the Colorado Plateau, and upper-crustal batholiths that may be associated with structurally disturbed zones in the deep lithosphere. If "gravitational anchors" associated with batholith formation extend entirely through the continental lithosphere, as seems likely for the Yellowstone region, can they affect interactions between tectonic plates? Can such perturbations along the northeast-trending structural flaws actually deform gently dipping subducted slabs, or, alternatively, are lithospheric discontinuities reactivated whenever they are appropriately oriented to be utilized by magmas ascending from subduction systems?

Questions such as these should present prime problems for many years for geologists, geophysicists, and geochemists interested in relations between volcanism and continental tectonics.

References

Armstrong, R. L., W. P. Leeman, and H. E. Malde (1975). K-Ar dating, Quaternary and Neogene volcanic rocks of the Snake River Plain, Idaho, *Am. J. Sci. 275*, 225-251.

Armstrong, R. L., W. H. Taubeneck, and P. O. Hales (1977). Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho, *Geol. Soc. Am. Bull. 88*, 397-411.

Atwater, T. (1970). Implications of plate tectonics for the Cenozoic tectonics of western North America, *Geol. Soc. Am. Bull. 81*, 3513-3536.

Atwater, T., and P. Molnar (1973). Relative motions of the Pacific and North American plates deduced from sea-floor spreading in the Atlantic, Indian, and South Pacific oceans, in *Proceedings of the Conference on Tectonic Problems of the San Andreas Fault System*, K. L. Kovach and A. Nur, eds., Stanford U. Publ. Geol. Sci. 13, pp. 136-148.

Barazangi, M., and B. L. Isacks (1976). Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America, *Geology 4*, 686-692.

Bromfield, C. S., A. J. Erickson, Jr., M. A. Haddadin, and H. H. Mehnert (1977). Potassium-argon ages of intrusion, extrusion, and associated ore deposits, Park City mining district, Utah, *Econ. Geol. 72*, 837-848.

Carr, M. J., R. E. Stoiber, and C. L. Drake (1973). Discontinuities in the deep seismic zones under the Japanese arc, *Geol. Soc. Am. Bull. 84*, 2917-2930.
Chapin, C. E., and W. R. Seager (1975). Evolution of the Rio Grande rift in Socorro and Las Cruces areas, in *New Mexico Geological Society Guidebook, 26th Field Conference*, pp. 297-321.
Christiansen, R. L., and P. W. Lipman (1972). Cenozoic volcanism and plate-tectonic evolution of the western United States. Part 2, late Cenozoic, *Proc. Roy. Soc. Lonoon 271*, 249-284.
Christiansen, R. L., and E. H. McKee (1978). Late Cenozoic volcanic and tectonic evolution of the Great Basin and Columbia intermontane region, in Cenozoic Tectonics and Regional Geophysics of the Western Cordillera, R. B. Smith and G. P. Eaton, eds., *Geol. Soc. Am. Mere. 152*, 283-312.
Coney, P. J. (1972). Cordilleran tectonics and North American plate motion, *Am. J. Sci. 272*, 603-628.
Coney, P. J. (1976). Plate tectonics and the Laramide orogeny, *N. Mex. Geol. Soc. Spec. Publ. 6*, 5-10.
Coney, P. J., and S. J. Reynolds (1977). Cordilleran Benioff zones, *Nature 270*, 403-406.
Cross, T. A., and R. H. Pilger (1978). Constraints on absolute motion and plate interaction inferred from Cenozoic igneous activity in the western United States, *Am. J. Sci. 278*, 865-902.
Dickinson, W. R. (1970). Relations of andesitic volcanic chains, granitic batholith belts, and derivative graywacke-arkose facies to the tectonic framework of arc-trench system, *Rev. Geophys. 8*, 813-860.
Dickinson, W. R. (1975). Potash-depth (K-h) relations in continental-margin and intraoceanic magmatic arcs, *Geology 3*, 53-56.
Dickinson, W. R. (1977). Inferred plate tectonic setting of classic Laramide orogeny, *Geol. Soc. Am. Abstr. Programs 9*, 950.
Dickinson, W. R., and W. S. Snyder (1979). Geometry of triple junctions and subducted slabs related to San Andreas transform, *J. Geophys. Res. 84*, 561-572.
Doe, B. R. (1967). The bearing of lead isotopes on the source of granitic magma, *J. Petrol. 8*, 51-83.
Eaton, G. P., R. L. Christiansen, H. M. Iyer, D. R. Mabey, H. R. Blank, Jr., I. Zietz, and M. E. Gettings (1975). Magma beneath Yellowstone National Park, *Science 188*, 787-796.
Elston, W. E., R. C. Rhodes, P. J. Coney, and E.G. Deal (1976). Progress report on the Mogollon Plateau volcanic field, southwestern New Mexico, No. 3—Surface expression of a pluton, *N. Mex. Geol. Soc. Spec. Publ. 5*. 3-28.
Fiske, R. S., O. T. Tobisch, R. W. Kistler, T. W. Stem, and M. Tatsumoto (1977). Minarets caldera, a Cretaceous volcanic center in the Ritter Range pendant, central Sierra Nevada, California, *Geol. Soc. Am. Abstr. Programs 9*, 975.
Gilluly, J. (1955). Geologic contrasts between continents and ocean basins, *Geol. Soc. Am. Spec. Paper 62*, 7-18.
Gilluly, J. (1973). Steady plate motion and episodic orogeny and magmatism, *Geol. Soc. Am. Bull. 84*, 499-514.
Hamilton, W., and W. B. Myers (1966). Cenozoic tectonics of the western United States, *Rev. Geophys. 4*, 509-550.
Hamilton, W., and W. B. Myers (1967). The nature of batholiths, *U.S. Geol. Surv. Prof. Paper 554-C*, 30 pp.
Hamilton, W., and W. B. Myers (1974). The nature of the Boulder batholith of Montana, *Geol. Soc. Am. Bull. 85*, 365-378.
Hotz, P. E. (1971). Plutonic rocks of the Klamath Mountains, California and Oregon, *U.S. Geol. Surv. Prof. Paper 684-B*, 20 pp.
Hurley, P.M., H. Hughes, G. Faure, H. W. Fairbairn, and W. H. Pinson (1962). Radiogenic strontium-87 model of continent formation, *J. Geophys. Res. 67*, 5315-5334.
James, D. E. (1978). Subduction of the Nazca plate beneath central Peru, *Geology 6*, 174-178.
Kistler, R. W., and Z. E. Peterman (1973). Variations in Sr, Rb, K, Na, and initial Sr^{87}/Sr^{86} in Mesozoic granitic rocks and intruded wallrocks in central California, *Geol. Soc. Am. Bull. 84*, 3489-3512.
Klepper, M. R., G. D. Robinson, and H. W. Smedes (1971). On the nature of the Boulder batholith of Montana, *Geol. Soc. Am. Bull. 82*, 1563-1580.
Laughlin, A. W. (1976). Late Cenozoic basaltic volcanism along theJemez zone of New Mexico and Arizona , *Geol. Soc. Am. Abstr. Programs 8*, 598.
Lindgren, W. (1915). The igneous geology of the Cordilleras and its problems, in *Problems of American Geology*, Yale University Press, New Haven, pp. 234-286.
Lipman, P. W., and H. H. Mehnert (1975). Late Cenozoic basaltic volcanism and development of the Rio Grande depression in the southern Rocky Mountains, *Geol. Soc. Am. Mere. 144*, 119-154.
Lipman, P. W., H. J. Prostka, and R. L. Christiansen (1971). Evolving subduction zones in the western United States, as interpreted from igneous rocks, *Science 148*, 821-825.
Lipman, P. W., H. J. Prostka, and R. L. Christiansen (1972). Cenozoic volcanism and plate-tectonic evolution of the western United States: I. Early and middle Cenozoic, *Phil. Trans. Roy. Soc. London 271*, 217-248.
Lipman, P. W., B. R. Doe, C. E. Hedge, and T. A. Steven (1978). Petrologic evolution of the San Juan volcanic field, southwestern Colorado: Pb and Sr isotope evidence, *Geol. Soc. Am. Bull. 89*, 59-82.
Livingston, D. E. (1973). A plate tectonic hypothesis for the genesisof porphyry copper deposits of the southern Basin and Range province , *Earth Planet. Sci. Lett. 20*, 171-179.
Lowell, J. D. (1974). Plate tectonics and foreland basement deformation, *Geology 2*, 275-278.
Luyendyk, B. P. (1970). Dips of downgoing lithospheric plates beneath island arcs, *Geol. Soc. Am. Bull. 81*, 3411-3416.
MacLeod, N. S., G. W. Walker, and E. H. McKee (1975). Geothermal significance of eastward increase in age of upper Cenozoic rhyolitic domes in southeastern Oregon, in *Proceedings 2nd U.N. Symposium on the Development and Use of Geothermal Resources*, Vol. 1, U.S. Government Printing Office, Washington, D.C., pp. 465-474.
Martin, R. F., and A. J. Piwinskii (1972). Magmatism and tectonic setting, *J. Geophys. Res. 77*, 4966-4975.
Mayo, E. B. (1958). Lineament tectonics and some ore districts of the southwest, *Min. Eng. 10*, 1169-1175.
McBirney, A. R., J. F. Sutter, H. R. Naslund, K. G. Sutton, and C. M. White (1974). Episodic volcanism in the central Oregon Cascade Range, *Geology 2*, 585-589.
Minster, J. B., T. H. Jordan, P. Molnar, and E. Haines (1974). Numerical modeling of instantaneous plate tectonics, *J. Roy. Astron. Soc. Can. 36*, 541-576.
Molnar, P., and T. Atwater (1978). The age of subducted oceanic lithosphere: a possible cause of interarc spreading or Cordilleran tectonics, *Geol. Soc. Am. Abstr. Programs 10*, 138.
Morgan, W. J. (1972). Plate motions and deep mantle convection, *Geol. Soc. Am. Mere. 132*, 7-22.
Peterman, Z. E., B. R. Doe, and H. J. Prostka (1970). Lead and strontium isotopes in rocks of the Absaroka volcanic field, Wyoming, *Contrib. Mineral. Petrol. 27*, 121-130.
Schweickert, R. A. (1976). Shallow-level plutonic complexes in

the eastern Sierra Nevada, California, and their tectonic implications, *Geol. Soc. Am. Spec. Paper 176*, 58 pp.
Shaw, H. R., and E. D. Jackson (1973). Linear island chains in the Pacific: result of thermal plumes or gravitational anchors?, *J. Geophys. Res. 78*, 8634-8652.
Shoemaker, E. M., R. L. Squires, and M. J. Abrams (1974). The Bright Angel and Mesa Butte Fault systems of northern Arizona, in *Geology of Northern Arizona: Part I, Regional Studies*, Karlstrom *et al*., eds., Geological Society of America, Cordilleran Section Meeting, Flagstaff, Ariz., pp. 355-391.
Silver, L. T., T. O. Early, and T. H. Anderson (1975). Petrological, geochemical, and geochronological asymmetries of the Peninsular Ranges batholith, *Geol. Soc. Am. Abstr. Programs 7*, 375.
Simpson, R. W., and A. Cox (1977). Paleomagnetic evidence for tectonic rotation of the Oregon Coast Range, *Geology* 5, 585-589.
Shayely, P. D., Jr., N. S. MacLeod, and H. C. Wagner (1968). Tholeiitic and alkalic basalts of the Eocene Siletz River volcanics, *Am. J. Sci. 266*, 454-481.
Snyder, W. S., W. R. Dickinson, and M. L. Silberman (1976). Tectonic implications of space-time patterns of Cenozoic magmatism in the western United States, *Earth Planet. Sci. Lett. 32*, 91-106.
Stauder, W. (1975). Subduction of the Nazca plate under Peru, as evidenced by focal mechanisms and by seismicity, *J. Geophys. Res. 80*, 1053-1064.
Steven, T. A., and P. W. Lipman (1976). Calderas of the San Juan volcanic field, southwestern Colorado, *U.S. Geol. Surv. Prof. Paper 958*, 35 pp.
Stewart, J. H., and J. E. Carlson (1976). Cenozoic rocks of Nevada, *Nev. Bur. Mines Geol. Map 52*.
Stewart, J. H., W. J. Moore, and I. Zietz (1977). East-west patterns of Cenozoic igneous rocks, aeromagnetic anomalies, and mineral deposits, Nevada and Utah, *Geol. Soc. Am. Bull. 88*, 67-77.
Stoiber, R. E., and M. J. Carr (1974). Quaternary volcanic and tectonic segmentation of Central America, *Bull. Volcanol. 37*, 304-325.
Suppe, J., C. Powell, and R. Berry (1975). Regional topography, seismicity, Quaternary volcanism, and the present-day tectonics of the western United States, *Am. J. Sci. 275A*, 397-436.
Swanson, D. A, T. L. Wright, and R. T. Heltz (1975). Linear vent systems and estimated rates of magma production and eruption for the Yakima basalt on the Columbia Plateau, *Am. J. Sci. 275*, 877-905.
Tweto, O., and P. Sims (1963). Precambrian ancestry of the Colorado Mineral Belt, *Geol. Soc. Am. Bull. 74*, 991-1014.
Warner, L. A. (1978). The Colorado lineament: a middle Precambrian wrench fault system, *Geol. Soc. Am. Bull. 89*, 161-171.
Wilson, J. T. (1965). Evidence from oceanic islands suggesting movement in the earth, *Phil. Trans. Roy. Soc. London 258*, 145-165.
Zartman, R. E. (1974). Lead isotope provinces in the Cordillera of the western United States and their geologic significance, *Econ. Geol. 69*, 792-805.
Zoback, M. L., and G. A. Thompson (1978). Basin and Range rifting in northern Nevada: clues from a mid-Miocene rift and its subsequent offsets, *Geology 6*, 111-116.

15

The Shape of North America during the Precambrian

William R. Muehlberger
The University of Texas

INTRODUCTION

Historically, the geological evolution of North America has been the subject of studies since they were first made on this continent. This largely led to the development of geosynclinal theory, which was rapidly applied to thick sedimentary sequences worldwide. In the first half of this century, continental drift was not widely accepted in North America, but the evidence suggested continental accretion. Continental accretion of North America was supported by the early isotopic studies of Precambrian rocks and the apparent progressive younging of Phanerozoic orogenic belts toward the continental margins. Continuing work showed the presence of older rocks in the cores of hinterlands of orogenic belts and thus led to the recognition that continental crust, which had been stabilized in some earlier event, could be reactivated, have extensive sedimentary belts (geosynclines) form within it, be deformed, metamorphosed, intruded, and once again be a part of the stable interior of a continent.

Currently being debated is whether most continental crust was formed early in geological time or whether it has grown progressively through time. Engel and Engel (1964) presented a good summary of these opposing views. Their paper supports the latter view of continental accretion. This paper supports the early formation of much if not most of continental crust and continued accretion, but at a much slower rate. Hargraves (1976) presents a lucid discussion of the progressive growth of continental crust throughout geological time.

The advent of plate tectonics has revolutionized geological thinking. The three end members—rifts, transforms, and collisions (subduction)—with their characteristic associations of rock types provide new insights into the interpretation of continental rocks older than the cycle of seafloor spreading preserved in the ocean floor. These concepts have been widely applied to the Phanerozoic with an ever-increasing recognition and understanding of the wide variety of events that can happen during the Wilson cycle of continental rifting, drifting, closing, and suturing (e.g., Burke *et al*., 1977; Dewey, 1977).

Structural evolutionary rates must be gradually slowing down because of an overall cooling of the earth through geological time. Thus, convective "rollover," a character istic feature of oceanic lithospheric generation, cooling, and subduction should be slower today relative to earlier episodes. This implies that crustal generation should also be slower now relative to the early Precambrian.

Figures 15.1-15.4* show, in a summary form, the progressive stabilization and "growth" of the North American craton during four intervals of the Precambrian. How far back in geological time can plate-tectonic processes be inferred? Late Paleozoic? Grenville? Hudsonian/Penokean? Archean? If plate-tectonics processes can be invoked in the Precambrian we have yet to determine most of the interrelationships of active and passive margins, what was rifted off, what was accreted, or what was involved in major transform margins.

Three divisions of geological time are commonly used in discussing plate tectonics and its applicability to tectonic features: Archean, Proterozoic, and Phanerozoic (Anhaeusser, 1975; Burke *et al*., 1977; Sutton, 1973; Windley, 1977). These are diachronous in space and time but roughly correspond to the standard geological time scale. They represent major intervals of differing response of "stable" cratons to orogenic deformation.

ARCHEAN

The Archean had a style that was unique and that was essentially terminated about 2.5 billion years (b.y.) ago in North America, earlier in some continents, later in others. The development of extensive gray gneisses, terranes of soda-rich granitic composition, is unique to this period. These probably represent the early formation of continental crust by processes that are still poorly understood (Glikson, 1976; Goldich and Hedge, 1974; Goodwin, 1974a, 1974b; Moorbath *et al*., 1972; Myers, 1976). The study of lunar samples has accelerated attempts at unraveling the earliest history of earth; comparative plane-taw studies are an obvious important topic needing major research to fill in the gap between the terrestrial rock record and the origin of the earth.

The abundance of greenstone-graywacke belts is another association that is nearly unique to the Archean. The peridotite/komatiite basal unit is also nearly unique—it requires mantle temperatures well above those available today for magma generation [two early Proterozoic examples (Arndt *et al*., 1977) and a possible Ordovician example (Upadhyay, 1978) have been identified thus far]. Green (1975) also shows that with his preferred steep isotherm model for the Archean, eclogite could not form and subduction would not occur; thus, the basaltic oceanic crust would be scraped off against primitive sialic nuclei. The absence of Archean alkaline complexes, carbonatites, or kimberlites, all characteristic of stable cratons (Windley, 1977) also attest to the mobility of the Archean crust.

Most workers do not invoke plate-tectonic regimes during the formation of the Archean high-grade gneiss terranes or the low-grade greenstone belts. The greenstone belts have been argued to be sufficiently distinctive (Anhaeusser *et al*., 1969; Douglas and Price, 1972; Engel and Kelm, 1972) that there is no comparable environment in present-day orogenic belts. The many theories that have attempted to explain the evolution of greenstone belts are summarized by Windley (1977, pp. 4.5-58). Structural style is unique and dominated by a developing gravitational instability as mafic volcanic piles and adjacent region were intruded by soda-rich melts and migmatized. These piles subsided into a thin lithosphere with folding occurring as the volcanic piles slid toward the subsiding flysch basins. These then were injected by potash-rich granites with later stabilization (cooling) of the developing continental mass. Vertical motions appear to dominate the tectonics of this environment, although initial undulations have been interpreted to be the result of the flexing of thin lithosphere plate on the overriding side of a subduction zone or back-arc spreading (Drury, 1977; Fyson, 1978).

Goodwin (1974b) proposed that the initial crustal units developed around an early Archean paleoplume centered in Hudson Bay. These units aggregated to form a large craton by 1.7 b.y., which underwent periods of epeirogenic uplift, rifting, and downwarping until it began a continuing period of downwarping during the Phanerozoic because of movement of the craton off of its parent plume. This thesis, as others (e.g., Hurley and Rand, 1969), supports the concept that widespread fragmentation and dispersal of continental crust, so obvious in Cenozoic plate-tectonic analyses, could not have occurred in the Precambrian. However, these concepts do not preclude large-scale horizontal movements of an entire shield nor marginal accretions throughout this interval of time.

Archean rocks in the United States (Figure 15.1) are limited to two major regions: Wyoming and vicinity and the southern extension of the Canadian Shield into the northern tier of states west of Lake Michigan.

Structural trends in the greenstone and gneiss belts of Archean rocks of Wyoming are north to northeast for the earlier Archean deformation and north to northwest for the later (Houston, 1971). Much of the Precambrian in Wyoming has yet to be mapped, and the possible effects of Laramide rotation have not been assessed. From a structural geologist's point of view, it is intriguing that Wyoming, the only region with Archean basement, is the home of the Wyoming upthrust province, a unique structural style and trends of Laramide foreland deformation.

Northern Minnesota and the subcrop extension into the Dakotas constitute the southwestern extension of the Superior province, the largest block of unreworked Archean crust in North America. The granitic gneiss un

* Principal sources for all figures in this paper are Baer, 1970; Bayley and Muehlberger, 1968; Drummond, 1974; Goodwin, 1974a; Jackson and Taylor, 1972; King and Beikman, 1974; Lidiak, 1971; Lidiak and Zietz, 1976; Muehlberger *et al*., 1967; Price and Douglas, 1972; Stewart, 1976; Tectonic map of Canada, 1969.

derlying the greenstone-graywacke belts across the southern third of the province is about 3.0 b.y. old, and each greenstone belt evolved rapidly—about 70 million years (m.y.) between the start of greenstone deposition and the end of granitic intrusions (Krogh and Davis, 1971; Krogh *et al*., 1974, 1975). Nunes *et al*. (1978) and Nunes and Thurston (1978), however, have shown that the volcanic rocks of the Abitibi greenstone belt were erupted in about 25 m.y., whereas the Uchi Lake greenstone belt spanned 220 m.y. The cooling and stabilization of the Superior province occurred near 2.5 b.y. as shown by the extensive K-Ar dating program of the Geological Survey of Canada (Stockwell, 1961).

Figure 15.1
Area of North America known to be underlain by rocks 2.5 b.y. old or older. Principal areas (provinces) are named Slave, Nain, Superior, and Wyoming. Short lines and dots are outcrop belts or wells where rocks of this interval have been identified; most of these areas have been involved in orogenic events shown on Figure 15.2.

In contrast, the Archean of southern Minnesota, Wisconsin, and northern Michigan has been involved in younger orogenic events and was remetamorphosed in those events. The Precambrian of the Minnesota Valley contains rocks older than any yet recognized from the Superior province (3.8 b.y.; Goldich and Hedge, 1974).

The Great Lakes region has been an area of intense study for many decades, initially to analyze the mineral deposits but gradually developing into a study of the broader geological environments and the use of isotopic determinations in unraveling the geological history of the region. Extensive cover of glacial debris and Phanerozoic rocks around the southern and western sides has made conventional geological mapping difficult. The availability of regional magnetic and gravity coverage speeded up these studies significantly and has made it possible to extrapolate successfully from the limited outcrop data (see Chapter 11).

These studies (see Sims, 1976; Morey and Sims, 1976, for a summary) have demonstrated that the Archean is divisible into two major terranes. One is an older ensialic gneiss terrane that extends across southern and central Minnesota, northern Wisconsin. and the Upper Peninsula of Michigan with ages as great as 3.8 b.y. These rocks were then sutured about 2.7-2.6 b.y. ago to an ensimatic greenstone belt that may have been formed in island-arc or continental borderland environments with oldest ages about 3.0 b.y. (Superior province of the Canadian Shield) but ages ranging from about 2.8-2.7 b.y. for the green-stone units and their intensive granitic plutons.

EARLY PROTEROZOIC

Sims (1976) shows that the Lake Superior Proterozoic basin overlies an Archean suture and that the younger greenstone terrane is the stable block during this interval and the major source of sedimentary materials. This basin is compared with the intracratonic Labrador Trough (Dimroth *et al*., 1970) and shown to be analogous in nearly all depositional and tectonic events including timing, insofar as they have been determined.

Thus the ancient gneiss terrane that rims the Archean greenstone belt on the south and east acted as the mobile basement until it was finally stabilized about 1.7 b.y. ago to form a larger cratonic nucleus. Some of this belt was again reactivated and now forms part of the Grenville province.

Figure 15.2 shows this boundary and its continuation across the midcontinent as derived from Muehlberger *et al*. (1967), Bayley and Muehlberger (1968), and Lidiak (1971) and a southwestern continuation (dotted) as proposed by Warner (1978). Warner (1978) proposed that it is part of the Colorado lineament, a middle Precambrian (Penokean orogeny: 2.0-1.7 b.y.) wrench fault (left-lateral?) system that resembles the San Andreas type. Detailed geological mapping of the Mullen Creek-Nash Fork shear zone in southeastern Wyoming (Houston *et al*., 1968) shows that it must have been a continental margin, thus a possible Proterozoic plate boundary (Hills *et al*., 1975). Peterman and Hildreth (1977) demonstrated a Proterozoic overprint using K-Ar ages on the Archean rocks over a band up to 150 km wide north of the shear zone in Wyoming suggesting a collision boundary. The Colorado lineament is a major discontinuity and deserves consider able effort to determine its true nature and regional significance.

Figure 15.2
Area of North America underlain by rocks 1.7 b.y. old or older. Archean provinces of Figure 15.1 are stable cratons of this episode. Dashed lines show location and trend of fold belts active during the interval 1.7-2.5 b.y. ago. Southern and western dashed-line boundaries are limit of known rocks oft his time interval. Western boundary in United States based on Rb-Sr isotopic data (Kistler and Peterman, 1978). Line of triangles in Wyoming craton shows northern limit of middle Proterozoic K-At age overprint. Colorado Lineament forms prominent discontinuity from Lake Superior southwest to western Arizona.

Figure 15.2 also shows the Proterozoic fold belts that transect the Archean of the Canadian Shield and that are now part of the Churchill province that was consolidated with the Archean elements about 1.7 b.y. ago. In contrast to the Archean, the Proterozoic initiated the era that still continues, of rigid continental platforms, cratons that stood above sea level and thus produced the first recognizable continental margin, rifled margins and associated aulacogens, and continental-margin geosynclines. The Coronation geosyncline and Athapuscow aulacogen (Hoffman, 1973) that flank the Slave province of the northwestern Canadian Shield are among the earliest recognized in the Proterozoic.

Intracratonic basins (Dimroth *et al*., 1970) or small ocean basins (Kearey, 1976), such as the Labrador geosyncline, are dominant interpretations of the Proterozoic belts that are involved with reactivated Archean crust of the Churchill province.

Wynne-Edwards (1976) has proposed a model to explain the ensialic basins seemingly typical of Proterozoic orogenesis. It requires the thinning of hot ductile sialic crust over mantle upwarps. These then become sites of sedimentary basins that close by migration of the upwarp. Broad zones of shear, rather than the present transform faults, divide the spreading areas into segments and are spatially related to base-metal deposits. The lines of plutonic rocks such as the anorthosite massifs in the Grenville province provide tracks for relative movement vectors. Martin and Porada (1977), using the same African examples as does Wynne-Edwards, proposed instead that gravitational instabilities caused the development and subsequent closing of a multiple aulacogen to explain Proterozoic ensialic orogens. Kroner (1977) expanded upon their concept and showed how linear zones of weakness may develop ensialic geosynclines from graben systems via aulacogens. He further proposed that the horizontal mobility of large crustal plates increased through geological time to develop the wide opening and closing of ocean basins characteristic of the Phanerozoic.

Gibb (1975; 1977) recognizes a set of major faults that extends westward from the northern termination of the Kapuskasing Gneiss Belt in the southernmost James Bay resulting from the suturing of the Superior and Churchill plates. He depicts a northward-moving Superior plate suturing progressively westward from James Bay and subducting the intervening oceanic lithosphere. He interprets this collision, following the Tibetan Plateau analog (Dewey and Burke, 1973), to have reactivated the Churchill basement. He points to the Wollaston Lake fold belt, 450 km to the northwest, whose shape mimics the curvature of the suture boundary, as the possible limit of penetrative deformation and crustal reactivation. However, this does not explain why all the metasedimentary belts of the Churchill province mimic the suture boundary nor why the entire province has a uniform K-Ar age. His concept should be expanded to include the entire Churchill province to be the result of the Superior province collision. Seyfert (1978) interprets paleomagnetic apparent polar wander curves to show that a North America-Gondwanaland collision at 1.85 b.y. caused the Penokean orogeny. He further suggests that Europe joined North America at this time; this might resolve the apparent paradox of having a collision margin completely circumscribing the Superior province and furnishes an additional collision direction to assist in remobilizing the Churchill province. Cavanaugh and Seyfert (1977) suggest that the Slave province collided with the Superior province about 1.75 b.y. ago. Their proposed suture extends from between the Foxe and Committee fold belts of Baffin Island southwest across the shield. Another proposed Proterozoic suture, based on the recognition of a zone of high electrical conductivity in the buried shield (Camfield and Gough, 1976), extends southward from the Wollaston fold belt to southeastern Wyoming.

The curvature of the belt that surrounds the Superior province suggests a flexibility (ductility?) of lithosphere that is greater than present-day plate tectonics will allow. However, the tight curve around the Ungava Peninsula of northern Quebec has the *same* radius of curvature as the present-day Banda Arc of the East Indies, the eastern Caribbean arc, or Scotia Arc between South America and Antarctica. An important difference for these modern examples, however, is that each has oceanic crust inside the arc rather than circumscribing a continental craton.

Important questions yet unresolved include: how broad was the ocean or seaways in which these fold belts were deposited? Paleomagnetic control on rocks of this age (2.5-1.7 b.y.) permits oceans limited to widths no greater than 500-1000 km (about the width of Baffin Bay) that must close so that their original relative positions are maintained (McElhinney and McWilliams, 1977). What was the sequence of formation? Are they successive events or plasterings onto the Superior margin, or were they ductile extension zones in a larger Archean continent that subsided, filled with sediments, and were later compressed? How can we best account for the circum-Superior collision without having it reactivated as well? How many of these proposed sutures have supporting geological evidence?

LATE PROTEROZOIC

The arcuate trends, extrapolated from the Canadian Shield along linear gravity anomalies and the Black Hills outcrop belt under the northern plains, appear to be truncated along the Colorado lineament. A southeast trend continues faintly across Nebraska and into northwestern Missouri, but geological confirmation is lacking. If it continues, then it is evidence for the Penokean orogeny being earlier than the Churchill deformations or it marks the western margin of the Archean crustal block of the Great Lakes region.

The southern limit of North America at about 1.7 b.y. appears to be as shown in Figure 15.3, although the steps in its evolution are only now being deciphered. Isotopic data in Colorado and southern Wyoming indicate that no Archean crust exists beneath the exposed Precambrian rocks (Hills and Armstrong, 1974); thus crust evolved since the Archean. How many lithological/structural belts are present in this band or their sequence of development is unknown for most of this region.

Proterozoic greenstone belts are recognized at many places in Arizona, Colorado, and New Mexico (Anderson and Silver, 1976; Barker, 1969; Robertson *et al*., 1978). Tectonic foundering of a continental margin in northern New Mexico and southwestern Colorado has been proposed by Barker *et al*. (1976) to explain the southeast-trending belt of trondhjemites and the nearly contemporaneous rhyolite-quartzite terrane that lies to the southwest. Sedimentary structures (Barrett and Kirschner, 1979; Montgomery, 1953) in the quartzites and intercalated schists, which lie southwest of the rhyolite terrane, show them to be continental-margin deposits. Field and isotopic studies thus far have not definitely established the time sequence, although considerable progress is being made (Denison *et al*., in press). From a resource point of view, it is important to know the sequence and environment of deposition of each belt so that intelligent mineral exploration programs can be planned.

Figure 15.3
Area of North America underlain by rocks 1.3 b.y. old or older. Craton now includes all of the Canadian Shield, except for its eastern border. The southeasternmost line marks probable southern limit of rocks 1.3 b.y. old or older. The central line is the southern limit of rocks 1.7 b.y. old or older. Between these lines is the anorogenic pluton belt of Silver et al. (1977) and Emslie (1977); plutons shown as dots at location of dated sample or dated pluton. The northern line marks the northern limit of intrusive or metamorphic activity giving ages in the 1.3- to 1.7-b.y. range. Western boundary is same as Figure 15.2. Cross-hatched areas are rhyolite/granite terranes; dotted areas are Sioux quartzite.

Figure 15.3 shows that rocks in the interval between 1.7 b.y. and about 1.35 b.y. ago dominate the midcontinent region of the United States. This division is a relic from our 1967 paper (Muehlberger *et al*., 1967), but because of more recent studies some of the distinctions made in that paper have blurred; others have sharpened. We used it to mark the end of the Elsonian orogeny (Stockwell, 1961); it now marks the end of the anorogenic pluton episode

(Emslie, 1977; Bickford and Van Schmus, 1978; Silver *et al*., 1977). The northern belt of Figure 15.3 consists of a 1.7-b.y. basement intruded by anorogenic plutons about 1.5-1.4 b.y. The belt south of the heavy dashed line contains no rocks known to be older than 1.4 b.y.

Much of the southern belt is overlain by younger Precambrian rhyolite and associated granite sills or Precambrian sedimentary units, which mask the older crustal geology (Muehlberger *et al*., 1967); thus evidence for an older basement, if any, has yet to be determined. Rhyolite-granite terranes become progressively younger from northeast to southwest (Bickford and Van Schmus, 1978; Denison *et al*., 1977). This younging of granite-rhyolite terranes continues on the next age interval map (Figure 15.4) across the Texas Panhandle and into the E1 Paso area. A prominent exception to the southwestward younging is the early Cambrian magmatism (Ham *et al*., 1964) associated with the development of the Southern Oklahoma Paleozoic aulacogen.

Rifting events have been proposed for the origin of the 1.4-b.y. to 1.5-b.y. anorogenic pluton belt, the 1.2-b.y. Mackenzie dike swarm (shown schematically on Figure 15.4), and the Coppermine basalts on the northwest Canadian Shield. Apparently slightly younger but possibly related to the Mackenzie dikes is the extensive intracontinental rift system of the Keweenawan. Sims (1976) has shown that the Lake Superior segment of Keweenawan rifling is bounded on the south by pre-existing continental faults that had substantial right-lateral movement before Keweenawan time. Sawkins (1976) has shown that this rifling event was a major period of copper mineralization. This rifting was soon terminated by the Grenville collision as indicated by the near coincidence of isotopic ages, and the truncation of Keweenawan marks the first opening(?). The location of the suture is obscured by late Paleozoic activity of the Appalachian belt. King and Zietz (1978) describe the New York-Alabama lineament that lies beneath the Appalachian basin and propose that it is a major strike-slip fault (possibly a suture) analogous to the currently active Altyn Tagh fault of Tibet, resulting from the India-Asia collision (Molnar and Tapponnier, 1975).

Across Texas we have evidence for suturing at Grenville time. The west-trending overthrust belt of the Van Horn region (King and Flawn, 1953) and the serpentinites of the southern Llano uplift in central Texas (Barnes, 1946; Garrison, 1978) both show movement of material northward (using present geographic coordinates) onto the North American continent. The suture has been proposed as the closing of an aulacogen or small ocean (Garrison and Ramirez-Ramirez, 1978; Sengor* and Butler, 1977).

This proposed suture is part of the zone that several investigators include in the Texas lineament, a major zone of recurrent tectonic activity that extends across the southwestern United States. The origin of the Texas lineament is shrouded in mystery (oldest demonstrated offsets are about 1.4 b.y.; Swan, 1975), but that it has been the site of recurrent tectonic activity is a fact (Wiley and Muehlberger, 1971). Thus it and the Colorado lineament are similar in style, but the southern zone is younger. Could these be the sites of oblique collision, zones that produce such a pervasive grain and fracture system that they never properly weld to pre-existing shield and thus are continuing zones of activity? The Grenville-Appalachian belt appears to be a younger and more extensive analog. The paucity of information on the Precambrian of Mexico

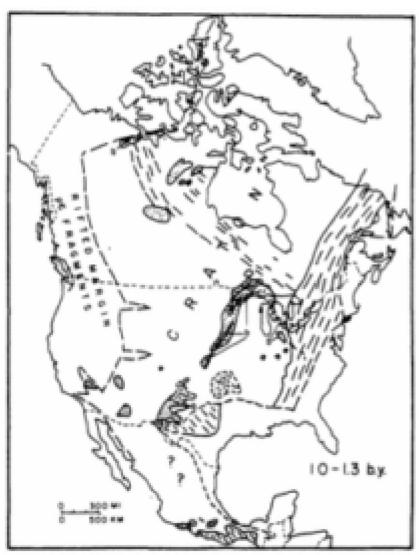


Figure 15.4
Area of North America at about 1.0 b.y. old; showing major tectonic and igneous events between 1.3 b.y. and 1.0 b.y. of age. The western margin (essentially that shown in Figures 15.2 and 15.3) is now seen to be a rifted margin (former western limit is not yet identified) with two major re-entrants (aulacogens). Fragments of Precambrian continental crust have been recognized in the northwestern part of the Cordilleran belt. Southwestern margin is distorted by late Phanerozoic transform tectonics. Dotted areas are sedimentary rock of Precambrian age; cross-hatched areas, rhyolite-granite terranes; b, basalt; dashed lines inside a linear boundary, Keweenawan basaltgabbro; dashed lines trending northwest across Canada, Mackenzie dike swarm; dashed zone along eastern North America, Texas, and southern Mexico are results of the Grenville collision. Black dot in Colorado is Pike's Peak granite (~1.1 b.y. old). Peripheral to this craton are the successive latest Precambrian and Phanerozoic sedimentary wedges.

because of extensive cover of Mesozoic and younger rocks leaves all reassemblies speculative; however, the "Grenville front" shown across west Texas is based on geophysical, petrographic, and isotopic data as described in Muehlberger *et al*. (1967).

The rifted western and northwestern margins of North America are reasonably well located because of the presence of the miogeoclinal wedge of Belt rocks and their equivalents. Where the margin was in earlier times is speculative. The recognition of Precambrian blocks in the western Cordillera of Canada and Alaska and extensive strike-slip displacements make reconstruction a major problem. Stewart (1976) has proposed a late Precambrian age (about 850 m.y.) for the rifting that made this region. The Belt and Uinta embayments (aulacogens?—Burke and Dewey, 1973) were filling early in Belt time, and thus a margin was in existence by at least 1.35 b.y. ago. Sears and Price (1978) present evidence that the western and southwestern boundaries of North America (Cordilleran Belt margin and Texas lineament) are pre-Belt rift margins and that the matching segments can be found in the Siberian cratons. Seyfert (1978) from paleomagnetic data suggests that Asia joined the west coast of North America between 1.4 b.y. and 1.15 b.y. ago. The absence of a western source of Belt and Windermere strata (Maxwell, 1974) suggests that Asia may only have been close to North America. Isotopic identification of continental crust in western United States places a western limit on it that essentially coincides with the latest Precambrian sedimentary wedges (Kistler and Peterman, 1978; Stewart, 1976). On the other hand, Badham (1978a) proposes that the western margin of North America has been both an active and passive margin but was never destroyed by rifting (the Athapuscow aulacogen is only a persistent fault zone; Badham, 1978b) or collision and thus has progressively grown westward.

Late Precambrian rifts, aulacogens, and other extensional phenomena (alkalic magmatism, anorogenic plutons) have been proposed for many regions. Some of these same belts are zones of seismicity today, for example, St. Lawrence rift system (Kumarapeli, 1978), Reelfoot rift (Ervin and McGinnis, 1975), and southern Oklahoma aulacogen (Denison, 1978; Wickham, 1978). Each of these structures has its origins in the Precambrian, although most of their documented activity is in the Phanerozoic. An understanding of how old tectonic lines can be reactivated will furnish significant data on paleostress orientations (Sykes, 1978).

SUMMARY

This brief description of the evolution of North America has attempted to show that we have very little detailed knowledge of how, where, or when successive belts of rocks were added or were rifted off or were reactivated (continental collisions?) or, for that matter, how well or even whether plate tectonics works in the Precambrian. Every aspect of the evolution of continents is in a state of active debate, research, and ferment (see Chapter 2).

Broad trends, however, are visible: Archean (2.5 b.y.); early Proterozoic (2.5-1.7 b.y.) with stable cratons and intracratonic mobile belts; middle Proterozoic (1.7-1.2 b.y.) with anorogenic plutons (incipient rifting) and widespread rhyolitic volcanism across the central midcontinent region; late Proterozoic to Cambrian (1.2-0.6 b.y.) with rifting and collision events of Phanerozoic style; and the Phanerozoic—the recognized realm of plate tectonics. Each of these intervals marks a period of fairly uniform structural style, evolving from more ductile to more rigid deformation styles that reflect an increasing rigidity and areal growth of the lithosphere and, presumably, a concurrent thickening (Jordan, 1975).

References

Anderson, C. A., and L. T. Silver (1976). Yavapai Series—a greenstone belt, *Ariz. Geol. Soc. Digest X*, 13-26.
Anhaeusser, C. R. (1975). Precambrian tectonic environments, *Ann. Rev. Earth Planet. Sci.*, 3 31-53.
Anhaeusser, C. R., R. Mason, M. J. Viljoen, and R. P. Viljoen (1969). A reappraisal of some aspects of Precambrian shield geology, *Geol. Soc. Am. Bull. 80*, 2175-2200.
Arndt, N. T., A. J. Naldrett, and D. R. Pyke (1977). Komatiitic and iron-rich tholeiitic lavas of Munro township, northeast Ontario, *J. Petrol. 18*, 319-369.
Badham, J. P. N. (1978a). Has there been an oceanic margin to western North America since Archean time?, *Geology 6*, 621-625.
Badham, J .P. N. (1978b). The early history and tectonic significance of the East Arm graben, Great Slave Lake, Canada, *Tectonophysics 45*, 201-215.
Baer, A. J., ed. (1970). Symposium on basins and geosynclines of the Canadian Shield, *Geol. Surv. Can. Paper 70-40*, 265 pp.
Barker, F. (1969). Precambrian geology of the Needle Mountains, southwestern Colorado, *U.S. Geol. Surv. Prof. Paper 644-A*, 35 pp.
Barker, F., J. G. Arth, Z. E. Peterman, and I. Friedman (1976). The 1.7 to 1.8 b.y.-old trondhjemites of Southwestern Colorado and northern New Mexico; geochemistry and depths of genesis, *Geol. Soc. Am. Bull. 87*, 189-198.
Barnes, V. E. (1946). Soapstone and serpentine in the Central Mineral Region of Texas, in *Texas Mineral Resources*, Bureau of Economic Geology, U. of Texas Publ. 4301, pp. 55-91.
Barrett, M., and C. Kirschner (1979). Depositional environments of the Precambrian Ortega and Rinconada formations, Picuris Mountains, northcentral New Mexico, in *New Mexico Geological Society Guidebook, Santa Fe Country*, pp. 121-126.
Bayley, R. W., and W. R. Muehlberger (1968). Basement rock map of the United States, exclusive of Alaska and Hawaii, scale 1:2,500,000, U.S. Geological Survey.
Bickford, M. E., and W. R. Van Schmus (1978). Chronology of middle and late Precambrian crest-forming events in the mid-continent region of North America, *Geol. Soc. Am. Abstr. Programs 10*, 2.
Burke, K., and J. F. Dewey (1973). Plume-generated triple junc

tions: key indicators in applying plate tectonics to old rocks, *J. Geol. 81*,406-433.
Burke, K., J. F. Dewey, and W. S. F. Kidd (1977). World distribution of sutures—the sites of former oceans, *Tectonophysics 40*, 69-99.
Camfield, P. A., and D. I. Gough (1976). A possible Proterozoic plate boundary in North America, *Eos Trans. Am. Geophys. Union 57*, 1003.
Cavanaugh, M. D., and C. K. Seyfert (1977). Apparent polar wander paths and the joining of the Superior and Slave provinces during early Proterozoic time, *Geology 5*, 207-211.
Denison, R. E. (1978). Basement rocks in the Arbuckle Mountains, in *Structural Style of the Arbuckle Region*, J. Wickham and R. E. Denison, field trip leaders, Geological Society of America, South-Central Sect., Field Trip 3, pp. 42-46.
Denison, R. E., E. G. Lidiak, M. E. Bickford, and E. B. Kisvarsanyi (1977). Precambrian geology and geochronology of the central interior province, North America, *Geol. Soc. Am. Abstr. Programs 9*, 947-948.
Denison, R. E., E.G. Lidiak, M. E. Bickford, and E. B. Kisvarsanyi (in press). Geology and geochronology of Precambrian rocks in the central interior region of the United States, *Econ. Geol*.
Dewey, J. F. (1977). Suture zone complexities: a review, *Tectonophysics 40*, 53-67.
Dewey, J. F., and K. Burke (1973). Tibetan, Variscan and Precambrian basement reactivation: products of continental collision, *J. Geol. 81*,683-692.
Dimroth, E., W. R. A. Barager, R. Bergeron, and G. D. Jackson (1970). The filling of the circum-Ungava geosyncline, in Basins and Geosynclines of the Canadian Shield, A. J. Baer, ed., *Geol. Surv. Can. Paper 70-40,* 45-142.
Douglas, R. J. W., and R. A. Price (1972). Nature and significance of variations in tectonic styles in Canada, in Variations in Tectonic Styles in Canada, R. A. Price and R. J. W. Douglas, eds., *Geol. Assoc. Can. Spec. Paper II*, 625-683.
Drummond, K. J. (1974). Paleozoic Arctic margin of North America, in *The Geology of Continental Margins*, C. A. Burk and C. L. Drake, eds., Springer-Verlag, Berlin, pp. 797-810.
Drury, S. A. (1977). Structures induced by granite diapirs in the Archean greenstone belt at Yellowknife, Canada: implications for Archean geotectonics, *J. Geol. 85*, 345-358.
Emslie, R. F. (1977). Anorthosite massifs, rapakivi granites, and late Proterozoic rifting of North America, *Precambrian Res. 7*, 61-98.
Engel, A. E. J., and C. G. Engel (1964). Continental accretion and the evolution of North America, *Advan. Frontiers Geol. Geophys*., 17-37e.
Engel, A. E. J., and D. L. Kelm (1972). Pre-Permian global tectonics: a tectonic test, *Geol. Soc. Am. Bull. 83*, 2325-2340.
Ervin, C. P., and L. D. McGinnis (1975). Reelfoot rift: reactivated precursor to the Mississippi Embayment, *Geol. Soc. Am. Bull. 86*, 1287-1295.
Fyson, W. K. (1978). Structures induced by granite diapirs in the Archean greenstone belt at Yellowknife, Canada: implications for Archean geotectonics: a discussion, *J. Geol. 86*, 767-769.
Garrison, J. R., Jr. (1978). A petrogenetic model for the southeastern Llano uplift, Texas, *Eos Trans. Am. Geophys. Union 59*, 1211.
Garrison, J. R., Jr.. and C. Ramirez-Ramirez (1978). The Llano uplift, central Texas: evidence for a Precambrian triple spreading system, *Geol. Soc. Am. Abstr. Programs 10*, 106.
Gibb, R. A. (1975). Collision tectonics in the Canadian Shield?, *Earth Planet. Sci. Lett. 27*, 378-382.
Gibb, R. A. (1977). A gravity survey of James Bay and its bearing on the Kapuskasing Gneiss Belt, Ontario, *Tectonophysics 45*, T7-T13.
Glikson, A. Y. (1976). Earliest Precambrian ultramafic-mafic volcanic rocks: ancient oceanic crust or relic terrestrial maria?, *Geology 4*, 201-205.
Goldich, S. S., and C. E. Hedge (1974). 3,800 m.yr. granitic gneiss in southwestern Minnesota, *Nature 252*, 467-468.
Goodwin, A.M. (1974a). The most ancient continental margins, in *The Geology of Continental Margins*, C. A. Burk and C. L. Drake, eds., Springer-Verlag, Berlin, pp. 767-780.
Goodwin, A.M. (1974b). Precambrian belts, plumes, and shield development, *Am. J. Sci. 274*, 987-1028.
Green, D. H. (1975). Genesis of Archean peridotitic magmas and constraints on Archean geothermal gradients and tectonics, *Geology 3*, 15-18.
Ham, W. E., R. E. Denison, and C. A. Merritt (1964). Basement rocks and structural evolution of southern Oklahoma, *Okla. Geol. Surv. Bull. 95*, 302 pp.
Hargraves, R. B. (1976). Precambrian geologic history, *Science 193,* 363-371.
Hills, F. A., and R. L. Armstrong (1974). Geochronology of Precambrian rocks in the Laramie Range and implications for the tectonic framework of Precambrian southern Wyoming, *Precambrian Res. 1*, 213-225.
Hills, F. A., R. S. Houston, and G. V. Subbarayudu (1975). Possible Proterozoic plate boundary in southern Wyoming, *Geol. Soc. Am. Abstr. Programs 7*, 614.
Hoffman, P. (1973). Evolution of an early Proterozoic continental margin, *Phil. Trans. Roy. Soc. London A273*, 547-581.
Houston, R. S. (1971). *Regional Tectonics of the Precambrian Rocks of the Wyoming Province and Its Relationship to Laramide Structure,* Wyoming Geological Association Guidebook Twenty-third Annual Field Conference, A. R. Renfro, ed., pp. 19-27.
Houston, R. S., *et al*. (1968). *A Regional Study of Rocks of Precambrian Age in That Part of the Medicine Bow M?? Lying in Southeastern Wyoming—With a Chapter on ??tionship Between Precambrian and Laramide St.??*, Mem. 1, Wyoming Geological Survey, 167 pp.
Hurley, P.M., and J. R. Rand (1969). Pre-drift continental nuclei, *Science 164*, 1229-1242.
Jackson, D. G., and F. C. Taylor (1972). Correlation of major Aphebian rock units in the northeastern Canadian Shield, *Can. J. Earth Sci. 9*, 1650-1669.
Jordan, T. H. (1975). The continental tectosphere, *Rev. Geophys. Space Phys. 13*, 1-12.
Kearey, P. (1976). A regional structural model of the Labrador. Trough, northern Quebec, from gravity studies, and its relevance to continent collision in the Precambrian, *Earth Planet. Sci. Lett. 28*, 371-378.
King, E. R., and I. Zietz (1978). The New York-Alabama lineament: geophysical evidence for a major crustal break in the basement beneath the Appalachian basin, *Geology 6*, 312-318.
King, P. B., and H. M. Beikman (1974). Geologic map of the United States, scale 1:2,500,000, U.S. Geological Survey.
King, P. B., and P. T. Flawn (1953). *Geology and Mineral Deposits of Precambrian Rocks of the Van Horn Area, Texas*, Publ. 5301, Bureau of Economic Geology, U. of Texas, 218 pp.
Kistler, R. W., and Z. E. Peterman (1978). Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks, *U.S. Geol. Surv. Prof. Paper 1071*, 17 pp.
Krogh, T. E., and G. L. Davis (1971). Zircon U-Pb ages of Ar

chean metavolcanic rocks in the Canadian Shield, *Carnegie Inst. Washington Yearb. 70*, 241-242.
Krogh, T. E., I. F. Ermanovics, and G. L. Davis (1974). Two episodes of metamorphism and deformation in the Archean rocks of the Canadian Shield, *Carnegie Inst. Washington Yearb. 73*, 573-575.
Krogh, T. E., G. L. Davis, N. B. W. Harris, and I. F. Ermanovics (1975). Isotopic ages in the eastern Lac Seul region of the English River gneiss belt, *Carnegie Inst. Washington Yearb. 74*, 623-625.
Kroner, A. (1977). Precambrian mobile belts of southern and eastem Africa—ancient sutures or sites of ensialic mobility? A case for crustal evolution towards plate tectonics, *Tectonophysics 40*, 101-135.
Kumarapeli, P.S. (1978). The St. Lawrence paleorift system: a comparative study, in *Tectonics and Geophysics of Continental Rifts*, I. B. Ramberg and E. R. Neumann, eds., D. Reidel, Dordrecht, Holland, pp. 367-384.
Lidiak, E.G. (1971). Buried Precambrian rocks of South Dakota, *Geol.Soc. Am. Bull. 82*, 1411-1420.
Lidiak, E.G., and I. Zietz (1976). Interpretation of aeromagnetic anomalies between latitudes 37ºN and 38ºN in the eastern and central United States, *Geol. Soc. Am. Spec. Paper 167*, 37 pp.
Martin, H., and H. Porada (1977). The intracratonic branch of the Damara orogen in South West Africa, *Precambrian Res. 5*, 311-338.
Maxwell, J. C. (1974). Early western margin of the United States, in *The Geology of Continental Margins*, C. A. Burk and C. L. Drake, eds., Springer-Verlag, Berlin, pp. 831-851.
McElhinney, M. W., and M. O. McWilliams (1977). Precambrian geodynamics—a paleomagnetic view, *Tectonophysics 40*, 137-159.
Molnar, P., and P. Tapponnier (1975). Cenozoic tectonics of Asia: effects of a continental collision, *Science 189*, 419-426.
Montgomery, A. (1953). Precambrian geology of the Picuris Range, northcentral New Mexico, *N. Mex. Bur. Mines Miner. Resour. Bull. 30*, 89 pp.
Moorbath, S., R. K. O'Nions, R. J. Pankhurst, N. H. Gale, and V. R. McGregor (1972). Further rubidium-strontium age determinations on the very early Precambrian rocks of the Godthaab district, West Greenland, *Nature Phys. Sci. 240*, 78-82.
Morey, G. B., and P. K. Sims (1976). Boundary between two Precambrian W terranes in Minnesota and its geologic significance, *Geol. Soc. Am. Bull. 87*, 141-152.
Muehlberger, W. R., R. E. Denison, and E.G. Lidiak (1967). Basement rocks in continental interior of United States, *Am. Assoc. Petrol. Geol. Bull. 51*, 2351-2380.
Myers, J. S. (1976). Granitoid sheets, thrusting, and Archean crustal thickening in West Greenland, *Geology 5*, 265-268.
Nunes, P. D., and P. C. Thurston (1978). Evolution of a single greenstone belt over 220 million years—a zircon study of the Uchi Lake area, northwestern Ontario, in Short Papers of the Fourth International Conference, Geochronology, Cosmochronology, Isotope Geology, R. E. Zartman, ed., U.S. *Geol. Surv. Open-File Report 78-701*, pp. 313-315.
Nunes, P. D., D. R. Pyke, and L. S. Jensen (1978). Toward an absolute age stratigraphy of the Abitibi greenstone belt—zircon ages from the Timmins and Kirkland Lake areas, *Geol. Soc. Am. Abstr. Programs 10*, 464-465.
Peterman, Z. E., and R. A. Hildreth (1977). Reconnaissance geology and geochronology of the Precambrian of the Granite Mountains, Wyoming, *U.S. Geol. Surv. Open-File Report 77-140*.
Price, R. A., and R. J. W. Douglas, eds. (1972). Variations in tectonic styles in Canada, *Geol. Assoc. Can. Spec. Paper 11*, 688 pp.
Robertson, J. M., K. C. Condie, and D.C. Mathewson (1978). The Pecos greenstone belt: a Proterozoic volcano-sedimentary sequence in the southern Sangre de Cristo Mountains, New Mexico, *Geol. Soc. Am. Abstr. Programs 10*, 479.
Sawkins, F. J. (1976). Widespread continental rifting: some considerations of timing and mechanism, *Geology 4*, 427-430.
Sears, J. W., and R. A. Price (1978). The Siberian connection: a case for Precambrian separation of the North American and Siberian cratons, *Geology 6*, 267-270.
Sengor*, A. M. C., and J. C. Butler (1977). The Llano uplift, central Texas: a Proterozoic example of continental collision?, *Geol. Soc.Am. Abstr. Programs 9*, 72-73.
Seyfert, C. K. (1978). Palaeomagnetic evidence for three Wilson cycles during the Precambrian, *Geol. Soc. Am. Abstr. Programs 10*, 490.
Silver, L. T., M. E. Bickford, W. R. Van Schmus, J. L. Anderson, T. H. Anderson, and L. G. Medaris (1977). The 1.4-1.5 b.y. transcontinental anorogenic plutonic perforation of North America, *Geol. Soc. Am. Abstr. Programs 9*, 1176-1177.
Sims, P. K. (1976). Precambrian tectonics and mineral deposits, Lake Superior region, *Econ. Geol. 71*, 1092-1127.
Stewart, J. H. (1976). Late Precambrian evolution of North America: plate tectonics implications, *Geology 4*, 11-15.
Stockwell, C. H. (1961). Structural provinces, orogenies, and time classification of rocks of the Canadian Precambrian Shield, *Can. Geol. Surv. Paper 61-17*, 108-118.
Sutton, J. (1973). Some changes in continental structure since early Precambrian time, in *Implications of Continental Drift to the Earth Sciences*, D. H. Tarling and S. K. Runcorn, eds., Academic, New York, pp. 1071-1081.
Swan, M. M. (1975). The Texas lineament-tectonic expression of a Precambrian orogeny, *Geol. Soc. Am. Abstr. Programs 7*, 1288-1289.
Sykes, L. R. (1978). Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation, *Rev. Geophys. Space Phys. 16*, 621-688.
Tectonic Map of Canada (1969). Geological Survey Canada Map 1251A.
Upadhyay, H. D. (1978). Phanerozoic, peridotitic, and pyroxenitic komatiites from Newfoundland, *Science 202*, 1192-1195.
Warner, L. A. (1978). The Colorado lineament: a middle Precambrian wrench fault system, *Geol. Soc. Am. Bull. 89*, 161-171.
Wickham, J. (1978). The southern Oklahoma aulacogen, in *Structural Style of the Arbuckle Region*, J. Wickham and R. E. Denison, field trip leaders, Geological Society of America, South-Central Sect., Field Trip 3, pp. 9-41.
Wiley, M. A., and W. R. Muehlberger (1971). The Texas lineament, in *The Geologic Framework of the Chihuahua Tectonic Belt*, K. Seewald and D. Sundeen, eds., Publ. 71-59, West Texas Geological Society, pp. 15-23.
Winalley, B. F. (1977). *The Evolving Continents*, John Wiley, New York, 385 pp.
Wynne-Edwards, H. B. (1976). Proterozoic ensialic orogenesis: the millipede model of ductile plate tectonics, *Am. J. Sci. 276*, 927-953.

16

An Outline of the Tectonic Characteristics of China

T. K. Huang
Chinese Academy of Geological Sciences

INTRODUCTION

Earlier in this century, western scholars were introduced to the geology of China from the works of Richthofe, Loczy, Obruchev, Willis, Grabau, and others. Unfortunately, no comprehensive treatise on the geology of China was available until the *Geology of China* (Lee, 1939) appeared. Lee stressed the concept of "tectonic systems" that led to the development of his well-known theory of geomechanics. Huang (1945) assembled and analyzed all available geological data and attempted to elucidate systematically the tectonic characteristics of China in his book, *On Major Tectonic Forms of China*. Recently, greater attention has been paid by Chinese geologists to the theories of seafloor spreading and plate tectonics. As most of the Chinese works are published in Chinese, this paper intends to give a current summary of the tectonic characteristics of China. Additional background information can be found in the papers of Huang (1945, 1960), Huang *et al*. (1974), Li (1975), and Terman (1973).

TECTIONIC UNITS OF CHINA

Sino-Korean Paraplatform

The Sino-Korean paraplatform (triangular in form) covers the entire territory of North China, the northern part of the Yellow Sea, and the northern part of Korea (Figure 16.1). This paraplatform is the oldest in China, its basement last affected by the Chungtiao* orogeny at about 1700 million years (m.y.) ago (Table 16.1). The marginal parts of the

This paper was originally published essentially in its entirety in *Eclogae Geol. Helv*. 71, 611-635, 1978.

* In the Luliang area, the Cambrian lies directly upon the lower Proterozoic, the upper Proterozoic is lacking. The term "Luliang orogeny" is here replaced by the Chungtiao orogeny, with the type locality of Chungtiaoshan, where the upper Proterozoic is well developed and lies unconformably upon the lower Proterozoic.

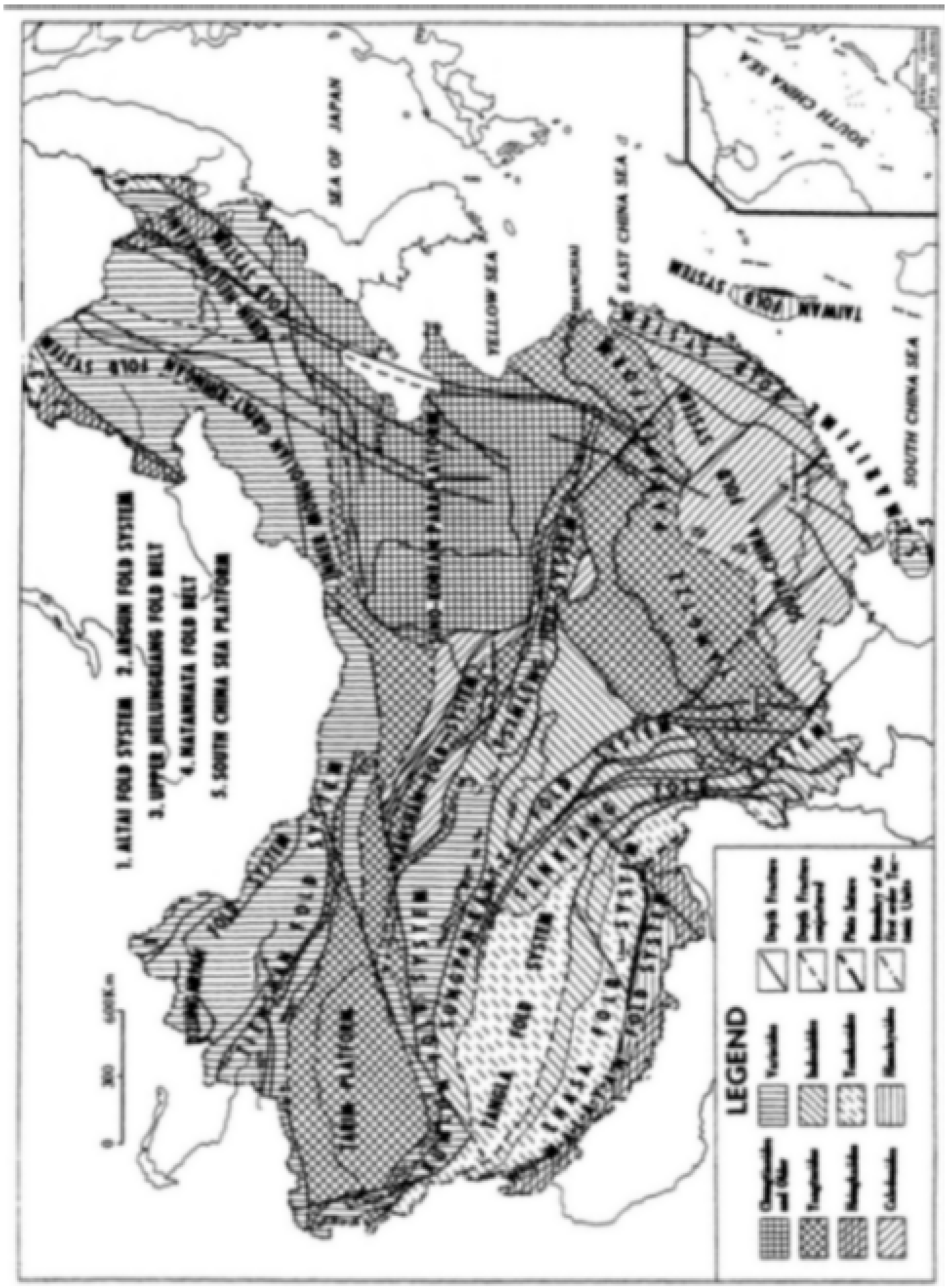


Figure 16.1
A simplified tectonic map of China.

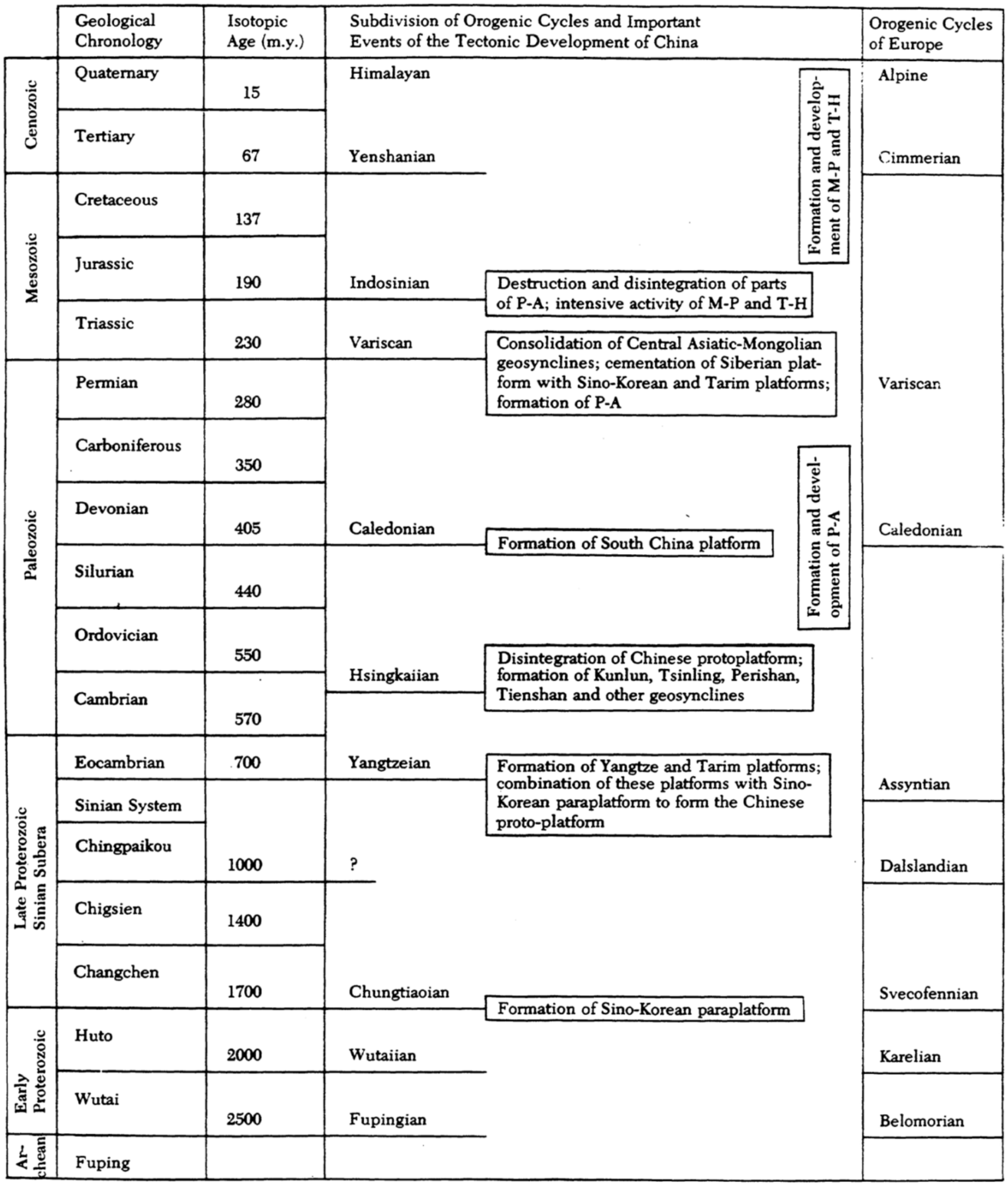


Figure 16.1
Subdivision of Orogenic Cycles and Important Events of the Tectonic Development of China

paraplatform, such as the Alashan region, were consolidated by the end of the Proterozoic through the Yangtze cycle. Geologists now agree that the main portion of the paraplatform developed through three stages, as revealed by three major unconformities (Ch'eng *et al*., 1973). These are (1) the unconformity between the Archean Fuping Group and the lower Proterozoic Wutai Group, representing the Fuping orogeny about 2350-2550 m.y. ago*; (2) the unconformity between the Wutai Group and the upper part of the lower Proterozoic, or the Huto Group, representing the Wutai orogeny about 2000 m.y. ago; and (3) the unconformity between the Huto Group or its equivalents and the upper Proterozoic Sinian Suberathem,* representing the Chungtiao orogeny. The blanket of the platform consists of Sinian and Cambro-Ordivician neritic sediments (largely carbonates), Pertmo-Carboniferous terrestrial sediments (with marine intercalations), as well as purely terrestrial Meso-Cenozoic sediments. The Sinian System is present in a few regions, while Silurian, Devonian, and lower Carboniferous are absent. An abundance of Mesozoic continental volcanics and granitoid intrusions occurs in the eastern part of the paraplatform, especially in the Yenliao depression, Shangtung, and other regions. Cenozoic basalts are widespread. Tectonic disturbances of the Indosinian cycle were limited in Inner Mongolia, Liaotung, and the Yenliao depression.

Yangtze Paraplatform

The Yangtze paraplatform (Figure 16.1) comprises the greater part of the Yangtze Basin from eastern Yunnan to Kiangsu, also including the southern part of the Yellow Sea. Previously, the consolidation of the Yangtze paraplatform was considered contemporaneous with that of the Sino-Korean paraplatform. Recent investigations using stromatolites, micropaleoflora, and isotopic geochronology of the basement rocks indicate that it was consolidated about 700 m.y. ago (Yangtze orogeny). The blanket oft he paraplatform consists chiefly of carbonate and clastic deposits ranging in age from Sinian System to Triassic, with Devonian and Carboniferous normally absent in Szechwan and North Kweichow, whereas terrestrial Jurassic, Cretaceous, and younger deposits occur in Szechwan, Central Yunnan, Hupeh, Kiangsu, and other regions. Tectonism was strong chiefly in the Yenshanian cycle with the formation of well-developed blanket folds. The Kam-Yunnan Axis and the Lower Yangtze, however, suffered polycyclic deformation and magmatism; the former belongs essentially to the Variscan cycle, while the latter belongs to the Yenshanian and Indosinian cycle. Formerly, the Huaiyang Massif was considered a portion of the Sino-Korean paraplatform; recent data show that it belongs to the Yangtze paraplatform.

TARIM PLATFORM

The Tarim platform (Figure 16.1) is defined by the Tienshan fold system in the north and by the Kunlun fold system in the south and is largely covered by Cenozoic deposits. Its basement, together with its Paleozoic blanket, crops out along the northern border as seen at Kelpin and Kuruk Tagh. In the latter region, late Proterozoic (Sinian) tillites and stromatolites are found, showing that the Tarim platform took shape near the end of the Proterozoic just as the Yangtze paraplatform did.

Tienshan-Khingan Geosynclinal Fold System and Argun Fold System

These two systems are components of the great central Asiatic-Mongolian arcuate fold region, which extends between the Siberian platform, on the north, and the Sino-Korean paraplatform and the Tarim platform, on the south (Figure 16.1). This fold region is separated into two halves by the Derbugan depth fracture (eastern extension of the Mid-Mongolian depth fracture); the southern half is the Tienshan-Khingan geosynclinal fold system, while the northern half includes the Argun fold system. The latter is Hsingkaiian, and the former consists, within the Chinese territory, of the Altai (Caledonides and Variscides), the Dzungarian (Variscides), the Tienshan (Variscides), and the Kirin-Heilungkiang (Variscides) fold systems. All of them, except the southern Tienshan, are eugeosynclinal in nature. During the Caledonian cycle, volcanic activity, chiefly submarine, was widespread in the Ordovician and Silurian. Radiolarian cherts were found closely associated with spilites and ultrabasic intrusives forming ophiolitic suites, such as in the western Dzungaria. During the Variscan cycle, the Devonian and Carboniferous were characterized by calc-alkaline volcanics (mainly andesites). From Carboniferous to the end of the Permian,* the geosynclinal formations accompanied by granitoid intrusives were brought into intense and complicated folds, thus converting the geosyncline into a craton. During the Yenshanian cycle, a general tendency of increasing tectonic intensity from east to west can be recognized, especially in the Kirin-Heilungkiang and Great Khingan regions, where strong remobilization and magmatism took place. During the Himalayan cycle, the general tendency of increasing intensity of tectonism was inverted. Faulting and uplift along the general strike of the Tienshan were strong and widespread, forming lofty mountain ranges with sunken northern and southern foredeeps as

** The author uses the classification of the Sinian as given in the new Geological Map of China (1:4,000,000). The term. "Sinian System," with its standard section in the Yangtze Gorges, comprises sediments ranging in age from 600-800 m.y. ago. The term "Sinian Suberathem," with its standard section in Chihhsien near Peking, comprises sediments ranging in age from 600-1700 m.y. ago. The term Sinian in this paper corresponds to Sinian Suberathem.

* This is the Variscan orogeny. Caledonian orogenic movements are also present but they are very limited in distribution.

* The oldest basement rocks are dated at 3200-3400 m.y. ago.

well as intermontane depressions such as the famous Turfan Basin.

Kunlun-Nanshan-Tsinling Geosynclinal Fold System and Tibet-Yunnan Geosynclinal System

These two tectonic units, separated from one another by the Chinshakiang-Red River depth fracture, occupy the vast territory south of the Tarim platform and Sino-Korean paraplatform, west of the Yangtze paraplatform, and north of the Tsangpo depth fracture (Figure 16.1).

The Kunlun-Nanshan-Tsinling geosynclinal fold system is a typical polycyclic geosynclinal fold system embracing three orogenic events. During the Caledonian cycle, a eugeosyncline came into existence mainly in the Nanshan, with well-developed ophiolite suites and glaucophane schist belts. During the Variscan cycle, a large portion of the system was miogeosynclinal, while eugeosynclines were maintained only in Burkhanbuddha, Amnemachin, and along the Chinshakiang. During the Indosinian cycle, miogeosynclinal development continued in the Sungpan-Kantze, Tsinling, and South Kokonor Range, while the area along the Chinshakiang remained eugeosynclinal. Different parts of the system differ in age of folding. The Nanshan is Caledonian, the Kunlun is Variscan, while the Sungpan-Kantze and Tsinling are basically Indosinian. The Yenshanian and Himalayan cycles are characterized by remobilization accompanied by a variable amount of magmatism. That Cambrian volcanics were overthrust upon late Tertiary red beds in the Lachishan of Chinghai and Jurassic sandstones were overthrust upon Quaternary gravels at the Hungliu Gorge near Yumen of Kansu demonstrate that intensive horizontal compression has been a factor even in more recent geological time.

The Tibet-Yunnan geosynclinal fold system, situated to the south and west of the Chinshakiang-Red River depth fracture, is a Mesozoic arcuate fold system, consisting of the Sankiang* or Three-River (Indosinides), the Tangla (Yenshanides), and the Lhasa (Yenshanides) fold systems. The existence of the proposed Precambrian Tibetan Massif (Terman, 1973) is negated by recent observations. The Sankiang fold system includes western Yunnan and the Changtu district of Tibet and is characterized by geosynclinal folds strictly controlled by depth fractures among which the Chinshakiang depth fracture, the Lantsangkiang depth fracture, and the Nukiang depth fracture are of prime importance. Along these depth fractures, eugeosynclinal activity took place, with polycyclic tectonism accompanied by intensive polycyclic magmatism. Not only the Paleozoic and Mesozoic, but also, in some localities, the Tertiary formations underwent tectonism and metamorphism in various grades and formed several tectono-magmato-metamorphic belts (the Ailaoshan, the Lantsangkiang, and the Kaolikungshan metamorphic belts).

Himalayan Geosynclinal Fold System

This tectonic unit (Figure 16.1), situated to the south of the Tsangpo and Indus Rivers, consists of Sulaiman (in Pakistan), the Himalaya, and the Arakan Yoma (in Burma). It must be pointed out that true geosynclinal deposits are absent in the Himalaya proper but appear immediately to the south of the Tsangpo, where a more or less continuous belt of ophiolites is well exposed. The lofty ranges of the main Himalaya form in fact the northern border of the Indian craton and associated Hsingkaiides, the metamorphic and sedimentary rocks of which were deeply involved in the Himalayan folding when the Indian craton collided with Tibet (Chang and Cheng, 1973; Gansser, 1964, 1966; Molnar and Tapponnier, 1975).

South China Geosynclinal Fold System

This tectonic unit is located to the south of the Yangtze paraplatform (Figure 16.1). The author believes that this unit is not entirely Caledonian as had been previously regarded but consists of two parts of different character. The major part, still called the South China fold belt, is truly Caledonian, but the subordinate part, including coastal regions of Chekiang, Fukien, and Kwangtung and the neighboring continental shelves as well as a greater part of the Hainan Island, belongs to the Variscan, to be named the Southeastern Maritime fold system. Because the latter was strongly transformed by Mesozoic tectonism and widely covered with Jurassic and Cretaceous continental volcanics, its tectonic character has long been misjudged.

Natanhata Eugeosynclinal Fold Belt and Upper Heilungkiang Miogeosynclinal Fold Belt

These are components of the Mesozoic geosynclinal fold systems located to the east of the Siberian platform (Figure 16.1). The Natanhata belongs to the Sikhote-Alin fold system, while the Upper Heilungkiang belongs to the Mongolo-Okhotsk fold system. Both are Yenshanides.

DEEP-SEATED STRUCTURES AND DEPTH FRACTURES IN CHINA

Deep-Seated Structure

From a comprehensive examination of available geophysical and seismic data, it is preliminarily concluded that the lithosphere within China can be subdivided into several heterogeneous layers, as elsewhere in the world. This layering character is demonstrated by recent investigations (Hsi *et al*., 1974a; 1974b; 1975) of seismic sound

* So named because it embraces the Chinshakiang, the Lantsangkiang, and the Nukiang drainage systems.

ing along a profile from Yuanshih to Tsinan in the North China Plain and another profile across the eastern section of the Tsaidam Basin. In these regions the earth's crust possesses layers of velocity gradients in addition to low-velocity zones.

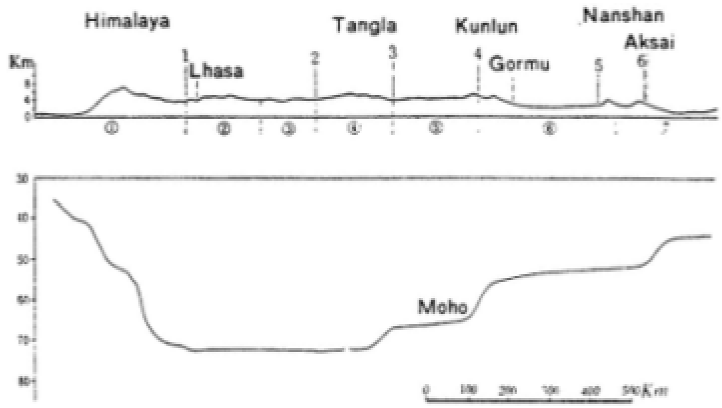


Figure 16.2
Cross section of the Moho from Himalaya to Nanshan. 1, Tsangpo depth fracture; 2, Lantsangkiang depth fracture; 3, Chinshakiang depth fracture; 4, East Kunlun depth fracture; 5, depth fracture along northern margin of Tsaidam; 6, Altyn depth fracture;①, Himalaya fold system; ②, Lhasa fold system; ③, Sankiang fold system;④, Tangla fold system;⑤, Sungpan-Kantze fold system; ⑥, Kunlun fold system; ⑦, Nanshan fold system.

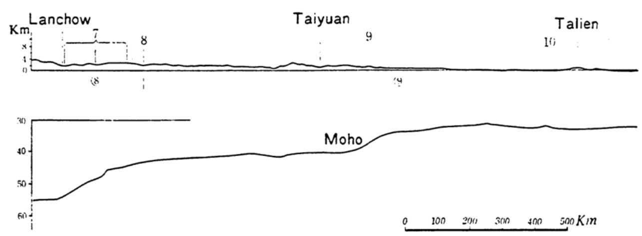


Figure 16.3
Cross section of the Moho from Lanchow to Talien. 7, depth fracture zone of North Nanshan; 8, depth fracture of western margin of Ordos; 9, depth fracture zone of Taihangshan; 10, Tancheng-Lukiang depth fracture zone;⑧), Nanshan fold system; ⑨, Sino-Korean paraplatform.

A blocklike pattern is developed for the Chinese mainland, especially its eastern part, when the areal variations of the depth to the Moho (Mohorovicic* discontinuity) is considered. Two .cross sections show the nature of the depth to the Moho (Figures 16.2 and 16.3). Each block is confined generally by depth fractures of persistent development. For example, the Yinchuan-Liupanshan-Kam-Yunnan* gravity gradient belt (Kunming-Yinchuan depth fracture system) separated eastern China from western China in the geological past. The Great Khingan-Taihangshan-Wulingshan* gravity gradient belt, a dividing line between the eastern and western belts of the Marginal-Pacific tectonic domain, came into existence long since the late Mesozoic. The northern section of the above-mentioned gravity gradient belt, i.e., the section from the Great Khingan to Taihangshan, coincides with a depth fracture zone. The North Tsinling-North Huaiyang depth fracture zone is a geological dividing line between North and South China. The gravity gradient belt of West Kunlun-Altyn-North Nanshan essentially coincides with the depth fracture zones between the Tarim platform and the Alashan Massif on the north and the geosynclinal fold belts on the south.

Moreover, the present geomorphic features of China form the mirror image of the Moho, as shown in Figures 16.2, 16.3, and 16.4; mountains and plateaus correspond to the "downwarps" of the Moho, while plains and basins agree with its "uplifts." This appears to be a result of a combined effect of the Pacific plate and the Indian plate acting against the continental block of China since the Indosinian, particularly since the Himalayan cycle. In these movements, not only the crust but also the upper mantle were involved.

Depth Fractures of China

The author classifies depth fractures into three classes according to their depth (Table 16.2). Translithospheric fractures either came into being during a definite geological period or continued in motion till today marked by a series of deep-focus earthquakes (Benioff zones). Lithospheric fractures are generally characterized by basic and ultrabasic complexes but without the development of extensive ophiolites. Crustal fractures are of much

* No N-S directed depth fractures are present along the Wulingshan.

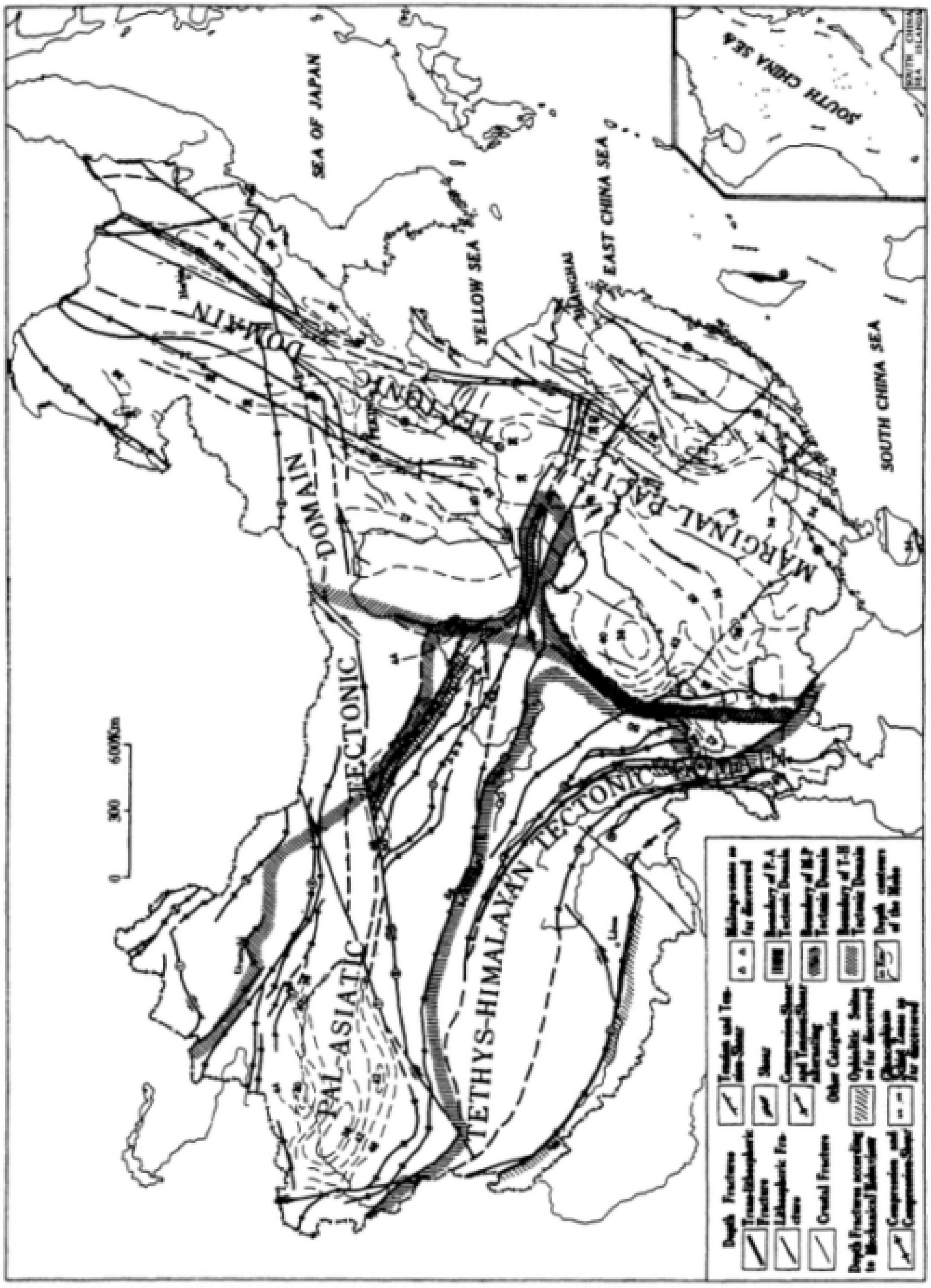


Figure 16.4
A simplified map showing depth fractures and tectonic domains of China.

smaller magnitude and are usually accompanied by acid and/or intermediate magmas (sialic fractures) or by basalts (simatic fractures).

TABLE 16.2 Classification of Depth Fractures

		Examples	
Class	Depth	World	China
Sialic crustal fracture	Dissecting the sial but not clearly extending into the sima	Many	Heyuan depth fracture
Simatic depth fracture	Dissecting the whole crust but not extending into the mantle	Many	East Tsangchow depth fracture
Lithospheric fracture	Dissecting the lithosphere but not extending into the asthenosphere	Rift Valleys of East Africa	Tancheng-Lukiang fracture
Translithospheric fracture	Dissecting the lithosphere and extending deep into the asthenosphere	Benioff zones along island arcs of the Pacific	Tsangpo depth fracture

Depth fractures can also be divided into three classes* according to their mechanical behavior: tension-depth fracture, compression-depth fracture, and shear-depth fracture. Fractures along midoceanic ridges and along the great rift valleys are tension-depth fractures. Fractures along the island arcs of the western Pacific (marked by deep-sea trenches) are examples of compression-or compression-shear-depth fractures. The San Andreas Fault of North America, the Altyn Depth Fracture, and the Tancheng-Lukiang Depth Fracture of China are examples of shear-depth fractures. Depth fractures are not unchangeable; on the contrary, their formation, development, transformation, and extinction came one after another in succession during the long history of the evolution of the earth. Numerous facts note that different sections of the same depth-fracture zone might have originated in different geological ages, belong to different types, and possess different characteristics.

Based on the available geological and geophysical data, the main depth fractures are listed in Table 16.3, and a brief description of them follows.

Those belonging to the *translithospheric depth fractures* are as follows:

1. The Tsangpo depth-fracture zone (Huang, 1960) extends westward following the Indus River valley and was termed the Indus suture (Fitch, 1972). It is now considered as the suture line between the Indian plate and the Eurasian plate and is marked by well-developed ophiolites extending about 2000 km within the confines of China (Chang and Teenage, 1973; Gansser, 1964).
2. The longitudinal depth-fracture zone of Taiwan forms only a small section of the long depth fractures (Benioff zones) along the island arcs of the western Pacific. Its presence is marked by ophiolites, glaucophane schists and mélanges, and frequent deep-focus earthquakes. This depth fracture is sinistral shear in mechanical behavior in contrast to the predominantly compressional Benioff zones.

Available data indicate that the North Nanshan depth fracture, the Chinshakiang-Red River depth fracture, and the Derbugan (Mid-Mongolian) depth fracture are translithospheric, or ancient plate sutures.

In China a great majority of depth fractures in geosynclinal regions and a part of those in platforms belong to *lithospheric depth fractures.* Among them the dominant ones are the following:

(a) The Tancheng-Lukiang depth fracture system is perhaps the major depth-fracture system of eastern Asia and is composed of the Tancheng-Lukiang, the Mishan-Tunhua, and the Yilan-Yitung depth fractures with a total length of about 2400 km. From south to north, it cuts through the Yangtze paraplatform, the Sino-Korean paraplatform, and the Kirin-Heilungkiang fold system. It also appears to be an important volcanic, metallogenic, and seismic zone, especially during Meso-Cenozoic times. Sinistral shear occurred in the geological past, but dextral shear is apparent from analyses of modern earthquake mechanisms. Some geologists consider that this fracture system existed in Pre-Sinian times, while others suggest that it was formed in the Indosinian cycle. The fracture system continues northward into the Far East of the U.S.S.R.

(b) The Kunlun-Tsinling depth fracture system includes the North Nanshan-North Tsinling-North Huaiyang fracture zone, the northern margin of Tsaidam-South Kokonor Range-North Tsinling-North Huaiyang fracture zone, and the east Kunlun-Tsinling fracture zone.

These complicated fracture zones controlled the origination and development of the Kunlun-Tsinling geosynclinal fold system and constitute a geological dividing line between northern and southern China. This dividing line shifted from the line of North Nanshan-North Tsinling-North Huaiyang during the Caledonian cycle to the line of the northern margin of Tsaidam-South Kokonor Range-North Tsinling-North Huaiyang during file early Variscan cycle, and finally to the line of East Kunlun-North Tsinling-North Huaiyang since the late Variscan cycle. It is important to note that these fracture zones deepen from east to west. Widespread submarine

* In fact, there are transitional types, such as compression-shear and tension-shear depth fractures, which are attributable respectively to compression, tension, and shear depth fractures according to their dominant mechanical behavior.

volcanics occur along the western section, where eugeosynclinal conditions prevailed, and especially in North Nanshan, where remarkable zones of ophiolites 700 km long and zones of glaucophane schists more than 100 km long were recently discovered, indicating the existence of an ancient plate suture. On the contrary, the eastern sections of these fractures, being characterized by flysch formations and scanty ultrabasic rocks, appear miogeosynclinal. This is perhaps the reason why the Tsinling geosyncline became appendix like toward the east between the Sino-Korean and the Yangtze paraplatforms.

TABLE 16.3 The Principal Depth Fractures in China[a]

Number	Depth Fracture Zone	Depth	Character	Age of Activity	Magmatism and Metamorphism
1	Darbut	L	s	Pz	OS
2	Irtish	L	c-s	Pz	Σ
3	Karameili	L	c-s	Pz	Σ
4	North Margin of Central Tienshan	L	c-s	Pt?,Pz	Σ
5	Derbugan	L	c-s	Pz	
6	Silamulun	L	c-s	Pz	Σ,gs
7	Cherchen	L	s-s	Pz	
8	Northern Margin of Inner Mongolian Axis	L	c-s	Pt?,Pz	Σ
9	Altyn	L	s-s	Pz	Σ
10	North Nanshan–Northern Margin of North Tsingling–North Huaiyang	L	c-s	Pt,Pz	OS,gs
11	Northern Margin of Tsaidam–Southern Margin of North Tsingling–North Huaiyang	L	c-s	Pz	Σ,m
12	East Kunlun	L	c-s	Pz	Σ,m
13	Kantze-Litang	L	c-s	Pz	Σ
14	Chinshakiang–Red River	L	c-s	Pz	Σ,m
15	Lantsangkiang	L	c-s	Pz	Σ
16	Nukiang	L	c-s	Pz	Σ
17	Anningho	L	c-s	Pz	Σ
18	Hsiaokiang	C	c-s	Pz	β
19	Lungmenshan	L	c-s	Pz	Σ
20	Lingshan	C	c-s	Pz	γ
21	Wuchuan-Szehwei	C	a	Pz	γ
22	Heyuan	C	a	Mz	γ
23	Lishui-Haifeng	C	c-s	Pz	γ
24	Changle-Amoy	C	c-s	Mz	γ
25	Taihangshan	C	c-s	Mz	γ
26	East Tsangchow	C	t-s	Mz	β
27	Liaocheng-Lankao	C	t-s	Mz	β
28	Tancheng-Lukiang	L	s-s	Pt?,Mz?	Σ
29	Yilan-Yitung	C	t-s	Mz	β
30	Fushun-Mishan	L	s-s	Pz	β
31	Tsango-Indus	T	c-s	Mz	OS,m
32	Longitudinal Valley of Taiwan	T	s-s	Mz	OS,m,gs

[a]Numbers correspond to those circled on Figure 16.4.

T, Translithospheric fracture zone
L, Lithospheric fracture zone
C, Crustal fracture zone
c-s, Compression and compression–shear
t-s, Tension and tension–shear
s, Shear
s-s, Sinistral shear
a, Compression–shear and tension–shear, alternating

Pt, Active since Proterozoic
Pz, Active since Paleozoic
Mz, Active since Mesozoic
OS, Ophiolite suites
Σ, Basic and ultrabasic complexes
γ, Granitoid plutons
β, Basalts
m, Mélanges
gs, Glaucophane schists

Crustal fractures are numerous in China as represented by large-scale acid-intermediate plutons with extensive volcanics of the "Pacific" type, lack of ultrabasic rocks, and zones of strong compression followed by strong tension along the same zones, thus giving rise to a series of Cretaceous-Tertiary basins filled with red beds along growth faults* of depth-fracture nature. Afterwards, in late Tertiary and Quaternary, they were subjected to compression again. In this respect, the Heyuan depth fracture is one of the fairly good examples.

Another type of crustal fracture in eastern China is represented by numerous tension fractures along which originated and developed Mesozoic and/or Tertiary fault basins, filled with terrestrial clastic deposits. Among them the East Tsangchow depth fracture, the Liaocheng-Lankao depth fracture, and the Weiho-Fenho graben (young Tertiary) are the most important. Unlike the Heyuan depth fracture, they originated as tension fractures in Cretaceous or old Tertiary time and are buried by young Tertiary and Quaternary deposits. Thus, they are typical growth faults, forming with clastic deposits, the so-called "dustpan-like basins" accompanied by basalt flows. Geophysical data show that these depth fractures dissect the entire earth's crust but do not extend deep into the upper mantle and therefore should belong to simatic fractures.

THE TECTONIC DEVELOPMENT OF CHINA

Two major stages of tectonic development of China before Paleozoic times can be distinguished. The first is the stage from Archean to early Proterozoic or the Pre-Sinian stage, through which the Sino-Korean paraplatform came into being about 1700 m.y. ago. The second is the late Proterozoic or Sinian stage, through which the Yangtze paraplatform and the Tarim platform came into being about 700-800 m.y. ago. The great importance of the Yangtze orogenic cycle should be emphasized here. Available data show that platform areas formed through the Yangtze orogeny were more extensive than the present Yangtze paraplatform and Tarim platform. Platform-type formations correlatable with the Sinian System lying unconformably upon metamorphosed Sinian Suberathem, are met with in East Kunlun, Tsinling, Altyn, and along the northern border of the Tsaidam Basin. Moreover, phosphatic sediments of early to middle Cambrian age have been discovered in the Tienshan and Peishan. These facts suggest that a gigantic craton, to which the name of Chinese proto-platform is temporarily given, might have been created by and maintained through the Yangtze orogeny with a time span of about 200 m.y. (Sinian System to lower Cambrian).

The tectonic development of China since Paleozoic time could be divided into two stages, the Paleozoic stage and the Meso-Cenozoic stage. In spatial distribution, three major units (tectonic domains) were developed during this long time period. They are the Pal-Asiatic (P-A,) tectonic domain, the Marginal Pacific (M-P) tectonic domain, and the Tethys-Himalayan (T-H) tectonic domain. The P-A, tectonic domain took shape and developed through the Hsingkaiian, the Caledonian, and the Variscan orogenies. The process was roughly as follows: as soon as the Sayan-North Mongolian-Argun Geosynclinal System was folded and uplifted at the end of the early Cambrian (Hsiangkaiian), the Chinese proto-platform was disintegrated and partially transformed to form the Tienshan, the Nanshan, the Tsinling, and other geosynclines, all of which were again folded, transformed, and consolidated at the end of the Variscan cycle. As a result, these newly consolidated geosynclinal folds joined and "cemented together" the pre-existing four platforms, the Siberian platform, the Tarim platform, the Sino-Korean paraplatform, and the Yangtze paraplatform, to form a gigantic craton called the Pal-Asia. During the Meso-Cenozoic stage, the tectonic development of China was under the control of the Marginal Pacific tectonic domain and the Tethys-Himalayan tectonic domain. The former may be subdivided into an inner or Cenozoic tectonic belt and an outer or Mesozoic tectonic belt. The Taiwan fold system is the only representative of the inner belt within the Chinese territory. The outer belt of great importance was superimposed upon older tectonic units of various ages. In addition to the Southeast Coast Variscides and the Mesozoic Northeast Asiatic geosynclinal fold system, they are, from north to south, the Inner Mongolian-Great Khingan Variscides, the Kirin-Heilungkiang Variscides, the Sino-Korean paraplatform, the Yangtze paraplatform, and the South China fold system (Caledonides). The belt is characterized by large swells and depressions trending NE or NNE, with Mesozoic blanket folds and faults, extensive "Pacific-type" volcanics, and large-scale granitoid intrusives.

THE POLYCYCLIC TECTONIC EVOLUTION OF CHINA

For many years, Stille's concepts of tectono-magnetic cycles have been prevalent in models of geosynclinal development:

1. A geosynclinal period with initial magmatism, mainly basic and ultrabasic rocks;
2. An orogenic period with synorogenic magmatism, mainly granitoid intrusions;
3. A quasicratonic phase with subsequent magmatism, mainly porphyries and andesites; and
4. A full cratonic period with final magmatism, mainly basalts.

Since then, many well-known tectonic and economic geologists have accepted this concept (monocyclic concept) especially in the Soviet Union and Western Europe, for example, Beloussov (1962), Rittmann (1960), deSitter (1964), and Aubouin (1965).

In the author's explorations of many key regions of the geology of China, he arrived at the conclusion that geosynclines developed polycyclically, both in orogeny and

* Also termed synsedimentary faults.

in magmatism (Huang, 1945, 1960; Huang and Chiang 1962; Huang *et al*., 1965, 1974). Van Bemmelen (1949) also advocated a polycyclic development for the Indosinian geosynclines. Recent geological mapping in the Tienshan and Nanshan again confirms the view of polycyclic development of geosynclinal fold belts, which is briefly summarized below.

1. The Tienshan fold system, especially the North Tienshan, consists of eugeosynclinal Variscides. The Nanshan fold system, especially the North Nanshan, consists of eugeosynclinal Caledonides. Both exhibit unquestionable polycyclic orogenies; seven orogenies, three being the most important, are found in the Tienshan, while four orogenies (Figure 16.5), of which the last is the most important, are found in the Nanshan. Roughly coinciding with the orogenies, intrusion of granitoid plutons took place, indicating that magmatism generally corresponds to important orogenies.
2. Submarine volcanic eruptions are also polycyclic. In the Tienshan, intermediate volcanics (andesites) prevail and are found in four cycles corresponding to four orogenic cycles. The Nanshan volcanics, generally basic in nature (tholeiites and spilites), can also be divided into four cycles corresponding to the four most important orogenic cycles.
3. Basic and ultrabasic rock complexes are likewise polycyclic. In the Nanshan they are best developed and divisible into five cycles, each corresponding to an orogenic cycle (Figure 16.5), of which the second cycle (E-O_1) is the most important. Basic and ultrabasic rocks, together with radiolarian chert, form three typical ophiolitic suites in three different ages (Z, E_2, O_1). On the contrary, ultrabasic rocks are poorly developed in the Tienshan, where ophiolites are absent.
4. Marine flyschoid sediments are well developed in both regions, and together with submarine volcanics form polycyclic volcano-flyschoid formations. However, they are andesite-flyschoid in the Tienshan and basaltic-flyschoid in the Nanshan. Continental molasse is developed in both regions at the close of the geosynclinal evolution. The Devonian molasse of the Nanshan is the most typical.

It is important to emphasize that polycyclicity of geosynclinal development must not be considered as simple repetitions of geological processes as some writers believe them to be, but rather they are processes of vectorial

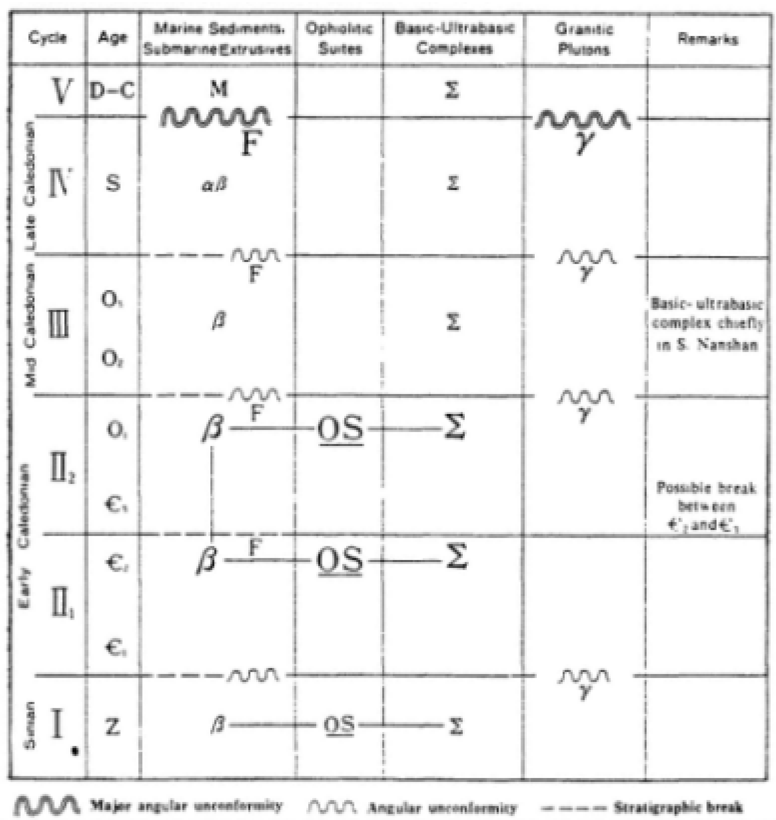


Figure 16.5
Diagram showing the polycyclic development of the Nashan eugeosyncline. β: basic submarine extrusives, mainly tholeiites and spilites; α: andesite; α^a: andesite having an acidic tendency; α^b: andesite having a basic tendency; ρ: rhyolite; F: flyschoid deposits, often alternating with β and α; M: molasse, divided into two stages, D_{1-2} and D_3, in Nanshan; Σ: basic and ultrabasic rocks, generally serpentinized; OS: ophiolitic suites, formed of serpentinites, basic extrusives (with pillow structures) and radiolarites; G: basic intrusives; γ: granitic plutons. The size of the letters representing the different rocks corresponds to the degree of their importance.

spiral-like evolution with a regular arrangement. In the Nanshan geosyncline, for instance, ophiolites began to develop in Cycle I, reaching their acme of development in Cycle II, especially in Subcycle II_2.* In Cycle III, although basic and ultrabasic rocks are present, no ophiolites were formed. In Cycle IV, they dwindled away. On the contrary, flyschoid sedimentation, poorly developed in Cycle I and Subcycle II_1, strengthened in Subcycle II_2 and Cycle III, and reached its acme of development in Cycle IV. Folding, faulting, and granitoid intrusion all played an important role in each cycle, but it is obvious that they possessed a tendency of increasing activity from Cycles II to III and reached their peak of development in Cycle IV. Cycle V indicates the close of geosynclinal sedimentation, when extensive molasse formed.

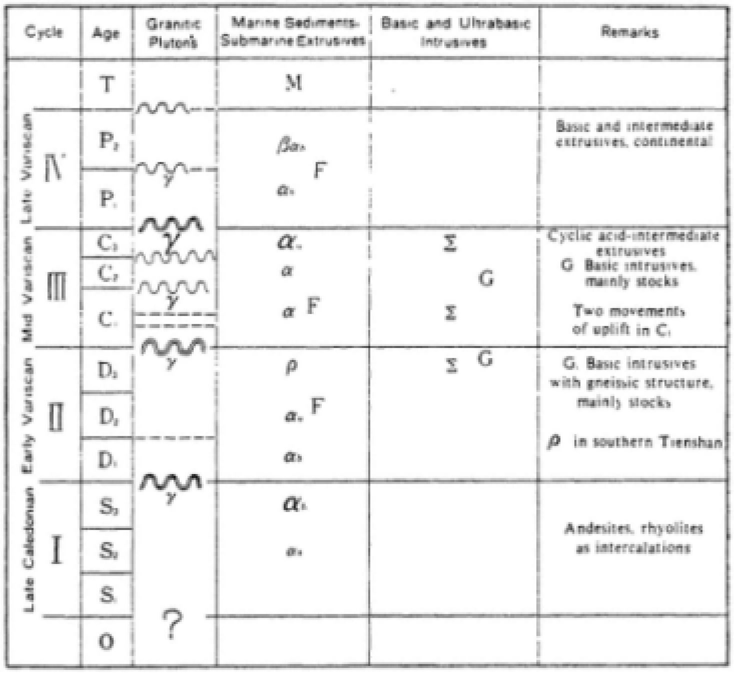


Figure 16.6
Diagram showing the polycyclic development of the Tienshan eugeosyncline. Symbols are as in Figure 16.5.

The polycyclic evolution toward a definite direction of the Tienshan geosyncline is even more prominent (Figure 16.6). There, granitoid magmatism was not important in Cycles I and II, but was greatly strengthened in Cycle III, and became best developed in the later part of Cycle III, i.e., in late Carboniferous time. Granitoid magmatism rapidly decreased in Cycle IV. Moreover, granitoid plutons of Cycles I and II show gneissic structures absent in those of Cycles III and IV. It must be pointed out that each of the five granitoid magmatisms is clearly connected with an orogeny, while the occurrence of large-scale granite batholiths is generally contemporaneous with principal orogenies. From the petrochemical point of view, granitoid intrusives of early Variscan abound in plagioclase granites and albite granites; those of middle Variscan are generally normal granites, while those of late Variscan are characterized by kali-granites and alaskites, and even syenites appear. In other words, the petrochemical characteristics of granitoid intrusives change from acid-intermediate to acid, then to acid-alkaline, and finally to alkaline.

From the above discussion, the author arrives at the preliminary conclusion that the Nanshan geosyncline is characterized by polycyclic basic extrusives (tholeiites and spilites), which, together with ultrabasic rocks, form polycyclic ophiolitic suites. Thus, the two types of geosynclines are quite different from each other in character.

Polycyclic geosynclinal development is also distinctly revealed in other geosynclines of China, among which the Tsinling geosyncline and the East Kunlun-Tangla geosyncline are typical (Huang *et al.*, 1974).

Many of the famous geosynclinal systems in the world, such as the Appalachian, the Cordilleran, the Uralian, and the Tasman geosynclines, are also characterized by polycyclic development. Consequently, polycyclic development of geosynclines is, without doubt, to be considered as the general rule.

For the sake of simplicity, a preliminary model for the

* The author divides an orogenic cycle into subcycles (see Figures 16.5 and 16.6).

development of geosynclinal fold belts as shown in Figure 16.7 is proposed. It can be seen from Figure 16.7 that the development of geosynclinal fold belts includes the following:

1. Early geosynclinal cycles, including Cycle I, Cycle II, and possibly Cycle III;
2. Principal geosynclinal cycles, including Cycles III and IV; and
3. Postgeosynclinal cycles, including Cycles I and II.

It must be noted that each geosynclinal cycle may include initial magmatism (ophiolites), synorogenic magmatism (granites), and subsequent magmatism (porphyries and andesites) but usually without final magmatism (plateau basalts). Moreover, flysch and molasse formations are also polycyclic, while the postgeosynclinal cycles are characterized by block-faulting accompanied by folding, again with polycyclic granites and polycyclic molasse (such as in Eastern China).

If we accept the theory of plate tectonics to interpret the origin and development of geosynclines, as quite a number of geoscientists are doing, we come to the conclusion that plate motions are likewise polycyclic in nature. This is particularly manifested in the case of the Tasman Geosyncline. Scheibner (1972) pictured the development of that geosyncline as a series of successively eastward-accreting continental blocks, subducted by a series of eastward-retreating oceanic crust. From his palinspastic maps, it appears clear that the Tasman fold belt was formed by polycyclic plate motions.

PRELIMINARY OBSERVATIONS ON PLATE TECTONICS IN CHINA

Tsangpo Depth Fracture Zone

Geological and geophysical data indicating the applicability of plate tectonics along the Tibetan part of the Himalayan geosyncline were collected and analyzed by Chinese geologists (Huang *et al*., 1974; Chang and Cheng, 1973). Recent observations in the Tsangpo Valley reveal the occurrence of abundant ophiolites, and to the east of Shigatze typical mélanges with exotic blocks of limestone containing Triassic and Jurassic fossils are found. It is thus probable that the Tsangpo depth-fracture zone is a subduction zone with the southern oceanic plate underthrusting the northern Asiatic plate. Such subducting activities might be polycyclic in nature.

Nanshan-Tsinling Region

The eugeosynclinal character of the northern Nanshan was described in previous articles (Wang and Liu, 1976; Huang *et al*., 1965, 1974), while its polycyclic development has already been stressed. The glaucophane schist zone chiefly consists of quartz-muscovite glaucophane schist and garnet-epidote glaucophane schist, and the majority of the tholeiites of the ophiolitic suites are characterized by very low K_2O contents (usually <<0.3 wt%). It is interesting to note that all of these features are similar to those of other paleogeosynclines in the world. The

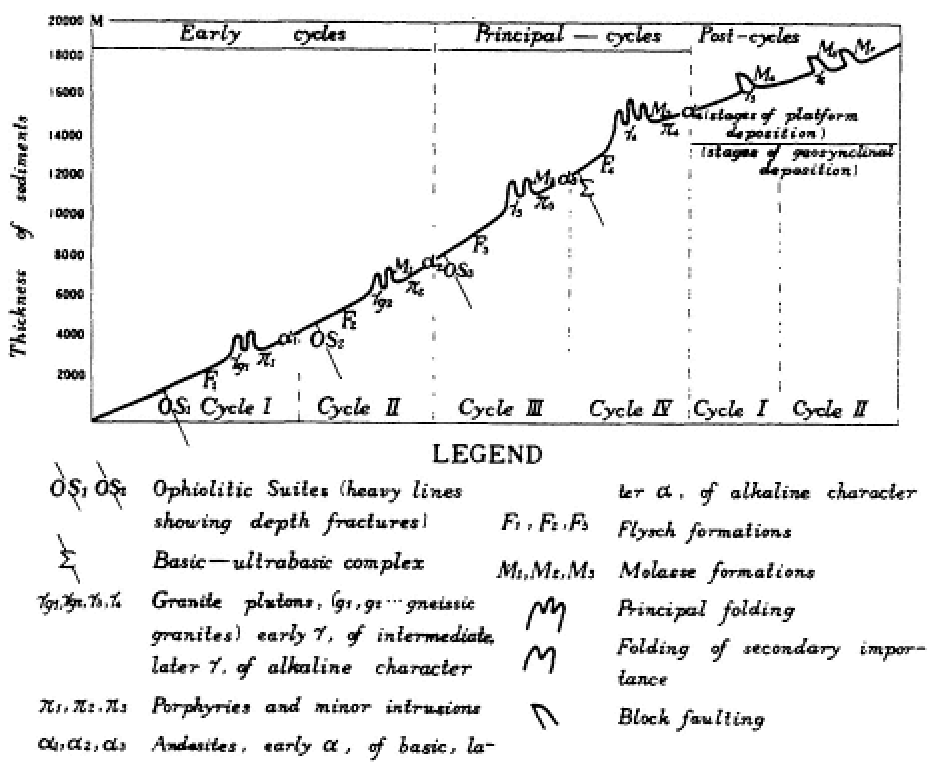


Figure 16.7
Diagram showing polycyclic development of geosynclinal foldbelts.

ophiolites are polycyclic, with successive motions being from the south to the north (Wang and Liu, 1976).

The Northern Tsinling geosyncline is the continuation of the Northern Nanshan. Observations show that the Caledonides of the Tsinling extend from south of Paochi to Nanyang Basin, probably including ophiolitic patches. Li (1975) advocates that the Tsinling is characterized by plate-tectonic interactions. It is possible that the eastern Tsinling fold belt was formed by the mutual approach between the Sino-Korean and the Yangtze paraplatforms.

Chinshakiang-Red River Depth-Fracture Zone

Recent mapping in the Chinshakiang-Red River region discloses evidence for paleoplate motions. South of Batang, in the western side of the upper Chinshakiang fracture zone, occur typical ophiolitic mélanges characterized by serpentinites, spilites with pillow structure, and various kinds of basic to ultrabasic rocks intercalated with radiolarian cherts, while exotic blocks of omphaciteeclogite; dillage-cinnamon stones; and particularly Devonian, Carboniferous, and Permian limestones are found in the matrix. In the eastern side of the same zone, wildflyschlike deposits characterized by a matrix of argillaceous and arenaceous rocks with various exotic blocks yield many fossils ranging in age from Silurian to Permian, which are, without exception, older than those in the matrix. Judging from all the facts observed so far, it is probable that a western oceanic plate approached and collided, from Permian to Late Triassic time, with the eastern suboceanic plate (this belongs to the Indosinian orogeny). Along the Red River valley, the existence of ophiolitic mélanges and glaucophane schists is also probable. Apparently, the Chinshakiang zone and the Red River zone belong to the same system of convergent plates.

Since in western Yunnan and northern Burma there occur a series of lithospheric fractures, i.e., Chinshakiang-Red River, Lantsangkiang, Nukiang, and Naga-Arakan Yoma, and since the principal ages of these fracture zones seem, respectively, to be Indosinian, early Yenshanian, late Yenshanian, and Himalayan, the hypothesis is suggested that polycyclic plate subductions happened from east to west in a successively retreating manner.

References

Aubouin, J. (1965). Geosynclines, in *Development in Geotectonics*, Elsevier, Amsterdam.

Beloussov, V. V. (1962). *Basic Problems in Geotectonics*, McGraw-Hill, New York, 809 pp.

Chang, C. F., and H. L. Cheng (1973). Some tectonic features of the Mt. Joimo Lungma area, southern Tibet, *Sci. Sin. 16*, 257-265.

Ch'eng, Y. C., F. D. Chung, and Y. C. Su (1973). The Chen-Tan series in northern and northwestern Chinese areas, *Acta Geol. Sin. 1973*, no. 1 (in Chinese).

Chinese Academy of Geological Sciences (1976). The Geological Map of the Peoples Republic of China (1:4,000,000), Map Publishers (in Chinese).

deSitter, L. U. (1964). *Structural Geology*, 2nd ed., McGraw-Hill, New York, 530 pp.

Fitch, T. J. (1972). Plate convergence, transcurrent faults and internal deformation adjacent to southeast Asia and the western Pacific, *J. Geophys. Res. 77*, 4432-4460.

Gansser, A. (1964). *Geology of the Himalayas*, Interscience, New York.

Gansser, A. (1966). The Indian Ocean and the Himalayas: a geological interpretation, *Eclogae Geol. Helv. 59*, 831-848.

Hsi, C. W., S. F. Feng, C. S. Li, H. P. Chen, K. T. Wen, C. J. Chang, and C. C. Hsiung (1974a). The background of the deep structure in the middle of the Northern China Plain and the Hsing-Ta earthquake, *Acta Geophys. Sin. 17*, no. 4 (in Chinese).

Hsi, C. W., S. F. Feng, C. S. Li, H. P. Chen, K. T. Wen, C. J. Chang, and C. C. Hsiung (1974b). The reflection wave of the deep earthquake in Chai-Ta-Mu Basin and crustal structure, *Acta Geophys. Sin. 17*, no. 2 (in Chinese).

Hsi, C. W., S. F. Feng, C. S. Li, H. P. Chen, K. T. Wen, C. J. Chang, and C. C. Hsiung (1975). The background of the deep structure in the middle of the Northern China Plain and the Hsing-Ta earthquake, *Acta Geophys. Sin. 18*, no. 3 (in Chinese).

Huang, T. K. (1945). On major tectonic forms of China, *Geol. Mem. Ser. A*, no. 20, National Geological Survey of China.

Huang, T. K. (1960). The main characteristics of the structure of China: preliminary conclusions, *Sci. Sin. 9*, 494-544.

Huang, T. K., and C. F. Chiang (1962). The primary study of the crust—rules developed from the polycyclic point of view, *Acta Geol. Sin. 40*, no. 1 (in Chinese).

Huang, T. K., C. K. Chang, C. M. Chang, and K. M. Chen (1965). On eugeosynclines and miogeosynclines and their polycyclic development, *Chin. Acad. Geol. Sci. Selec. Prof. Papers Reg. Tectonics, Sec. C*, Peking (in Chinese).

Huang, T. K., C. S. Jen, C. Chiang, C. M. Chang, and C. K. Chang (1974). The new knowledge of some characteristics for Chinese plate tectonics, *Acta Geol. Sin. 1974* (in Chinese).

Lee, J. S. (1939). *The Geology of China*, Thomas Murby & Co., London, 528 pp.

Li, C. Y. (1975). The primary analysis for the development of some parts of Chinese regional structure using plate tectonic theory, *Acta Geophys. Sin*. 18, no. 1 (in Chinese).

Molnar, P., and P. Tapponnier (1975). Cenozoic tectonics of Asia: effect of a continental collision, *Science 189*, 419-426.

Rittmann, A. (1960). *Vulkane und ihre Tatigkeit* 2. *umgearbeite-ten und erw., Aufl*., Enke, Stuttgart, 336 pp.

Scheibner, E. (1972). Actualistic models in tectonic mapping, *24th Sess. Int. Geol. Congr. Can*. 405-422.

Terman, M.J. (principle compiler) (1973). *Tectonic Map of China and Mongolia*, (1:5,000,000), Geol. Soc. Am.

van Bemmelen, R. W. (1949). *The Geology of Indonesia*, Government Printing Office, The Hague, 732 pp.

Wang, C., and H. Y. Liu (1976). The ancient oceanic crust in western China's Ch'i-Lien mountains and its plate tectonic meanings, *Geol. Sci. 1974*, no. 1 (in Chinese).

PLATE 11.1
Aromagnatic map of north-central United States (each color represents 200 gammas).