

THE STRUCTURE AND THERMAL TECTONICS OF PLANETARY LITHOSPHERES:

MID-OCEAN RIDGES AND LUNAR IMPACT BASINS

by

Steven Richard Bratt

B.S., The Pennsylvania State University
(1979)

SUBMITTED TO THE DEPARTMENT OF
EARTH, ATMOSPHERIC, AND PLANETARY SCIENCES
IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS
FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

November 1984

© Massachusetts Institute of Technology 1984

Signature of Author

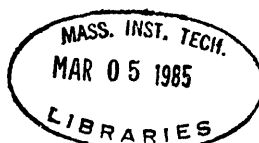
Department of Earth, Atmospheric, and Planetary Sciences

Certified by

Sean C. Solomon
Thesis Supervisor

Accepted by

Theodore R. Madden
Chairman, Department Committee



ARCHIVES

THE STRUCTURE AND THERMAL TECTONICS
OF PLANETARY LITHOSPHERES:
MID-OCEAN RIDGES AND LUNAR IMPACT BASINS

by

Steven Richard Bratt

Submitted to
the Department of Earth, Atmospheric, and Planetary Sciences
on November 21, 1984 in partial fulfillment of the
of the requirements for the degree of Doctor of Philosophy in
Geophysics

ABSTRACT

The structure and tectonics of planetary lithospheres reflect both their initial state and their subsequent evolution. This thesis is an investigation of several aspects of these relationships for terrestrial mid-ocean ridges and lunar impact basins. A common thread linking oceanic lithosphere and impact basins is that both are born hot and cool with time. The objectives of this thesis are to determine the structure of young oceanic lithosphere and large lunar basins, to interpret these structures in terms of the processes responsible for their formation, and to assess the contribution of cooling, thermal contraction, and thermal stress to the tectonics of oceanic and lunar lithosphere.

Seismograms recorded by a three-component ocean-bottom seismometer from two orthogonal refraction lines in the ROSE area of the East Pacific Rise at 11°20'N are used to constrain the structure of young oceanic crust and the nature of magma bodies beneath the rise axis. Converted shear waves are clearly observed on the horizontal seismometers for both lines, including paths crossing the rise axis. Neither a travel-time delay nor measurable attenuation is observed for shear or compressional waves which have passed beneath the rise crest at depths down to 4 km. On the basis of these observations, there can be no large crustal magma chamber beneath the rise at this latitude. The P wave velocity distribution, as inferred from travel-time and amplitude data, does not differ markedly from that of "normal" young oceanic crust. The shear velocity distribution, as inferred from the travel times of converted S arrivals, however, indicates a Poisson's ratio (0.32) that may be more typical of fractured crust in rise-axis regions. A combined consideration of the

results of this experiment and other observations of rise-crossing crustal phases elsewhere in the ROSE area leads to the conclusion that any magma within the crust beneath the rise axis of this region must be confined to narrow dikes or pipes or to small isolated bodies with a vertical thickness of less than about 1 km.

Over 500 seismograms collected by five ocean-bottom hydrophones are used to examine the structure and variability of 0.5-m.y.-old oceanic crust in the ROSE area. There are no significant differences among the first arrival travel times for 10 independent refraction lines along the isochron. However, line-to-line variability of an amplitude pattern signaling the transition from seismic layer 2 to layer 3 suggests that structural boundaries within the upper crust vary by hundreds of meters in depth over lateral distances of tens of kilometers within this 200 km segment of lithosphere. Based on the modeling of these amplitude patterns and on comparisons with the upper crust sampled at DSDP hole 504B and in ophiolite complexes, layer 2 can be divided into three units: 2A corresponds to extrusive volcanic rocks with a steep velocity gradient, 2B to extrusives with a near-zero velocity gradient, and 2C to a complex region of transition from extrusives to sheeted dikes. The total thickness of layer 2 varies by about 0.7 km over the length of the line. The variability of the layer 2 -layer 3 amplitude event provides direct evidence that the along-strike structure of oceanic crust in the ROSE area is definably heterogeneous and that the process of crustal accretion is likewise variable on the scale resolved by amplitude and travel-time information.

A series of exploratory models tests the hypothesis that thermoelastic stress is a significant contributor to the state of stress in young oceanic lithosphere (less than 35 m.y. in age). In support of this hypothesis is the concentration of seismicity in lithosphere younger than 15 m.y. where cooling rates are relatively high. Most recent near-ridge earthquakes occurred beneath the Moho but shallower than the depth of the 800°C isotherm. Thrust and strike-slip events dominate the shallower seismicity, while the focal mechanisms of the deeper events generally indicate normal faulting. To assess the importance of thermal stress in the total stress field indicated by these earthquake characteristics, the lithosphere is modeled as a half-space in which the elastic response to cooling is computed. Normal faulting is predicted throughout the lithosphere in models based on simple half-space cooling. Models in which a cooled surface layer is used to simulate the effect of hydrothermal circulation on shallow thermal structure predict stresses that match the locations of both thrust and normal faulting observed near mid-ocean ridges. The addition of a "ridge push" stress to the computed thermal stress field enhances the likelihood of shallow thrust faulting at ages greater than about 15 m.y., but contributes little to the stress field at younger ages. These simple half-space models suggest that thermoelastic stress can play a significant role in the tectonics of young oceanic lithosphere.

Models for the structure of the crust and upper mantle beneath lunar impact basins are derived from an inversion of gravity and topographic data from the nearside of the Moon. All basin models display a thinner crust and an elevated Moho beneath the central basin region compared to surrounding areas, a signature of the processes of basin excavation and mantle uplift during collapse of the transient cavity. A general decrease in the magnitude of apparent uplift of mantle material with increasing basin age is attributable to enhanced rates of ductile flow of crustal material early in lunar history when crustal temperatures were relatively high. The more relaxed topographic and Moho relief associated with older basins on the central nearside may, in particular, be at least partly a consequence of the extensive subsurface heating associated with the formation of the large Procellarum basin. The deep structure of the youngest basins constrains the geometry of the cavity of excavation and the amount of crustal material ejected beyond the basin rim. From the volumes of the topographic basin, of mare basalt fill, and of uplifted mantle material, the volume of crustal material ejected beyond the basin rim for an Orientale-size event was on the order of 10^7 km³. A near constant thickness of non-mare crustal material beneath the central regions of young basins of various diameters and pre-impact crustal thicknesses suggests that the transient cavity excavated to at least the base of the crust for the largest basins; significant excavation into the mantle may have been impeded by an abrupt increase in strength at the lunar Moho.

Thermal stress may have been a significant contributor to the topography and tectonics of large lunar impact basins. Heat converted from impact kinetic energy and contributed from the uplift of isotherms during early cavity modification are important components in the energy budget of a newly-formed basin. That subsequent cooling may have been an important factor in the tectonic history of lunar basins is partially supported in the Orientale basin by a deep central depression and a surrounding region of extensive fissuring. The significance of thermal stress to the formation of these features is tested by examining the response of an elastic half-space to cooling of the sub-basin region. All models predict subsidence of the basin floor and a state of stress in which fissuring is likely. In addition, the rates of cooling and of accumulation of thermal stress are consistent with the inferred timing of fissure formation in Orientale. The sensitivity of the displacement and stress fields predicted by our models to the distribution of initial temperatures allows us to use the observed topography and tectonics to place approximate bounds on the quantity and distribution of impact heat emplaced during basin formation. The impact heat deposited beneath the Orientale basin was probably between 1 and 6×10^{32} erg. It is likely that most of this heat was concentrated within a distance of 100 km from the point of impact.

Thesis Supervisor: Sean C. Solomon, Professor of Geophysics

ACKNOWLEDGEMENTS

My family deserves the first word of thanks. It was they who gave me love during the formative years and always encouraged me to follow my dreams. Many of my high school teachers, coaches and friends directly contributed to my development as an individual and as a scientist. At Penn State, Professors Graham, Alexander, Lavin, and Dachtler fed my curiosity about processes which mold the Earth and planets and convinced me that getting paid to attend graduate school was not such a bad idea.

There have been so many people in Cambridge and at M.I.T. who have made my 5 years here both educational and fun. The runners of the Cambridge Sports Union motivated me to try a marathon or two thus bolstering my confidence at a time when my biggest academic achievements included inverting 3×3 matrices and logging onto Multics. I thank Steve Roecker, Lynn Hall, Paul Huang, Bob Nowack, Mark Murray, Dan Davis, Craig Jones, Arthur Cheng, and Carl Godkin for periodic spiritual and intellectual nourishment. I thank Steve, Joey, Toscanini, and the Felafel King for nourishment of the edible sort. Intramurals, especially softball, have generated a camaraderie that will endure past M.I.T. Three of the best friends (as well as best housemates, officemates and scientists) that anyone could ask for just happen to anchor most of our departmental teams: Scott Phillips, Dave Olgaard, and Roger Buck. Darby Dyar, Kaye Shedlock, and Lind Gee have made life around the Institute a joy and have been great sources of warmth and support over the years.

People, agencies, and ships have contributed greatly to the work contained in this thesis. Anne Trehu, Paul Spudich, Jeff Karson, John Orcutt, and Jan Garmany offered words of wisdom on the work of Chapters 2 and 3. I am grateful to Woods Hole's Dicky Allison, for her expertise in data processing, and Mike Purdy, for contributing both

his thoughts and elbow grease to the work of Chapter 3. I'd also like to thank the captains, crews, technical personnel, and scientists involved in Project ROSE. The work of Chapter 4 was largely motivated by Eric Bergman's compilation of the characteristics of near-ridge earthquakes. The lunar gravity and topographic data sets used in Chapter 5 were generously provided by Bill Sjogren and Bruce Bills, respectively. Jim Head has contributed planetological insight on countless occasions and Cliff Thurber has received desperate phone calls from me for geophysical advice more times than I can recount. I thank both of them for their help. The bills for this thesis were paid by NASA grants NSG-7081 and NSG-7297, and contract NAS 5-27339, by ONR contracts N00014-80-C-0273 and N00014-82-C-0019, and by NSF grant EAR-815908.

Much thanks to the members of my thesis committee, Bill Brace, Marcia McNutt, Mike Purdy, and Rob Comer, for volunteering the time and effort required to examine a thesis and its author. I also thank Bill for running adventures in Northern New England where the temperatures stay down but the topography only goes up. I cannot express adequately enough my appreciation of Jan Barbaro, for her word-processing wizardry, and Sharon Feldstein, for her funding finesse. Other invaluable assistance in the preparation of this manuscript has come from Dorthy Frank, Jeannie Kahl and Amy Sterling-Bratt.

I consider myself lucky to have had Sean Solomon as an advisor. Not only has he allowed me to freely pursue both seismological and planetary problems, but he possesses the expertise to have guided me well in both fields. His scientific qualities have become my criteria for excellence.

Finally, I thank my wife Amy for her unrequited support and love during the last two years and especially during the thesis crunch.

I dedicate this work to her.

TABLE OF CONTENTS

	<u>page</u>
ABSTRACT.....	ii
ACKNOWLEDGEMENTS.....	v
CHAPTER 1. INTRODUCTION.....	1
References.....	7
CHAPTER 2. COMPRESSIONAL AND SHEAR WAVE STRUCTURE OF THE EAST PACIFIC RISE AT 11°20'N: CONSTRAINTS FROM THREE-COMPONENT OCEAN BOTTOM SEISMOMETER DATA.....	10
Introduction.....	11
Experiment Description.....	16
Data Processing.....	19
Compressional Velocity Structure.....	21
The rise-crossing line.....	21
The rise-flanking line.....	26
Effects of a Sediment layer.....	27
The Nature of Refracted Shear Energy.....	30
The rise-crossing line.....	32
The rise-flanking line.....	36
Shear Velocity Structure and Attenuation Beneath the Rise.....	39
Discussion.....	43
Conclusions.....	47
References.....	50
Figure Captions.....	57
Figures.....	62
CHAPTER 3. STRUCTURE AND VARIABILITY OF OCEANIC CRUST ON THE FLANKS OF THE EAST PACIFIC RISE BETWEEN 11° AND 13°N.....	86
Introduction.....	87
Seismic Scale.....	89
The Experiment.....	91
Data Reduction and Uncertainty.....	93
Travel Time Interpretation.....	98
Amplitude Interpretation.....	101
Discussion.....	113
Summary and Conclusions.....	124

	<u>page</u>
References.....	129
Figure Captions.....	134
Tables.....	139
Figures.....	141
 CHAPTER 4. THERMOELASTIC STRESS: HOW IMPORTANT AS A CAUSE OF EARTHQUAKES IN YOUNG OCEANIC LITHOSPHERE?.....	 157
Introduction.....	158
Characteristics of Near-Ridge Earthquakes.....	159
Sources of Stress in Oceanic Lithosphere.....	160
Previous Models of Thermal Stress.....	162
Calculation of Thermal Stress.....	164
Temperature structure.....	164
Thermoelastic displacement potentials.....	166
Thermal stress.....	169
Assessment of procedure.....	172
Thermal Stress Models.....	174
Halfspace cooling.....	176
Effects of a hydrothermally cooled layer.....	178
Effects of a blocking temperature.....	180
Increasing the stress accumulation time	182
Contribution of ridge push.....	183
Discussion.....	185
Conclusions.....	187
References.....	189
Figure Captions.....	195
Tables.....	198
Figures.....	200
 CHAPTER 5. THE DEEP STRUCTURE OF LUNAR BASINS: IMPLICATIONS FOR BASIN FORMATION AND MODIFICATION.....	 212
Introduction.....	213
Procedure.....	216
Computation of gravity anomalies.....	217
Constraints and assumptions.....	219
Inversion for crustal structure.....	221
Gravity and Topographic Data.....	224
Basin Structural Models.....	230
Geologic Implications.....	237
Early cavity geometry.....	237
Basin modification.....	243
Summary and Conclusions.....	247
References.....	250

	<u>page</u>
Figure Captions.....	257
Table.....	261
Figures.....	262
 CHAPTER 6. THERMAL STRESS AND THE TECTONIC EVOLUTION OF LUNAR IMPACT BASINS.....	 274
Introduction.....	275
Geologic Observations: Orientale.....	276
Thermal Evolution.....	280
Thermal Displacement and Stress.....	281
Assessment of Procedure.....	285
Initial Thermal Structure.....	287
Impact heating.....	288
Isotherm uplift.....	292
Adopted Physical Parameters.....	293
Thermo-Tectonic Models: Orientale.....	295
Isotherm uplift.....	295
Impact heating.....	301
Effect of a blocking temperature.....	306
Discussion.....	309
Conclusions.....	312
References.....	315
Figure Captions.....	322
Tables.....	326
Figures.....	328
 CHAPTER 7. CONCLUSIONS.....	 346
References.....	351
 APPENDIX A MASS DISK PARAMETERS AND TRIAXIAL GRAVITY COEFFICIENTS USED TO COMPUTE LUNAR FREE- AIR GRAVITY ANOMALY.....	 352
References.....	354
Table.....	355
 APPENDIX B 5° by 5° AVERAGES OF LUNAR TOPOGRAPHY.....	 364
References.....	365
Table.....	366
 APPENDIX C THE POTENTIAL OF A THIN DISK.....	 383
 BIOGRAPHICAL NOTE.....	 386

Chapter 1

INTRODUCTION

Volcanism, impact cratering, and tectonism are three geologic processes which have at one time or another been important in the shaping of the surfaces of all the terrestrial planets. This thesis examines aspects of all three of these processes, including the structure of volcanically-emplaced crust near a terrestrial mid-ocean ridge, the deep structure of impact basins on the Moon, and a process potentially important to the tectonics of the lithosphere in both settings: thermal stress.

The most extensive zone of active volcanism in the inner solar system is located along the Earth's mid-ocean ridges. This essentially continuous chain of ridges and rises extends for over 50,000 km beneath the oceans. It is here that over 70% of the terrestrial lithosphere has been accreted and structurally ordered. Though the structural heterogeneity of oceanic crust and upper mantle is definable, the seismically measured depth to any structural horizon probably varies by at most a few kilometers within lithosphere of a given age [e.g., Spudich and Orcutt, 1980].

On the Moon, impact cratering has been the most prominent sculptor of the outer shell of that one-plate planet. The largest craters, called impact basins, exceed hundreds of kilometers in diameter and dominate our naked-eye view of the Moon as seen from the Earth. Impact basins were the center of volcanism and tectonic activity early in lunar history [Solomon

and Head, 1979; Solomon et al., 1982]. The shaping of the lunar lithosphere bore no resemblance to the relatively steady accretion of oceanic plates seen on our planet today. The deep structure of the Moon's lithosphere is largely a product of an intense and random period of bombardment that ended almost 4 b.y. ago. Gravity data over lunar basins and adjacent terrain suggest variability in crustal thickness of tens of kilometers over lateral distances of 100 km [Bills and Ferrari, 1977].

A key ingredient in the creation of both oceanic lithosphere and lunar impact basins is heat. The loss of this heat during lithospheric cooling can significantly affect the structural and tectonic evolution of crust and upper mantle. A first-order consequence of oceanic heat loss is the observed subsidence of the seafloor with increasing age [e.g., Parsons and Sclater, 1977]. That the amount of impact kinetic energy converted to heat during the formation of a single 600-km-diameter basin [$\sim 10^{32}$ erg, O'Keefe and Ahrens, 1975] is probably at least 10^4 times the annual loss of heat through the ocean floor [$\sim 10^{28}$ erg, Sclater et al., 1980] attests to the potential importance of cooling and associated thermal stress to the modification of impact basins.

In this thesis, I use seismic, gravity and topographic data to examine the structure of the lithosphere near the East Pacific Rise (Chapters 2, 3 and 4) and beneath impact basins on the lunar nearside (Chapters 5 and 6). The structures determined provide important information on the processes

responsible for the formation and modification of the respective lithospheres. Finally, I assess the contribution of cooling, thermal contraction and thermal stress to the tectonic history of lithosphere in the vicinity of terrestrial mid-ocean ridges and lunar impact basins.

Chapter 2 presents an investigation of the structure of crust less than 0.5 m.y. old at 11°20'N on the East Pacific Rise. The use of a three-component ocean-bottom seismometer (OBS) to record refraction seismic data from lines both crossing and flanking the rise facilitates the analysis of both compressional and converted-shear arrivals. In addition to determining the P- and S-wave velocity structure of the youngest oceanic crust, I appraise the efficiency of shear wave propagation beneath the rise axis as a test for the existence of large magma bodies within the crust. Comparison of these results with refraction data collected elsewhere on the rise allows the degree of heterogeneity in the process of crustal accretion to be explored. This work has been recently published [Bratt and Solomon, 1984].

The structure and variability along a 200-km-long line of 0.5-m.y.-old crust just north of the OBS is examined in Chapter 3. Over 500 seismograms recorded by 5 identical ocean-bottom hydrophones comprise a large, uniform data set with which arrival-time and amplitude data can be interpreted in terms of structure. Also, the variability within the data set provides clues to the heterogeneity of the shallow lithosphere and variability in the processes responsible for its creation.

Comparison of the results of this analysis with observations of the structure of ophiolite complexes [e.g., Casey et al., 1981; Christensen and Salisbury, 1975] and in Deep Sea Drilling Project hole 504B [Anderson et al., 1982; Becker et al., 1982] suggests geologic analogs to the seismic structural units in young oceanic crust. Also, comparison of the P-wave data recorded on 0.5 m.y. old crust with data from a similar study on 140 m.y. old crust [Purdy, 1983] gives additional insight into the evolution of lithosphere with time. Chapter 3 has also recently appeared in published form [Bratt and Purdy, 1984].

A well known product of the aging and cooling of oceanic lithosphere is thermal subsidence [e.g., Parsons and Sclater, 1977]. In Chapter 4, I examine the possibility that cooling and contraction of the lithosphere also produces a thermo-elastic stress field important to the state of stress in young lithosphere. This study has been motivated by recent determinations of the focal mechanisms and depths of near-ridge earthquakes [e.g., Bergman and Solomon, 1984] which suggest a concentration of seismicity in lithosphere younger than 15 m.y., where cooling rates are relatively high. To assess the importance of thermal stress to near-ridge seismicity, the lithosphere is modeled as a half-space in which the elastic response [Mindlin and Cheng, 1950] to cooling is computed. The thermal stress field predicted by standard half-space cooling and by cooling models which simulate the effects of hydrothermal circulation are calculated and compared to the state of stress inferred from the earthquake record. A paper

reporting on the results of this analysis has been submitted for publication [Bratt et al., 1984].

The deep structure of lunar impact basins offers information on the processes of basin formation and modification not available from observations of surface morphometry and morphology. Models for the structure of the crust and upper mantle are derived in Chapter 5 from an inversion of gravity and topographic data from the lunar nearside. Other global models lacking the resolution of this analysis [e.g., Thurber and Solomon, 1978] suggest that lunar basins are characterized by generally thinned crust, a remnant of crater excavation and mantle uplift during basin formation. Comparison of the preserved mantle uplift beneath basins as a function of basin age furnishes insight into changes in the processes of basin formation and modification as the Moon's elastic lithosphere cooled and thickened with time. The amount of crustal thinning displayed beneath the youngest basins can also provide constraints on the volume of material excavated and ejected beyond the basin rim, a quantity that for the largest basins is presently uncertain by an order of magnitude [Head et al., 1975]. The analysis and results of this chapter are in press [Bratt et al., 1984b].

Basin formation focusses large quantities of heat into the shallow lithosphere beneath the point of impact [e.g., O'Keefe and Ahrens, 1975]. Chapter 6 is an assessment of the contribution of cooling, subsidence, and thermal stress to the topography and tectonics of impact basins. The two primary

components of a basin's initial energy budget are heat converted from the kinetic energy of the projectile and the uplift of isotherms during collapse of the excavated cavity and formation of the basin. The quantity and distribution of the first component is highly uncertain. Isotherm uplift is constrained by the structural models of Chapter 5. The importance of thermal stress as a tectonic process in impact basins is examined by modeling the response of an elastic half-space to cooling of a cylindrically symmetric initial temperature field. The modeling procedure differs from that of Chapter 4 only in the symmetry of the changing temperature field. The quantity and distribution of impact heating is varied so as to match the relief of the central depression, the location of a band of fissures, and the timing of deformation within the relatively young and unmodified Orientale basin [Church et al., 1982]. A sensitivity of the models presented here to the quantity and distribution of initial heat allows the observed topography and tectonics to be used as constraints on the quantity and distribution of impact heat emplaced during basin formation.

In Chapter 7, I discuss some general conclusions of this thesis, address questions generated by the analyses contained in this work, and suggest future work designed to help answer these questions.

REFERENCES

- Anderson, R.N., et al., DSDP hole 504B, the first reference section over 1 km through layer 2 of the oceanic crust, Nature, 300, 589-594, 1982.
- Becker, K., et al., In situ electric resistivity and bulk porosity of the oceanic crust, Costa Rica Rift, Nature, 300, 594-598, 1982.
- Bergman, E.A., and S.C. Solomon, Source mechanisms of earthquakes near mid-ocean ridges from body waveform inversion: Implications for the early evolution of oceanic lithosphere, J. Geophys. Res., in press, 1984.
- Bills, B.G., and A.J. Ferrari, A lunar density model consistent with topography, gravitational, librational, and seismic data, J. Geophys. Res., 82, 1306-1314, 1977.
- Bratt, S.R., E.A. Bergman, and S.C. Solomon, Thermoelastic stress: How important as a cause of earthquakes in young oceanic lithosphere?, J. Geophys. Res., submitted, 1984a.
- Bratt, S.R., and G.M. Purdy, Structure and variability of oceanic crust on the flanks of the East Pacific Rise between 11° and 13°N, J. Geophys. Res., 89, 6111-6125, 1984.
- Bratt, S.R., and S.C. Solomon, Compressional and shear wave structure of East Pacific Rise at 11°20'N: Constraints from three-component ocean-bottom seismometer data, J. Geophys. Res., 89, 6095-6110, 1984.
- Bratt, S.R., S.C. Solomon, J.W. Head, and C.H. Thurber, The deep structure of lunar basins: Implications for basin

formation and modification, J. Geophys. Res., in press, 1984b.

Casey, J.F., J.F. Dewey, P.J. Fox, J.A. Karson, and E. Rosencranz, Heterogeneous nature of oceanic crust and upper mantle: A perspective from the Bay of Islands ophiolite complex, in The Sea, vol. 7, edited by C. Emiliani, pp. 305-338, Wiley-Interscience, New York, 1981.

Christensen, N.I., and M.H. Salisbury, Structure and constitution of the lower oceanic crust, Rev. Geophys. Space Phys., 13, 57-86, 1975.

Church, S., J.W. Head, and S.C. Solomon, Multi-ringed basin interiors: Structure and early evolution of Orientale (abstract), Lunar Planet. Sci., 13, 98-99, 1982.

Head, J.W., M. Settle, and R.S. Stein, Volume of material ejected from major lunar basins and implications for the depth of excavation of lunar samples, Proc. Lunar Sci. Conf. 6th, 2805-2829, 1975.

Mindlin, R.D., and D.H. Cheng, Thermoelastic stress in the semi-infinite solid, J. Appl. Phys., 21, 931-933, 1950.

O'Keefe, J.D., and T.J. Ahrens, Shock effects from a large impact on the moon, Proc. Lunar Sci. Conf. 6th, 2831-2844, 1975.

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

Purdy, G.M., The seismic structure of 140 m.y. old crust in the

western central Atlantic Ocean, Geophys. J. R. Astron. Soc., 72, 115-138, 1983.

Sclater, J.G., C. Jaupart, and D. Galson, The heat flow through oceanic and continental crust and the heat loss of the earth, Rev. Geophys. Space Phys., 18, 289-311, 1980.

Solomon, S.C., R.P. Comer, and J.W. Head, The evolution of impact basins: Viscous relaxation of topographic relief, J. Geophys. Res., 87, 3975-3992, 1982.

Solomon, S.C., and J.W. Head, Vertical movement in mare basins: Relation to mare emplacement, basin tectonics, and lunar thermal history, J. Geophys. Res., 84, 1667-1682, 1979.

Spudich, P., and J. Orcutt, A new look at the seismic velocity structure of the oceanic crust, Rev. Geophys. Space Phys., 18, 627-645, 1980.

Thurber, C.H., and S.C. Solomon, An assessment of crustal thickness variations on the lunar near side: Models, uncertainties, and implications for crustal differentiation, Proc. Lunar Planet. Sci. Conf. 9th, 3481-3497, 1978.

Chapter 2

COMPRESSIONAL AND SHEAR WAVE STRUCTURE OF THE
EAST PACIFIC RISE AT 11°20'N:
CONSTRAINTS FROM THREE-COMPONENT OCEAN-BOTTOM SEISMOMETER DATA

INTRODUCTION

Of substantial interest to marine seismologists during the past two decades has been the structure of the lithosphere at the locus of plate creation: the mid-ocean ridge or rise. It is there that a systematic compositional order develops in oceanic crust and upper mantle. Most of the currently accepted models for the formation of oceanic crust require that magmas originating in the mantle first undergo some degree of differentiation in a crustal magma chamber before material is accreted onto neighboring young crust [e.g., Cann, 1974]. Many recent seismic experiments have focused on testing for the presence of such proposed bodies of magma or partial melt beneath mid-ocean ridge axes as well as on delineating the velocity structure of young oceanic lithosphere. Because theoretical models for viscous dissipation of seismic energy in a partially molten material [Walsh, 1968, 1969] predict that the effects of melt on the velocity and amplitude of a passing seismic wave are much greater in shear than in compression, shear wave propagation can be a particularly sensitive indicator of melt beneath a mid-ocean ridge. Few observations of shear wave propagation at crustal depths beneath the axis of a ridge, however, have been reported. In this chapter, we present an analysis of data from two seismic refraction lines oriented parallel to and across the East Pacific Rise and recorded by a three-component ocean-bottom seismometer (OBS) during the Rivera Ocean Seismic Experiment, or Project ROSE [Ewing and Meyer, 1982]. The data allow us to identify

refracted shear energy, to test directly for the existence of a crustal magma chamber, and to investigate more completely than would be possible with only single-component data the structure of the crust beneath and adjacent to the East Pacific Rise.

Seismic surveys conducted on the Mid-Atlantic Ridge (MAR) and the East Pacific Rise (EPR) have led to the view that while crustal magma chambers may be steady-state features beneath intermediate to fast spreading ridges, they are intermittent structures along slow spreading ridges [e.g., Macdonald, 1982; Sanford and Einarsson, 1982]. From observations of converted shear waves that had passed beneath the median valley of the MAR at crustal depths at latitudes 37° and 45°N , Fowler [1976, 1978] concluded that no crustal magma chamber could be resolved beneath the slowly spreading ridge in either survey area. Efficient propagation of crustal compressional energy beneath the crest of the MAR led Steinmetz et al. [1977] to suggest that no crustal magma body was present beneath the ridge axis at 40°N . Microearthquakes have been shown to occur throughout the crustal column beneath the MAR axial valley at 45°N [Lilwall et al., 1978] and 23°N [Toomey et al., 1982], probably precluding crustal magma chambers in both regions. In fact, we know of no studies which have convincingly demonstrated the presence of a shallow magma body at any location beneath the MAR. Current models for plate formation at slow spreading rates based on thermal constraints [Sleep, 1975] and on seismological, thermal, petrological, and geochemical data [Nisbet and Fowler, 1978] are consistent with these findings

and predict that any crustal magma resides in regions too narrow to be resolved by conventional seismic refraction investigations. Several studies, it should be noted, have resolved zones of low velocity and high attenuation in the upper mantle beneath the MAR [e.g., Solomon and Julian, 1974; Forsyth, 1977; Steinmetz et al., 1977; Fowler, 1978], but such regions need not be related to crustal magma chambers.

In contrast to the results from experiments on the Mid-Atlantic Ridge, travel-time delays and attenuated compressional arrivals seen in seismic refraction data from several locations along the East Pacific Rise [e.g., Rosendahl et al., 1976; McClain and Lewis, 1980] have been interpreted as evidence for low velocity zones, possibly magma chambers, within the crust beneath the rise crest. Reid et al. [1977] noted delayed first arrivals from explosive shots and attenuated shear arrivals from earthquakes when wave paths from either type of source to a vertical-component ocean bottom seismometer passed obliquely beneath the EPR at 21°N. They suggested partial melting at a depth of about 2.5 km beneath the ridge as an explanation of their observations. Orcutt et al. [1976], on the basis of refraction data, and Herron et al. [1978], on the basis of a multi-channel reflection profile, both concluded that a magma chamber was present at about 2 km depth beneath the rise crest at 9°N. The proposed magma chamber along this section of rise is about 5-10 km wide, and its roof deepens slightly with distance from the axis [Rosendahl et al., 1976; Herron et al., 1978, 1980; Hale et

al., 1982]. Lewis and Garmany [1982], working with vertical-component ocean bottom seismometer data collected between 11° and 13°N on the East Pacific Rise, cited amplitude attenuation and travel-time offsets for waves crossing beneath the rise as possible indicators of an axial low-velocity zone less than 2 km wide and about 3 km deep. Most of the geophysical evidence gathered to date support the thermal models of Sleep [1975], which predict that a steady-state magma chamber should exist beneath intermediate and fast-spreading ridges such as the East Pacific Rise.

With the exception of the work of Reid et al. [1977], no published seismic studies in the Pacific have examined shear propagation beneath the EPR as a means of testing for the existence of molten material in the crust, and none have employed three-component ocean-bottom instruments to resolve more clearly the later arrivals such as converted shear waves. This chapter presents an analysis of two orthogonal seismic refraction lines recorded by a three-component OBS located 25 km to the east of the EPR axis. One 45 km-long line crosses the rise crest; a 40 km-long line parallel to the rise axis is used as a basis for comparison with the rise-crossing line. The various phases studied and the nomenclature used in the text are illustrated in Figure 1.

Briefly, our analysis of the two refraction data sets consists of the following stages:

(1) We invert the travel times of first arrivals (the phases marked P in Figure 1) to obtain the compressional

velocity structure of the upper crust beneath and adjacent to the rise crest. Amplitude and waveform modelling of crustal refractions (P) and Moho reflections (PmP) are used to constrain further the structure of the lower crust and Moho along the rise-crossing line.

(2) The reverberatory character of the seismograms, the steep incidence angles of the first arrivals, and the presence of a converted shear phase (Ps) suggest that the OBS is sitting on well sedimented crust. We use the difference between the Ps and P arrival times to estimate the thickness of the sediments beneath the OBS.

(3) Converted shear refractions (Ss) arrive with large amplitudes from all shots on the rise-crossing line, including shots with wave paths passing beneath the rise crest. Refracted shear waves arriving at the OBS from shots along the rise-flanking line, in contrast, are considerably less energetic. From the arrival times and amplitudes of these converted shear waves for the rise crossing line, we derive the shear velocity structure of the crust and we examine the implications for a magma chamber beneath the rise axis.

(4) Finally, we compare the information on travel times, amplitudes, and particle motions obtained from this analysis with independent observations made elsewhere within the ROSE area, and we attempt to synthesize these findings into a consistent view of the seismic structure of the region.

EXPERIMENT DESCRIPTION

From January through March 1979, groups from twelve institutions, including the Massachusetts Institute of Technology (MIT), cooperated in a study of the structure and evolution of young oceanic lithosphere and the structure and tectonics of an active transform [Ewing and Meyer, 1982]. Project ROSE brought together 5 research vessels, 67 ocean-bottom seismometers and hydrophones, and several towed hydrophone arrays in a two-part study located between 11° and 16°N latitude near the East Pacific Rise (Figure 2). During the first experiment phase, refraction data were recorded along lines parallel and perpendicular to the rise crest.

In this chapter we examine the seismic data collected along two lines which intersected near the site of a three-component MIT OBS (Figure 3). Line 6 is perpendicular to the strike of the rise and crosses the EPR axis at 11°20'N latitude. Line 4 is parallel to and 25 km east of the rise axis. The local spreading rate is 104 mm/yr (full rate) with an azimuth of N80°E [Klitgord and Mammerickx, 1982]. The MIT instrument was located within the Brunhes magnetic polarity epoch and sits on crust that is 0.5 m.y. in age. Bathymetric data along line 6 (Figure 4a) resolve no signature of an axial trough nor of the severe topography common along slower spreading ridges. However, there is 550 m of seafloor relief between the MIT OBS and the rise crest 25 km to the west. The dominant bathymetric feature of line 4 (Figure 4b) is a seamount with 600 m of relief located 14 km to the north of the OBS.

A total of 29 seismograms from shots along line 6 were picked for use in this study on the basis of clarity of first or later arrivals. Included in this set are records from 6 large (217 kg) charges spaced 10 km apart and 23 small (11 kg) charges spaced 1.5 km apart. Of the 20 records examined from line 4, 4 are from large charges (217 kg) shot at irregular intervals and the remainder are from small charges (11 kg) spaced at 1.5 km intervals.

The MIT OBS is a free-fall, pop-up instrument designed primarily for recording earthquakes [Mattaboni and Solomon, 1977; Duschenes et al., 1981]. The three orthogonal geophones are housed in a separate pressure vessel which is deployed from the main instrument package after the instrument comes to rest on the seabed. The geophones are suspended from a gimbal joint to insure that the vertical component remains nearly vertical for instrument tilts up to 30°. Events are recorded digitally for ease of later processing and to ensure adequate signal dynamic range. The MIT OBS utilizes a simple event-detector scheme with an 18.3 s memory to allow for the preservation of first motion information from earthquakes and from most refraction shots. The first arrival from an explosion will not be recorded if the water wave rather than a crustal phase triggers the event detector and if the shot-receiver offset is sufficiently large so that the water wave arrives more than 18 s after the first refracted arrival. As a result, first arrivals from small charges at ranges greater than 25 km and from large charges at ranges greater than about 40 km were

usually not preserved.

A receiver planted on the seafloor has a distinct advantage over ocean-surface receivers because it can provide clear recording of both compressional and shear energy as well as better resolution of shallow crustal structure. Inadequate coupling of the geophones to the mechanically weak sediments, however, may be a cause of the reverberatory character of some OBS records [Sutton et al., 1981a]. Poor coupling causes signal magnification at one or more resonant frequencies that depend on sediment properties and on instrument geometry and mass [Sutton et al., 1981b; Zelikovitz and Prothero, 1981; Garmany, 1984]. Because the instrument response is relatively undistorted at frequencies below and attenuated at frequencies above such resonant frequencies, poor coupling can act as a low-pass filter operating on the incoming signal.

Transient tests conducted to measure OBS instrument responses at Lopez Island, Washington [Sutton et al., 1981a] and Woods Hole, Massachusetts [Tréhu and Solomon, 1981] have indicated that the MIT instrument records with relatively little noise and resonant distortion, but the resonant frequency for the vertical geophone can fall in the range typically used for marine refraction work (2-15 Hz) if the OBS is situated on sediment with a low shear modulus and high Poisson's ratio. In addition, the horizontal geophones have a resonant frequency that is lower than that of the vertical geophone and are therefore even more prone to site-related

filtering effects when located on soft sediments. At Lopez Island, where the sediments were perhaps slightly more consolidated than typical deep sea muds [Sutton et al., 1981a], the resonances of the MIT OBS vertical and horizontal geophones were at about 22 Hz and 6-10 Hz, respectively [Tréhu and Solomon, 1981]. If sediments similar to or less consolidated than those at Lopez Island existed under the ROSE site of the OBS in this study, some distortion of horizontally received signals would be likely.

Three-component record sections of the data from lines 6 and 4 are shown in Figures 5 and 6, respectively. Vertical seismograms appear narrow banded, the horizontal seismograms are even more band-limited, and the entire data set has a somewhat reverberatory character. These features are at least partly ascribable to the source spectrum, which is dominated by energy at a frequency equal to the reciprocal of the bubble pulse period appropriate to the charge size and detonation depth [e.g., Dobrin, 1976]. There is also a probable contribution from imperfect coupling of the OBS to the seafloor, indicating that the instrument likely sits on poorly consolidated sediments. Any reverberatory quality, however, does not greatly inhibit the interpretation of the several different phases (see Figure 1 for nomenclature) which can be seen as clear arrivals consistent across the record sections.

DATA PROCESSING

Horizontal component orientation was determined by

calculating the mean of the differences between the shot-receiver azimuth and the observed direction of horizontal particle motion for 16 water-wave arrivals on lines 4 and 6 [Tréhu, 1984]. Assuming that the water wave follows a direct path and that the shot and receiver locations are known, the horizontal components have orientations of 351° (H1) and 81° (H2) measured clockwise from north (Figure 7). The seismograms and particle motion plots of water-wave arrivals in Figure 7 show that the OBS plant was fortuitous: component H1 was oriented within 2° of being colinear with all but the closest shots on line 6. We refer to the component parallel to the particular line in question as radial and the component perpendicular to that line as transverse.

Shot-receiver ranges were determined from water-wave arrival times after shot instant corrections had been applied. First refracted arrivals were clear from all large shots on both lines, and picking errors were less than 0.02 s. Picking errors for first arrival travel times from small charges were generally less than 0.02 s with the exception of three charges near the rise crest which were assigned a picking error of 0.05 s. Topographic defocussing may be responsible for the quality of these three arrivals. Also, the presence of a low velocity gradient near the base of the crust would cause refracted arrivals in this range to be relatively less energetic. Figure 8 depicts the character of first arrival (P) information on a representative group of seismograms.

From the bathymetry along the shot line and the observed

horizontal slowness obtained from first arrival times, we next applied water-path corrections [Purdy, 1982a] to the data. This correction effectively places the shot at the point where the ray path from that shot to the OBS intersects the seabed and requires no assumptions to be made about the velocity structure of the uppermost crust. The total uncertainty associated with the assumptions, measurements, and corrections applied to travel time data of this type is discussed in Chapter 3. We estimate the overall uncertainty in the travel-time data on lines 4 and 6 to be between 0.05 and 0.11 s.

All record sections presented here have been reduced by a velocity of 8 km/s and have been left unfiltered beyond the effects of the anti-alias filter in the MIT OBS [Mattaboni and Solomon, 1977]. For ease of display, the amplitudes of all seismograms in the record sections in Figures 5 and 6 are scaled linearly with range and normalized to a range of 5 km. The amplitudes of small shots are increased by a factor of 4 relative to large shots for all three components, and the vertical component is shown amplified by a factor of 4 relative to the horizontal components for records generated by shots of both charge weights.

COMPRESSIONAL VELOCITY STRUCTURE

The Rise-Crossing Line

As a first step toward understanding the structure of the East Pacific Rise along line 6, we examine the travel times of

first compressional arrivals (denoted by P in Figure 5a). Figure 9a shows the 19 first-arrival travel times from shots west of the OBS, all picked from vertical-component seismograms and corrected for water path. A piecewise parabolic spline, constrained to have a negative second derivative, was fit through the travel times with a root-mean-square (rms) deviation of 0.02 s. This deviation is well below the minimum uncertainty in any travel time on the line. As is typical of most refraction experiments using surface sources, energy which refracts in the shallow crust is obscured by the direct water wave. We follow the work of Ewing and Purdy [1982] and approximate the shallow crustal structure with a linear velocity gradient obtained from three observables: the range and travel time of the data point at least range and, from the spline fit, the slowness at that range. After correcting the travel time curve for the effect of this linear-gradient layer, we determine the slowness and intercept at each shot point and invert these data directly for a velocity function which increases with depth [Purdy, 1982b]. The resulting compressional velocity (V_p) structure for the upper crust is shown in Figure 10a.

A velocity gradient of 4.5 s^{-1} was calculated for the uppermost 0.4 km of crust using the method of Ewing and Purdy [1982]. This procedure gives a velocity of 3.1 km/s at the seafloor. Substantial gradients such as that computed for the young shallow crust along line 6 can be attributed to a decrease in the volume of pores and cracks with depth in the basaltic extrusive layer [e.g., Chapter 3]. It is likely

that a decrease in shallow crustal velocity gradients with age occurs as a result of sediment infilling and chemical sealing of cracks and pores, and that this metamorphosis is a major contribution to the structural evolution of oceanic crust [Spudich and Orcutt, 1980a; Chapter 3].

Between 0.4 and 1.4 km depth, in the lower 1.0 km of what is commonly referred to as layer 2, the velocity gradient is reduced to 1.3 s^{-1} . The upper 2.3 km of layer 3 (between 1.3 and 3.7 km depth) is characterized by gentle gradients of between 0.1 and 0.2 s^{-1} . The deepest crust sampled by compressional waves along line 6 is at 3.7 km depth, where the velocity is 7.3 km/s. The shape of the velocity-depth curve below 3 km depth is poorly constrained by travel times alone because of the limited quantity of data from shots west of the rise.

The P wave from a shot 7 km west of the rise crest passes more than 3 km beneath the rise and arrives with a clear first motion (Figure 5a) and with no apparent time delay (Figure 9a). The travel-time delays and amplitude attenuation of entire seismograms seen by others [e.g., Rosendahl et al., 1976; Orcutt et al., 1976; Lewis and Garmany, 1982] on vertical-component records of P waves with rise crossing paths are not evident on any of our vertical seismograms. Our data show no evidence for a zone of low compressional velocity or enhanced P-wave attenuation within at least the upper 3 km beneath the East Pacific Rise crest at $11^{\circ}20'N$.

The modelling of amplitudes of different phases as they appear on a seismic refraction section can often provide complementary information to that derived from travel-time interpretation alone. Several factors, however, limit this approach. The data for line 6 are sparse and range-limited. Further, the focussing and defocussing effects of topography at scales greater than seismic wavelengths can profoundly affect seismic waveforms, especially in areas such as the EPR where the bathymetric signature is rough relative to older, well sedimented areas [see Figure 2 of Purdy, 1982a]. With the above cautions in mind, we nonetheless consider amplitude information as an additional constraint on P velocity structure. In particular, we seek models that match amplitude characteristics consistent on several seismograms within the vertical-component seismic section.

As a starting model, we use the upper crustal structure determined from the first arrival travel-time inversion, and we assume a lower crustal and upper mantle structure similar to that determined from waveform modelling of OBH data on line 4 [Chapter 3]. Waveforms are synthesized using the WKBJ method [Chapman, 1978]. The starting model is perturbed to fit the key characteristics of the observed seismic section for line 6. The final synthetic section is compared to the vertical-component observations collected on the rise-crossing line in Figure 11.

The steep velocity gradient in the uppermost crust determined from the travel-time data satisfactorily accounts for the focussing of energy at ranges of 13 km and less. The

amplitudes in this range decrease with distance from the OBS. This is in contrast to structures with a distinct transition between layers 2 and 3 which display amplitudes that typically build to a maximum at 5 to 10 km range [Spudich and Orcutt, 1980b; Chapter 3]. The amplitude data for line 6 cannot resolve a transition zone at the base of layer 2 beneath the EPR. Because the velocity signature of this transition has been shown to be variable along strike in this region [Chapter 3], however, the absence of a strong signature from such a transition zone on line 6 and the apparent presence of such a signature on a rise-crossing line (line 1, see Figure 3) to the north [Lewis and Garmany, 1982] is not surprising.

The existence of supercritical Moho reflections (PmP) from 20 km range to the end of our section suggests that a Moho is present beneath at least the eastern flank of the EPR. To model the strength of PmP relative to lower crustal refractions (P_3), we hold the velocity constant at 7.3 km/s in the lower crust. To model the time separation between P_3 and PmP, we set the thickness of the Moho transition to 0.5 km, and we place the top of the Moho at a depth of 6 km. In a limited data set such as that from line 6, the Moho thickness and depth can be traded off to match the interaction of P_3 and PmP phases. No mantle refractions (P_n) were recorded which might provide information on the velocity and depth of the uppermost mantle and thus assist in the modelling of the Moho. Nevertheless, the structure for the lower crust and Moho derived from waveform modelling (Figure 10a, at depths below the asterisk) of line 6

seismograms are in general agreement with those derived from larger data sets along line 1 [Lewis and Garmany, 1982] and line 4 [Lewis and Garmany, 1982; Chapter 3].

The Rise-Flanking Line

Although the limited quantity of data along line 4, especially in the range where P_mP , P_3 and P_n are expected to arrive, precludes the possibility of using amplitudes to model the structure of the crust and upper mantle, we can analyze the first-arrival travel times from line 4 to constrain further the shallow crustal structure in the vicinity of the EPR. The number of shots useful for a first-arrival time analysis on this line is, however, limited by the end of shooting 10 km to the south of the OBS and by the complication of the large seamount 14 km to the north. The 14 water-path corrected travel times from line 4 shown in Figure 9c are a merged set of the arrival times from north and south of the OBS. The data out to 11 km range can be fit with a piecewise parabolic spline with an rms deviation of 0.01s. The effects of the seamount on the travel times of the arrivals from the two most distant shots is evident in Figure 9c. Wave paths from these large charges are refracted beneath the seamount where the velocity structure most likely is laterally heterogeneous. Because of this effect, we invert only the data at ranges less than 11 km to determine the shallow crustal structure near the instrument.

The resulting compressional velocity structure is shown in Figure 10. The solution is very similar to that determined for the crust along line 6. A velocity of 2.9 km/s was calculated

for the top of the crust along line 4. In the uppermost 0.8 km of crust the velocity gradient is 4.0 s^{-1} . The velocity gradient in the final 0.6 km depth interval sampled by this small data set is 1.0 s^{-1} .

Figure 12 compares the travel-time data from lines 4 and 6 of this study with the merged set of first-arrival travel-time data examined in Chapter 3. The merged data set includes travel times from 10 independent refraction line segments along line 4 and collected by 5 OBHs located to the north of the MIT OBS (Figure 3). The data from that study were processed using techniques identical to those used in this chapter. The travel-time curve fit through the merged OBH data lags the travel time curves from lines 4 and 6 in this study by 0.10 to 0.15 s. Comparisons of the travel-time curves shown in Figure 12 and of the velocity structures derived from the travel time data sets (Figure 6 of Chapter 3 and Figure 10 of this chapter) suggest that the velocity of the shallow crust along line 6 and near the OBS along line 4 is higher than the average of this quantity determined for line 4 in Chapter 3. Possible explanations for this difference in shallow structure will be discussed below.

EFFECTS OF A SEDIMENT LAYER

On the horizontal-component records from both lines 4 and 6, energy associated with the first refracted P arrivals is curiously absent. Particle motion analysis indicates near-vertical motion for all first arrivals except those from shots located closest to the OBS. Compressional energy arriving

at a phase velocity of 6.8 km/s, typical of arrivals from 14 to 22 km range, should arrive at an incident angle of 27° for a receiver sitting on low velocity basement (3.1 km/s). The apparent angle of incidence, due to energy reflected and transmitted at the seabed-water boundary, would be 10° less than this value for a basement with a Poisson's ratio as high as 0.40 [Tréhu, 1984]. Even at ranges less than those appropriate to a 6.8 km/s phase velocity, however, the first P wave is incident much closer to the vertical than would be expected for an instrument situated on even highly porous basement.

Typical marine sediments with a P-wave velocity near 1.5 km/s [Hamilton, 1972] would not only decrease the actual angle of incidence at the OBS, but because of their high Poisson's ratio [$\nu > 0.48$; Hamilton, 1976] would transmit most of the arriving compressional energy into the water and thus decrease the apparent angle of incidence to that found from the particle motion analysis for lines 4 and 6. The presence of mechanically weak sediments would also result in poor OBS-substrate coupling if the shear modulus is sufficiently small. It is therefore likely that a sediment layer is present beneath the MIT OBS and that its high Poisson's ratio and low shear modulus are responsible for the steep apparent incidence angles of compressional arrivals and at least partially responsible for the resonant character of the seismograms.

Additional information about the sediment layer can be gained by studying the earliest arrivals (denoted by Ps in Figures 5 and 8) visible on the horizontal components from shots

on lines 4 and 6. This energy arrives at phase velocities equal to that of first refracted arrivals, but is delayed by 0.33 to 0.36 s relative to the initial P arrival observed on the vertical component. Lewis and McClain [1977] recognized a similar phenomenon in refraction records from the Gulf of California and identified the second arrival as a shear wave converted from the refracted compressional wave at the basement-sediment interface. Compressional-to-shear mode conversion at a basement-sediment boundary is quite efficient at horizontal phase velocities greater than the basement compressional velocity (Figure 13). Lewis and McClain [1977] derived a mean sediment-layer shear velocity (V_S) of 0.15 km/s using a sediment thickness (150 m) and compressional velocity (1.5 km/s) obtained from reflection profiles and bottom shots and a Ps-P lag time and Ps phase velocity measured from their OBS records.

Unfortunately, there are no reflection data with which to constrain sediment thickness and velocity along lines 4 and 6. However, given the young age of the crust, an assumed sedimentation rate of 35 $\mu\text{m/yr}$ [Purdy, 1982b], and the possible topographic effects on that rate on a local scale, a sediment thickness less than 100 m is likely. Setting the Ps-P lag time equal to the difference between the travel times of the s and p waves through the sediment layer, we can estimate the sediment thickness beneath the OBS. We assume that $V_p = 1.5$ km/s and $V_S = 0.2$ km/s [Hamilton, 1972; 1976] in the sediments and that the phase velocities of Ps and P arrivals are equal. A Ps-P lag time of 0.33 to 0.36 s then gives a sediment thickness of about

80 m at the OBS site.

A possible contributor to the ringing quality of the seismograms could be shear energy that is multiply reflected within the sediment layer. The near-vertically incident (incidence angle $< 2^\circ$) Ps arrivals should be almost totally reflected at the sediment-water interface. Multiples with a phase velocity greater than the P-wave velocity in the basement are also efficiently reflected at the base of the sediments. Figure 13 shows the reflection coefficient for SV waves incident from above at a representative sediment-basement contact. The two-way travel time for vertically propagating shear waves in a sedimentary layer 80 m thick with $V_S = 0.2$ km/s is 0.8 s. Packets of energy at approximately 0.8 s intervals can be seen especially well on the radial component records from small charges along line 6 (Figure 5b) and may represent shear multiples trapped within the sediment layer. Alternatively, these reverberatory pulses on the seismograms from the small charges may be the effect of frequency-dependent OBS-sediment coupling. Distinguishing between these two alternatives is precluded by a lack of further information on the OBS site characteristics.

THE NATURE OF REFRACTED SHEAR ENERGY

In a marine refraction experiment, efficient generation of shear energy occurs only at the contact of the sediment or water layer with the underlying basement [White and Stephen, 1980]. The appearance and strength of converted shear refractions have

been noted by many investigators to vary within a given region and even between adjacent shots and seismometers [e.g., Ewing and Ewing, 1959; Lewis and McClain, 1977]. Because the characteristics of an OBS site are, of course, constant for a given deployment, the variability of arriving shear refractions in such a situation can be largely attributed to lateral variation in the crust beneath the the seafloor at entry points of waves from different shots. Young ocean crust in general is highly cracked and porous near the surface. The low shear velocity of the shallow pillow basalts and rubble greatly reduces the energy of the transmitted S wave at the water-basement interface [Spudich and Orcutt, 1980b]. Lewis and McClain [1977] concluded that conversion from p to S is more likely to occur at a ray entry point where a sediment blanket is present. Cementation and infilling of cracks and pore spaces resulting from a cover of sediments, by this view, could raise V_S and decrease shear attenuation enough to make p to S mode conversion more efficient. It is unclear just how much the substitution of low velocity sediments ($V_p < 2.0$ km/s) for water in the porous upper crust would raise the bulk compressional velocity of young seafloor, however. Alternatively, a sediment layer may provide a capping blanket to hydrothermal circulation and thereby accelerate the cementation of cracks and pores by hydrothermal mineralization [Stephen and Harding, 1983]. A third hypothesis is that the porosity of young seafloor in areas where efficient p-to-S conversion is observed is inherently low,

and the P and S wave velocities intrinsically high, perhaps as a result of the presence of large-scale lava flows.

The conversion coefficient for downgoing p-to-S waves at the base of the sediments is shown in Figure 14 for representative physical properties of sediments and basement. The figure shows that a higher shear velocity for the basement enhances conversion to shear. Shear conversion at a water-basement interface is similarly affected. Variability in the amplitudes of shear arrivals for a given average structure may be due to topographic focussing and defocussing, crack and pore space heterogeneity in the basement, or lateral variation in both the amount of sediment present and the degree to which the underlying basement has been cemented and infilled. Also, when the phase velocity approaches V_p for the basement, p to S conversion efficiency is low (Figure 14). Since typical phase velocities for refracted shear arrivals (3.0 to 4.0 km/s) are within the range of characteristic basement compressional velocities (2.6 to 5.5 km/s [Spudich and Orcutt, 1980a]), weak shear arrivals may also be attributed to waves propagating at phase velocities near V_p for the local basement.

In light of the above discussion, we examine refracted shear energy on lines 4 and 6, and we comment on the implications of such arrivals for the sediment and basement structure along these lines.

The Rise-Crossing Line

The prominent later arrivals with phase velocity

substantially less than that for P on the vertical, radial, and transverse sections from the rise-crossing line (Figures 5 and 8) are shear waves converted from p at the top of the basement near the source and then refracted as S through the crust. All seismograms on the radial section (Figure 5b), including those from shots up to 12 km west of the rise crest, contain strong shear arrivals. This efficient conversion of p to S energy suggests that the shear-wave velocity at the top of the volcanic basement may be high relative to what is thought typical for young crust.

The refracted shear energy arrives at the OBS primarily as a shear wave (Ss phase in Figure 1). In areas with a high shear velocity at the top of the basement, it is also possible to record the arrival of energy refracted as S but converted to p at the basement-sediment boundary beneath the OBS. (The transmission coefficient curve for upgoing S to p conversion at phase velocities greater than 3.1 km/s is equal to the coefficient shown in Figure 14.) The seismograms in Figure 8 indicate that both Sp and Ss phases are present. Because the Sp phase is a doubly-converted arrival (p-S-p), however, it arrives with only a small fraction of the energy of the Ss arrivals. The weak Sp phase shows up best on the vertical channel (Figures 5 and 8), and the Ss phase appears strongest on the horizontal channels about 0.25 to 0.45 s later. These Ss-Sp time differences, together with the sediment velocities assumed earlier and the phase velocities of the shear refractions, give sediment thicknesses of 60 to 100 m beneath the OBS, a range compatible

with the estimate obtained from the Ps-P lag times.

Because the doubly converted phase Sp is not very energetic along line 6, an OBH or vertical-component OBS situated near the OBS site might fail to detect shear refractions propagating through the young oceanic lithosphere. In older, more heavily sedimented areas, however, such as the 15 m.y. old site in the Pacific examined by Spudich and Orcutt [1980b] and the 140 m.y. old site in the Atlantic studied by Purdy [1983], efficient p to S to p conversion produced strong arrivals on near-surface and ocean-bottom hydrophones, respectively, at phase velocities appropriate to refracted S waves. In these regions, chemical alteration and sediment infilling of the basalts with age presumably enhances p to S and S to p conversion by increasing the upper crustal velocities and thus forming a sharper velocity contrast at the sediment-basement contact where conversion takes place.

Particle motion diagrams for refracted shear arrivals recorded on the horizontal components are shown in Figure 15. The particle motion is dominated by the Ss arrival. The Sp phase is virtually undetectable on these diagrams because the background energy is large compared to the Sp arrival on the horizontal components (see the amplitude scales on the seismograms in Figure 8). Resonance of the horizontal geophone-sediment system may be partially responsible for the relative amplification on the horizontal components [Garmany, 1984]. The examples shown in Figure 15 are typical of all Ss phases recorded on line 6. Motion is not in the radial direction as

determined from the direct water wave, but consistently arrives at azimuths 10 to 35° southward of the expected westerly direction. The rectilinearity seen in many of the plots suggests that much of the energy is in phase on the two horizontal components. No obvious correlation is evident between the degree of deflection from a radial azimuth and whether the wave path crosses beneath the rise.

The deviations of these particle motions from the radial direction are probably too large and too consistent as a function of range to be ascribed to the deflection of ray paths from the shot to the OBS by a velocity heterogeneity near the rise axis, such as a low-velocity magma body. A three-dimensional analysis of the OBS-seafloor system by Garmany [1984] suggests that the oblique particle motion observed for arriving shear energy on line 6 could be the result of preferential excitation of one of several possible system resonances which can fall within the frequency band of a seismic refraction experiment. An alternative explanation for the observed deflections is that the shear arrivals have both radial (SV) and transverse (SH) polarizations. The arrival of both SV and SH phases from explosive sources would indicate some form of lateral heterogeneity along line 6. Because the particle motion of the Ss arrivals is fairly constant with range and because both SV and SH arrive at about the same time, the contributing heterogeneity would, most probably, be located near the OBS. A non-horizontal interface between sediments and basement is one possibility.

The Rise-Flanking Line

In sharp contrast to the line 6 data, shear arrivals (Ss or Sp) on all three components of the line 4 record sections (Figure 6) are relatively weak and are not coherent from shot to shot. Furthermore, no shear refractions were detected on any of the 10 refraction profiles recorded by OBHs to the north of the MIT OBS on line 4 [Chapter 3]. While particle motion and amplitude spectral studies therefore provide little information about the propagation of shear energy or the shear velocity structure along line 4, the absence or relative weakness of shear energy along this line nonetheless permits some further constraints to be placed on the shallow velocity structure along this flank of the EPR.

One explanation suggested to us for the difference in shear wave excitation between the two orthogonal lines is anisotropy of the oceanic crust [J.D. Garmany, personal communication, 1983]. Geologic observations from deep tow cameras and submersibles suggest that young seafloor is characterized by fissures striking parallel to the rise axis [e.g., Macdonald, 1982]; the consistent orientation of these fissures should result in some degree of velocity anisotropy. Garmany's suggestion is that SV waves travelling in the upper crust across the rise would have larger vertical components for a given phase velocity than SV waves travelling parallel to the rise and, as a consequence, the efficiency of p-to-S conversion would be greater for the rise-crossing line. A difficulty with the suggestion of velocity anisotropy, however, is that the inferred

compressional wave velocities in the upper crust on lines 4 and 6 and near the OBS are quite similar. Further, although some refracted shear energy appears in the vertical component data recorded by our OBS on line 6 (Figure 5a), none is apparent in the data recorded on the parallel line 1 (Figure 3) presented by Lewis and Garmany [1982]. Thus a simple and consistent geometric pattern to the observed shallow crustal velocities and mode conversions is not manifest throughout the ROSE area.

Alternatively, the lack of strong shear arrivals on either the OBS or OBH data from line 4 relative to line 6 could be due to differences in upper crustal structure between the two regions. Based on the previous discussion, if the pore and crack volume within the basement at the crustal entry points of waves propagating to the instruments were high along much of rise-flanking line relative to that of the crust along the rise-crossing line, the shear velocity contrast (and probably the compressional velocity contrast) at the top of the basement would not be as sharp and conversion of compressional to shear energy would be less efficient on line 4 than along line 6. The mean compressional velocity structure along line 4 determined in Chapter 3 supports this suggestion. The mean seafloor velocity inferred from the line 4 OBH data (2.1 km/s) is lower than that inferred from the line 6 OBS arrivals (3.1 km/s). The comparison of travel-time curves from both lines in Figure 12 further illustrates the difference in the upper crustal velocity structures. It should be noted that the seafloor velocity derived from the merged OBH data along line 4 is lower than

that calculated from the limited OBS data for the same line. The most likely explanation for this apparent inconsistency is that waves from the few shots included in the line-4 OBS profile propagate mainly through higher velocity upper crust near and beneath the instrument. If shots from ranges greater than 11 km to the north of the OBS had been available for study, by this suggestion, the shear arrivals from those shots should be even less energetic than for near ranges and the arrival times for the P waves from those shots should approach the mean travel-time curve from the OBH study.

If the difference in the efficiency of S wave conversion between lines 6 and 4 is ascribed to a greater infilling by sediments of the cracks and pores within the seafloor basement along the former line, a puzzling question is why more sediments have accumulated near the OBS and along line 6 than along the generally older line 4. The seamount to the north and EPR to the west may help to focus sediment deposition in the region near the OBS. Another possibility is that the local sediment accumulation rate is influenced by a bathymetric low in the basement along line 6. Records from the standard shipboard echo-sounder resolve a small topographic depression along line 4 at its intersection with line 6 and also along both lines near the instrument. However, the resolution of the echo-sounder is not sufficient to draw a strong conclusion about the nature of any sediment layer.

SHEAR VELOCITY STRUCTURE AND ATTENUATION BENEATH THE RISE

Using the travel times of the refracted shear waves (Ss), we have determined the shear-velocity structure along line 6. To determine arrival times, we have used--in addition to the individual components of data (V, R and T)--the product of the two horizontal components (Figure 8). Multiplication of R and T improves signal-to-noise ratio [Key, 1967; Fowler, 1976] and, because a substantial amount of shear energy was recorded on both horizontal components, tends to accentuate Ss arrivals. The shear travel times were corrected by the same methods used to correct the compressional first arrivals. Figure 9b shows the shear-wave travel times corrected for water path and plotted against range. The spline fit through these data has an rms deviation of 0.07 s. The onset of arriving Ss phases is not as sharp as that for first compressional arrivals so that picking errors are larger (>0.04 s). Also, reverberations interfere with picking accuracy, though the use of multiplied components appears to circumvent much of this problem (Figure 8).

The shear velocity structure beneath the EPR (Figure 10) can be determined following the procedure used earlier, by correcting the travel-time data for the effects of a surficial linear gradient [Ewing and Purdy, 1982], converting time and range to slowness and intercept for each shot, and then inverting these data for velocity as a function of depth. An average gradient of 1.3 s^{-1} was calculated for the upper 1.2 km of crust. The gradient in the highly porous upper few hundreds of meters is probably much higher than this, and the shear velocity at the surface is likely to be lower [Spudich and

Orcutt, 1980a] than the 1.8 km/s computed using the first clear shear arrival and the method of Ewing and Purdy [1982]. However, a substantially lower value for V_S at the top of the basement would decrease p-to-S conversion efficiency (Figure 14). The lowest shear velocity directly measurable from an observed refraction is 3.4 km/s at a depth of 1.2 km. The shear velocity increases smoothly below this depth with a gradient of 0.1 to 0.2 s⁻¹ to a velocity of 3.8 km/s at 4.6 km. The layer-3 shear velocity found by Fowler [1976] beneath the Mid-Atlantic Ridge (3.8 km/s) is similar to that at a depth of 4.5 km in our study. The entire shear structure calculated from the line-6 data, however, is in general slow compared to the shear velocity structures for crust older than 20 m.y. compiled by Spudich and Orcutt [1980a].

No attempt was made to model the refracted shear amplitudes with synthetic seismograms because of the variability in the generation of shear energy at the sediment-basement or water-basement contact and the possible distortion of observed shear amplitudes because of poor coupling between the horizontal geophones and the oceanic sediments.

Given both the compressional and shear velocity structures for the upper 3.7 km of crust, we can calculate Poisson's ratio ν as a function of depth for the crust along line 6 (Figure 10). For depths from which clear refracted arrivals were recorded (1.2 to 3.7 km), ν is nearly constant at 0.32. This Poisson's ratio is high relative to values reported elsewhere for normal layer 3 [e.g., Helmburger and Morris, 1970; Spudich and Orcutt,

1980b] and for some tectonically active regions [e.g., Fowler, 1978; Tréhu and Solomon, 1983]. Compressional and shear velocity measurements from refraction [Fowler, 1976] and earthquake [Francis, 1969] investigations on the Mid-Atlantic Ridge yielded Poisson's ratios of 0.31, a value which is essentially equivalent to that computed for most of the crust beneath the EPR in this study. The V_p/V_s ratio necessary to produce the observed Poisson's ratio along line 6 is the result of crustal shear velocities that are less than in older, perhaps better sealed, oceanic crust.

Because the generation of shear energy is highly dependent on the velocity structure of the uppermost crust at the ray entry point, a lack of shear energy recorded in a region does not necessarily indicate a zone of high shear attenuation at depth. If, however, shear energy does propagate through a region, the presence of a large magma chamber or zone of partial melt is unlikely. Fowler [1976] and Nisbet and Fowler [1978], for instance, concluded that no large crustal magma chamber exists beneath two sections of the Mid-Atlantic Ridge on the basis of clear arrivals of crustal shear waves that had passed beneath the ridge axis. From an examination of the seismograms in Figures 5, 6, and 8, it is obvious that shear energy propagates across the East Pacific Rise at crustal depths.

The amplitudes and ray paths of refracted shear energy may be used to constrain the shear-wave attenuation and to limit the dimensions of any magma body beneath the crest of the East Pacific Rise. Figure 16 shows the amplitude spectra of Ss

phases from four large shots on the rise-crossing line 6. The spectra of Ss. from shots at ranges of 32.6 and 37.1 km are for waves that have propagated beneath the rise. The amplitude of the shear arrival from 32.6 km is at least as strong as those from shorter ranges. Though the 37.1 km shear arrival is an order of magnitude less energetic than the arrival from 32.6 km, both arrivals show up clearly on the horizontal seismic sections (Figures 5 and 6). As was the case for compressional refractions through layer 3, a near-zero shear velocity gradient in the lower crust could explain the comparatively weak shear arrival at 37.1 km. Also, three-dimensional refraction effects, poor p-to-S conversion at the ray entry point, or some small amount of partial melt could result in a less energetic shear arrival from the most distant shot. However, no travel time delay, which might be expected for a shear refraction through partial melt, was resolvable (Figure 9b).

Because of the highly variable nature of shear wave generation in marine refraction experiments, the calculation of Q from spectral amplitude ratios (either between lines or between arrivals along the same line) would not be meaningful. However, it is evident from both the line-6 shear spectra in Figure 16 and the seismic profiles (Figure 5) that shear energy propagates efficiently across the EPR at crustal depths and that no large zone of high attenuation for shear exists in the crust beneath the rise at this latitude.

Ray tracing from shots along line 6 to the OBS through the shear velocity structure determined earlier (Figure 10) helps

to define the volume of crust sampled by the shear waves that passed beneath the rise axis in this experiment (Figure 17). Shear waves recorded from 8 shots along line 6 passed beneath the EPR within 2 km of the rise axis. On the basis of Figure 17, an isolated magma body up to 1 km in vertical extent could have been present within the crust beneath the EPR axis and still have escaped detection by this experiment. A magma body at a depth of more than 4 km beneath the rise axis would also not have been resolved by the data from this experiment.

DISCUSSION

The data presented in this chapter may be compared with the findings of Lewis and Garmany [1982] on crustal structure near and beneath the rise axis in the ROSE area (see Figure 3 for the locations of shot lines and instruments). Lewis and Garmany presented vertical-component seismograms from two OBSs along a line crossing the EPR (line 1), and their results are in general agreement with data obtained on the MIT OBS along line 6. Neither data set contains significantly attenuated compressional amplitudes that might be expected if a large crustal magma chamber were situated under the EPR in those regions. One of the Lewis and Garmany rise-crossing lines (line 1 shot to OBS UW-7) does, however, exhibit a 0.1 s travel-time delay (after correcting for topography) for arrivals which sample depths greater than 3 km below the seafloor of the rise axis. No such delay was resolved from the few shots on line 6 that sample comparable depths. Lewis and Garmany also recorded seismograms

from fan lines (lines 6 and 7 to OBSs UW-3 and UW-6) oriented obliquely to the rise crest. In contrast to the data collected perpendicular to rise strike, on these fan lines there was an abrupt attenuation of amplitude for paths crossing the rise axis for a prominent arrival corresponding to a multiple reflection at the seafloor of the caustic caused by the Moho transition zone. A zone of low-velocity material, located within the crust beneath the rise axis and with a width of 0.5 to 4 km and a poorly developed Moho transition, was proposed by Lewis and Garmany [1982] to explain both sets of observations.

A conclusion that follows directly from our data set together with the results of Lewis and Garmany [1982] and of other experiments conducted elsewhere along the East Pacific Rise [e.g., Orcutt et al., 1976; Rosendahl et al., 1976; Herron et al., 1978] is that any axial magma chamber in the ROSE area is not a steady-state phenomenon. If a magma chamber of the size inferred by Rosendahl et al. [1976], Herron et al. [1978, 1980], and Hale et al. [1982] beneath the rise axis once existed beneath the line-6 region, it has since been cooled and either completely or partially solidified, presumably by vigorous hydrothermal circulation [Lister, 1983]. This conclusion is in contrast to commonly held views on the structure of ridge axes at intermediate to fast spreading rates [Sleep, 1975; Macdonald, 1982; Sanford and Einarsson, 1982].

Within the ROSE area of the East Pacific Rise (Figure 3), a reasonable question is whether a single two-dimensional structure for the rise axis region can account for the data

presented in this chapter as well as those reported by Lewis and Garmany [1982]. While along-strike variability cannot be excluded, we can envision a model for the rise axis region that might satisfy both sets of results.

In this model, magma exists at crustal depths beneath the rise crest, but only in narrow dikes or pipes or in small isolated pockets. Because a finite-viscosity fluid is capable of supporting some shear stress, shear propagation through thin regions of even total melt is possible. Shear wave energy was observed to pass through a small magma lens at Kilauea Iki Crater, and this result has been used to model the in situ bulk properties of molten rock [Fehler and Aki, 1976; Aki et al., 1978]. A coefficient of only 0.05, however, was determined for transmission of the shear wave through that magma body, which had a thickness of about half a shear wavelength. The frequency content of the data reported here dictate that shear energy passing through viscous melt ($V_S = 0.2$ km/s, $\eta \sim 10^7$ poise [Aki et al., 1978]) would have a wavelength of about 30 m. While the observation of energetic shear arrivals on rise-crossing seismograms does not preclude the existence of magma in the region of the crust sampled by the shear waves beneath the rise axis, it does suggest that any continuous curtain of melt transversed by such waves must be much thinner than a wavelength. Magma intrusions in the form of narrow dikes or pipes beneath the rise crest, however, would most likely have gone undetected by the shear energy that propagated along line 6. Alternatively, a small magma body could be present

beneath the axis but lie between regions sampled by the rise-crossing shear waves or missed by shear energy diffracting around the low velocity structure. The distribution of ray paths shown in Figure 17 suggests that a small magma chamber cannot be much greater than 1 km in vertical extent. Whatever the distribution of crustal magma bodies (if any), the axial structure would be more akin to the "infinite leak" model proposed by Nisbet and Fowler [1978] for the slower spreading Mid-Atlantic Ridge than to the more conventional magma chamber models [e.g., Cann, 1974] invoked to explain the EPR structure at 9°N [Rosendahl et al., 1976; Orcutt et al., 1976; Herron et al., 1978, 1980].

The significant attenuation observed on Lewis and Garmany's rise-crossing fan lines, by the above model, is ascribed at least partly to the oblique azimuths at which the waves pass beneath the rise. Along such paths the waves must pass through a greater volume of rise-axis material than do waves travelling in planes normal to the rise strike. Further, since the waves observed by Lewis and Garmany [1982] to be attenuated by the rise consist of more nearly vertically-travelling energy, a thin axial magma body or a small high-level magma chamber could attenuate such waves more dramatically than for the shear waves propagating nearly horizontally to 4 km depth beneath the rise crest and observed in this study.

Another feature of the refraction results for line 6 worthy of explanation is the higher Poisson's ratio ν than is commonly observed for older ocean crust. As noted earlier, comparably

high values for this ratio have been seen in a few other mid-ocean ridge regions [e.g., Fowler, 1976; Francis, 1969]. An increase in the temperature of typical earth materials (gabbros, basalts, dunites) will increase v by less than 2% per 100°K, at least for temperatures between 25 and 400°C [e.g., Hughes and Maurette, 1957], so temperature alone is unlikely to be sufficient to account for the difference in v between very young and old oceanic crust. Much of the enhancement in v may be due fluid-filled fractures in the young rocks of a mid-ocean ridge. The self-consistent analysis of O'Connell and Budianski [1974] indicates that an increase in the number of saturated, elliptical cracks per unit volume will generally be accompanied by an increase in v in the bulk solid. It is likely that as the crust in this region ages, chemical alteration and further sediment deposition will act to seal the fractures in the young crust and thus decrease Poisson's ratio to values (0.25 to 0.30) more typical of mature oceanic crust [e.g., Spudich and Orcutt, 1980b].

CONCLUSIONS

We have presented an analysis of two refraction lines, across and adjacent to the crest of the East Pacific Rise at latitude 11°20'N, as recorded by a three-component ocean-bottom seismometer. Two general conclusions can be drawn from a comparison of this study with related published work: (1) The use of three-component ocean-bottom instruments to record marine refraction data can greatly enhance the amount of information on the structure of oceanic lithosphere that can be gathered from a

seismic refraction profile, particularly on the shear velocity structure. (2) The quantity and distribution of molten material in the crust along the axis of the East Pacific Rise must be discontinuous in time or along strike.

The compressional velocity structure for the crust beneath and adjacent to the EPR at 11°20'N is, within the bounds of uncertainty, similar to structures found for 'normal' young crust in tectonically inactive regions [Spudich and Orcutt, 1980a]. The principal difference between the crustal model derived here and the model computed for young crust immediately north of the area of this study [Chapter 3] is that the P-wave velocity in the upper crust is somewhat higher in this area. The presence of strong refracted shear arrivals on the horizontal components indicates that the S-wave velocity at the seafloor may also be unusually high for young oceanic crust. We suggest that either filling of the cracked and porous volcanic basement with sediments or hydrothermal mineralization or the presence of basement volcanic deposits with atypically low porosity has served both to improve the efficiency of p-to-S conversion at the top of the basement and to raise the seismic velocities in the uppermost part of layer 2 relative to other areas of similarly young seafloor. The shear velocity structure derived for this region indicates a crustal Poisson's ratio higher than for normal oceanic crust but not significantly higher than has been observed in other ridge environments. From an examination of particle motions and amplitudes of converted and direct seismic phases and the reverberatory character of the seismograms, a thin layer of

sediment is inferred to be present beneath the OBS.

Both shear and compressional arrivals for rise-crossing paths are recorded with high amplitudes and without any apparent travel-time delays associated with the rise axis. A crustal magma body of the size inferred from seismic data to be present at 9°N [Orcutt et al., 1976; Herron et al., 1978, 1980] is therefore not present beneath the EPR at 11°20'N. On the basis of the data reported here and those presented by Lewis and Garmany [1982] for another portion of the ROSE area, we conclude that if magma is present within the crust beneath the rise axis in this region, it is confined to narrow dikes or pipes or to small isolated bodies not much larger than a kilometer in vertical extent.

REFERENCES

- Aki, K., and P.G. Richards, Quantitative Seismology, W.H. Freeman and Co., San Francisco, pp. 150-151, 1980.
- Aki, K., B. Chouet, M. Fehler, G. Zandt, R. Koyanagi, J. Colp, and R.G. Hay, Seismic properties of a shallow magma reservoir in Kilauea Iki by active and passive experiments, J. Geophys. Res., 83, 2273-2282, 1978.
- Bratt, S.R., and G.M. Purdy, Structure and variability of oceanic crust on the flanks of the East Pacific Rise between 11° and 13°N, J. Geophys. Res., in press, 1984.
- Bullen, K.E., An Introduction to the Theory of Seismology, Cambridge University Press, pp. 129-130, 1963.
- Cann, J.R., A model for oceanic crustal structure developed, Geophys. J. R. Astron. Soc., 39, 169-187, 1974.
- Chapman, C.H., A new method for computing synthetic seismograms, Geophys. J. R. Astron. Soc., 54, 481-518, 1978.
- Christensen, N.I., and M.H. Salisbury, Structure and constitution of the lower oceanic crust, Rev. Geophys. Space Phys., 13, 57-86, 1975.
- Dobrin, M.B., Introduction to Geophysical Prospecting, 3rd ed., McGraw-Hill Book Co., New York, 1976.
- Duschenes, J.D., T.W. Barash, P.J. Mattaboni, and S.C. Solomon, On the use of an externally deployed geophone package on an ocean bottom seismometer, Mar. Geophys. Res., 4, 437-450, 1981.
- Ewing, J., and M. Ewing, Seismic refraction measurements in the Atlantic Ocean basins, in the Mediterranean Sea, on the

- Mid-Atlantic Ridge, and in the Norwegian Sea, Geol. Soc. Amer. Bull., 70, 291-318, 1959.
- Ewing, J.I., and R.P. Meyer, Rivera Ocean Seismic Experiment (ROSE) overview, J. Geophys. Res., 87, 8345-8357, 1982.
- Ewing, J.I., and G.M. Purdy, Upper crustal velocity structure in the ROSE area of the East Pacific Rise, J. Geophys. Res., 87, 8397-8402, 1982.
- Fehler, M., and K. Aki, Numerical study of diffraction of plane elastic waves by a finite crack with application to location of a magma lens, Bull. Seismol. Soc. Amer., 68, 573-598, 1978.
- Forsyth, D.W., The evolution of the upper mantle beneath mid-ocean ridges, Tectonophysics, 38, 89-118, 1977.
- Fowler, C.M.R., Crustal structure of the Mid-Atlantic Ridge crest at 37°N, Geophys. J. R. Astron. Soc., 47, 459-492, 1976.
- Fowler, C.M.R., The Mid-Atlantic Ridge: Structure at 45°N, Geophys. J. R. Astron. Soc., 54, 167-183, 1978.
- Francis, T.J.G., Upper mantle structure along the axis of the Mid-Atlantic Ridge near Iceland, Geophys. J. R. Astron. Soc., 17, 507-520, 1969.
- Garmany, J.D., The recovery of true particle motion from three component ocean bottom seismometer data, J. Geophys. Res., submitted, 1984.
- Hale, L.D., C.J. Morton, and N.H. Sleep, Reinterpretation of seismic reflection data over the East Pacific Rise, J. Geophys. Res., 87, 7707-7717, 1982.
- Hamilton, E.L., Compressional wave attenuation in marine sediments, Geophysics, 37, 620-646, 1972.

- Hamilton, E.L., Shear-wave velocity versus depth in marine sediments: A review, Geophysics, 41, 985-996, 1976.
- Helmberger, D.V., and G.B. Morris, A travel time and amplitude interpretation of a marine refraction profile: Transformed shear waves, Bull. Seismol. Soc. Amer., 60, 593-600, 1970.
- Herron, T.J., W.J. Ludwig, P.L. Stoffa, T.K. Kan, and P. Buhl, Structure of the East Pacific Rise crest from multi-channel seismic reflection data, J. Geophys. Res., 83, 798-804, 1978.
- Herron, T.J., P.L. Stoffa and P. Buhl, Magma chamber and mantle reflections - East Pacific Rise, Geophys. Res. Lett., 7, 989-992, 1980.
- Houtz, R.E., and J. Ewing, Upper crustal structure as a function of plate age, J. Geophys. Res., 81, 2490-2498, 1976.
- Hughes, D.S., and C. Maurette, Variation of elastic wave velocities in basic igneous rocks with pressure and temperature, Geophysics, 22, 23-31, 1957.
- Key, F.A., Signal-generated noise recorded at the Eskdalemuir seismometer array station, Bull. Seismol. Soc. Am., 57, 27-37, 1967.
- Klitgord, K.D., and J. Mammerrickx, Northern East Pacific Rise: Magnetic anomaly and bathymetric framework, J. Geophys. Res., 87, 6725-6750, 1982.
- Lewis, B.T.R., and J.D. Garmany, Constraints on the structure of the East Pacific rise from seismic refraction data, J. Geophys. Res., 87, 8417-8425, 1982.
- Lewis, B.T.R., and J. McClain, Converted shear waves seen by ocean bottom seismometers and surface buoys, Bull. Seis. Soc. Amer., 67, 1291-1302, 1977.

- Lilwall, R.C., T.J.G. Francis, and I.T. Porter, Ocean-bottom seismograph observations on the mid-Atlantic ridge near 45°N - further results, Geophys. J. R. Astron. Soc., 55, 255-262, 1978.
- Lister, C.R.B., On the intermittency and crystallization mechanism of sub-seafloor magma chambers, Geophys. J. R. Astron. Soc., 73, 351-365, 1983.
- Macdonald, K.C., Mid-ocean ridges: Fine scale tectonic, volcanic and hydrothermal processes within the plate boundary, Ann. Rev. Earth Planet. Sci., 10, 155-190, 1982.
- Mattaboni, P.J., and S.C. Solomon, MITOBS: A seismometer system for ocean-bottom earthquake studies, Mar. Geophys. Res., 3, 87-102, 1977.
- McClain, J.S., and B.T.R. Lewis, A seismic experiment at the axis of the East Pacific Rise, Marine Geol., 35, 147-169, 1980.
- Nisbit, E.G., and C.M.R. Fowler, The Mid-Atlantic Ridge at 37° and 45°N: Some geophysical and petrological constraints, Geophys. J. R. Astron. Soc., 54, 631-660, 1978.
- O'Connell, R.J., and B. Budiansky, Seismic velocities in dry and saturated cracked solids, J. Geophys. Res., 79, 5412-5426, 1974.
- Orcutt, J.A., B.L.N. Kennett, and L.M. Dorman, Structure of the East Pacific Rise from an ocean bottom seismometer survey, Geophys. J. R. Astron. Soc., 45, 305-320, 1976.
- Purdy, G.M., The correction for the travel time effects of seafloor topography in the interpretation of marine seismic data, J. Geophys. Res., 87, 8389-8396, 1982a.

- Purdy, G.M., The variability in seismic structure of layer 2 near the East Pacific Rise at 12°N, J. Geophys. Res., 87, 8403-8416, 1982b.
- Purdy, G.M., The seismic structure of 140 m.y. old crust in the western central Atlantic Ocean, Geophys. J. R. Astron. Soc., 72, 115-138, 1983.
- Reid, I., J.A. Orcutt, and W.A. Prothero, Seismic evidence for a narrow zone of partial melting underlying the East Pacific Rise at 21°N, Geol. Soc. Amer. Bull., 88, 678-682, 1977.
- Rosencrantz, E., Formation of uppermost oceanic crust, Tectonics, 1, 471-494, 1982.
- Rosendahl, B.R., R.W. Raitt, L.M. Dorman, L.D. Bibee, D.M. Hussong, G.H. Sutton, Evolution of oceanic crust, 1, A physical model of the East Pacific Rise crest derived from seismic refraction data, J. Geophys. Res., 81, 5294-5304, 1976.
- Sanford, A.R., and P. Einarsson, Magma chambers in rifts, in Continental and Oceanic Rifts, Geodynamic Series, vol. 8, edited by G. Palmason, pp 147-168, AGU, Washington, D.C., 1982.
- Sleep, N.H., Formation of oceanic crust: Some thermal constraints, J. Geophys. Res., 80, 4037-4042, 1975.
- Solomon, S.C., Shear-wave attenuation and melting beneath the Mid-Atlantic Ridge, J. Geophys. Res., 78, 6044-6059, 1973.
- Solomon, S.C., and B.R. Julian, Seismic constraints on ocean-ridge mantle structure: Anomalous fault plane solutions from first motions, Geophys. J. R. Astron. Soc., 38, 265-285, 1974.

- Spudich, P., and J. Orcutt, A new look at the seismic velocity structure of oceanic crust, Rev. Geophys. Space Phys., 18, 627-645, 1980a.
- Spudich, P., and J. Orcutt, Petrology and porosity of an oceanic crustal site: Results from waveform modeling of seismic refraction data, J. Geophys. Res., 85, 1409-1433, 1980b.
- Steinmetz, L., R.B. Whitmarsh, and V.S. Moraira, Upper mantle structure beneath the mid-Atlantic Ridge north of the Azores based on observations of compressional waves, Geophys. J. R. Astron. Soc., 50, 353-380, 1977.
- Stephen, R.A., and A.J. Harding, Travel-time analysis of borehole seismic data, J. Geophys. Res., in press, 1983.
- Sutton, G.H., F.K. Duennebier, B. Iwatake, J.D. Tuthill, B.T.R. Lewis and J. Ewing, An overview and general results of the Lopez Island OBS experiment, Mar. Geophys. Res., 5, 3-34, 1981a.
- Sutton, G.H., F.K. Duennebier, and B. Iwatake, Coupling of ocean bottom seismometers to soft bottom, Mar. Geophys. Res., 5, 35-51, 1981b.
- Toomey, D.R., S.C. Solomon, G.M. Purdy, and M.H. Murray, Micro-earthquakes on the Mid-Atlantic Ridge near 23°N: New observations with a large network (abstract), EOS Trans. AGU, 63, 1103, 1982.
- Tréhu, A.M., On the use of observed incident angles and azimuths to infer lateral velocity variations in the Orozco transform fault, Geophys. J. R. Astron. Soc., in press, 1984.
- Tréhu, A.M., and S.C. Solomon, Coupling parameters of the MIT OBS at two nearshore sites, Mar. Geophys. Res., 5, 69-78, 1981.

- Tréhu, A.M., and S.C. Solomon, Earthquakes in the Orozco transform zone: Seismicity, source mechanisms, and tectonics, J. Geophys. Res., 88, 8203-8225, 1983.
- Walsh, J.B., Attenuation in partially melted material, J. Geophys. Res., 73, 2209-2216, 1968.
- Walsh, J.B., New analysis of attenuation in partially melted rock, J. Geophys. Res., 74, 4333-4337, 1969.
- Weertman, J., Theory of water-filled crevasses in glaciers applied to vertical magma transport beneath ocean ridges, J. Geophys. Res., 76, 1171-1183, 1971.
- White, R.S., and R.A. Stephen, Compressional to shear conversion in oceanic crust, Geophys. J. R. Astron. Soc., 63, 547-565, 1980.
- Zelikovitz, S.J., and W.A. Prothero, Jr., The vertical response of an ocean bottom seismometer: Analysis of the Lopez Island vertical transient tests, Mar. Geophys. Res., 5, 53-67, 1981.

FIGURE CAPTIONS

- Figure 1. Schematic cross-section of the crust and upper mantle, showing the wave paths of phases examined in this study and the associated nomenclature used in the text. Compressional and shear waves propagating in the sediment layer are denoted by p or s, respectively, while such waves in the igneous crust and mantle are denoted by P or S, respectively.
- Figure 2. Geographic location of Project ROSE, from Ewing and Meyer [1982]. Bathymetric contours, in m, are from Klitgord and Mammerickx [1982]. The numbered lines at the bottom of the figure are the seismic refraction lines from the first phase of that project. The box in the upper left encloses the microearthquake study region [e.g., Tréhu and Solomon, 1983].
- Figure 3. Locations of refraction lines and instruments discussed in this chapter in relation to the local bathymetry [after Klitgord and Mammerickx, 1982]. The locations of instruments from the Woods Hole Oceanographic Institute (WHOI) and the University of Washington (UW) are shown in addition to that for the MIT instrument. The interval between the solid contours is 500m.
- Figure 4. Bathymetric profiles along the two refraction lines discussed in this study as a function of range from

the OBS. Shot locations are indicated by dots along the profile; selected shot numbers are shown. (a) Line 6. (b) Line 4.

Figure 5. Three-component record sections from the MIT OBS for line 6. All travel times have been reduced to $t-x/8$, where t is travel time in s and x is range in km. The amplitudes have been adjusted for range and charge weight as explained in the text; no bathymetric corrections have been made. Each seismogram stops at the arrival time of the water wave. The radial component is the horizontal component most nearly aligned with the strike of the refraction line. The travel-time curves for P and Ss phases are also shown. (a) Vertical component section. (b) Radial component section. (c) Transverse component section.

Figure 6. Three-component record sections from the MIT OBS for line 4. For other information, see Figure 5.

Figure 7. Orientation of the horizontal components of the MIT OBS, determined from water-wave particle motion. Shown are selected water-wave seismograms and horizontal-component particle motion from shots (dots) located at various azimuths from the instrument (star). The axis of the East Pacific Rise is indicated by a dashed line.

Figure 8. Representative seismograms from four shots on lines 6 and 4. The product of the two horizontal components ($H1 \times H2$) is shown at the top followed by the horizontal ($H2$ and $H1$) and vertical (V) component

seismograms. For line 6, H2 is approximately radial and H1 transverse; for line 4, the reverse holds. Phases discussed in the text and illustrated in Figure 1 are labeled. The amplitude scale of the records varies and is shown, in arbitrary digital units, on each seismogram. The time scale is represented as vertical lines across each seismogram spaced 1 s apart. The range from each shot is indicated. Note the Ps-P time lag between the horizontal and vertical components and the clear shear arrivals on the horizontal and H1xH2 seismograms. (a) Shot number 2223 (11 kg) on line 4. (b) Shot number 3216 (11 kg) on line 6. (c) Shot number 3206 (11 kg) on line 6. (d) Shot number 4431 (217 kg) on line 6.

Figure 9. Travel times versus range for refracted compressional (P) and shear (Ss) waves. A water path correction [Purdy, 1982a] has been applied to the data as explained in the text. (a) Line 6, compressional arrivals. (b) Line 6, shear arrivals. (c) Line 4, compressional arrivals.

Figure 10. Compressional and shear velocity structure along lines 6 and 4 determined principally from the inversion of travel times. The structure below the asterisk for line 6 was determined by amplitude modelling. Also shown is the inferred Poisson's ratio versus depth.

Figure 11. (Top) WKBJ synthetic seismic section calculated for the crustal structure (shown in Figure 10) for line 6. The source time function is an exponentially damped cosine

function, $\cos(2\pi ft)\exp[-(2\pi ft/\gamma)^2]$, where $f = 8$ Hz and $\gamma = 5$. The scaling of seismograms with range is the same as in Figures 5 and 6. (Bottom) The vertical-component data from line 6 to the west of the OBS, shown for comparison.

Figure 12. Comparison of the P-wave travel-time data collected by 5 OBHs on line 4 [Chapter 3] to the north of the MIT OBS with the travel time curves for lines 6 and 4 determined in this chapter. Both sets of data were identically processed. The difference between the sets of travel times from the two studies in the ROSE area is probably due to lower velocity in the upper crust beneath the OBHs.

Figure 13. Transmission coefficient for upgoing P-to-s plane waves and reflection coefficient for downgoing s-to-s (SV polarization) as functions of horizontal phase velocity [Aki and Richards, 1980]. The boundary is that between igneous basement ($V_p = 3.1$ km/s, $V_s = 1.8$ km/s) and overlying sediment ($V_p = 1.5$ km/s, $V_s = 0.2$ km/s). Coefficients shown are relative to a unit displacement. This figure illustrates the efficiency of converting refracted compressional (P) energy to shear (s) energy at a typical sediment-basement interface and of internal reflection of that shear energy within the sediment layer.

Figure 14. Transmission coefficient for downgoing p-to-S plane waves at a sediment-basement interface as a function of

horizontal phase velocity [Aki and Richards, 1980]. Curves are shown for three values of the basement shear velocity (v); other velocity parameters are as in Figure 13. Coefficients shown are relative to a unit displacement. Infilling of the highly porous extrusive basalts by sediments can raise the basement shear velocity and thus foster more efficient generation of refracted shear energy from explosive sources in the ocean.

Figure 15. Particle motion of refracted shear arrivals (Ss) from selected shots on line 6. Motion is shown in the horizontal plane. Polarization is not in the radial direction, but is aligned along an azimuth southward of the expected radial direction of approach.

Figure 16. Amplitude spectra of refracted shear arrivals (Ss) from four large charges on line 6. The spectra from shots at 32.6 and 37.1 km range measure shear energy that has propagated through the lower crust beneath the East Pacific Rise crest. For clarity, each spectrum has been raised by 1 unit of $\log_{10}$ amp with respect to the spectrum immediately beneath; note the sliding labels in the upper part of the ordinate scale.

Figure 17. Paths for Ss waves from shots on line 6 to the OBS. Paths are calculated by ray tracing through the shear velocity structure determined for the crust along line 6 (Figure 10).

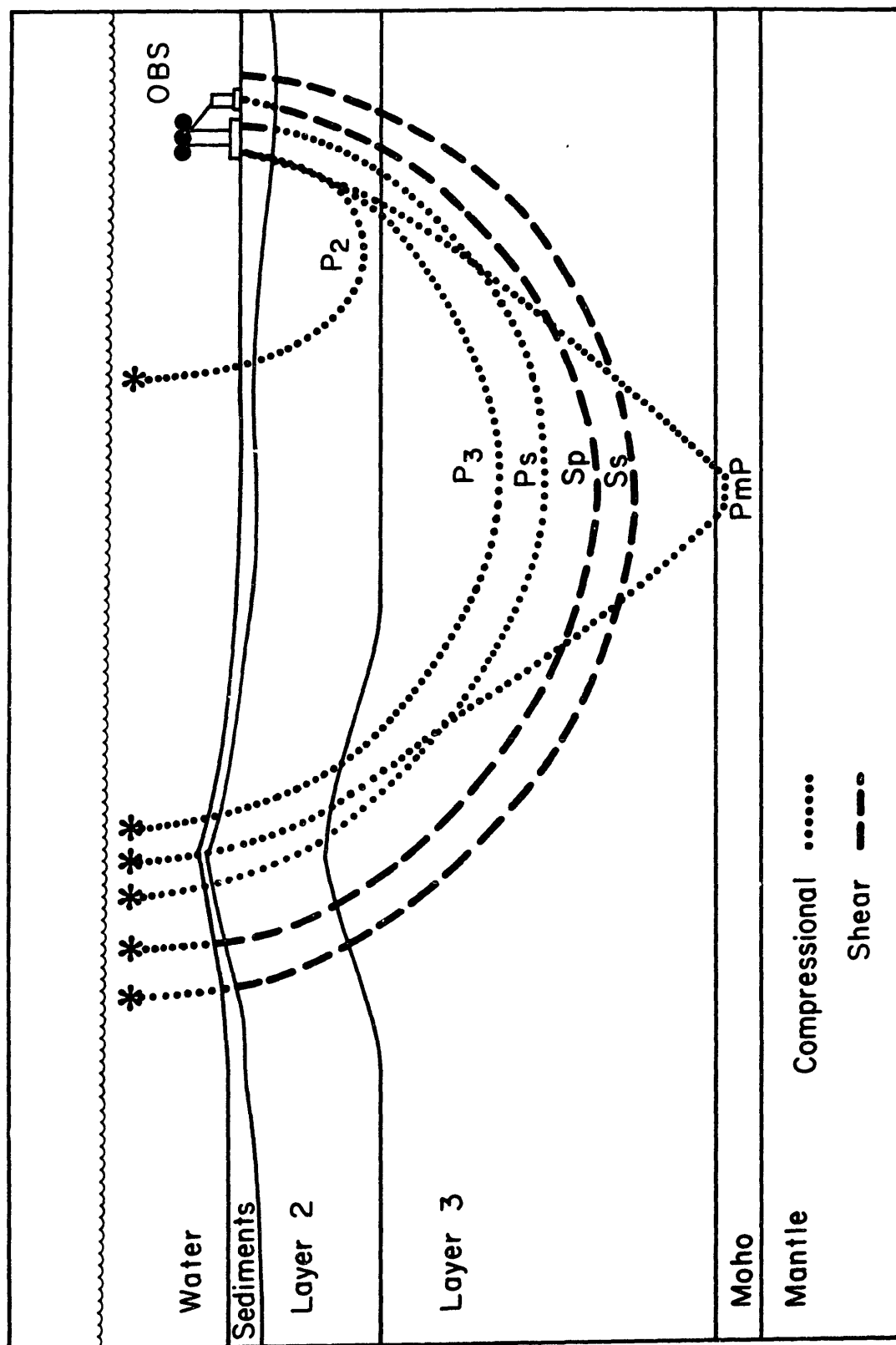


Figure 1

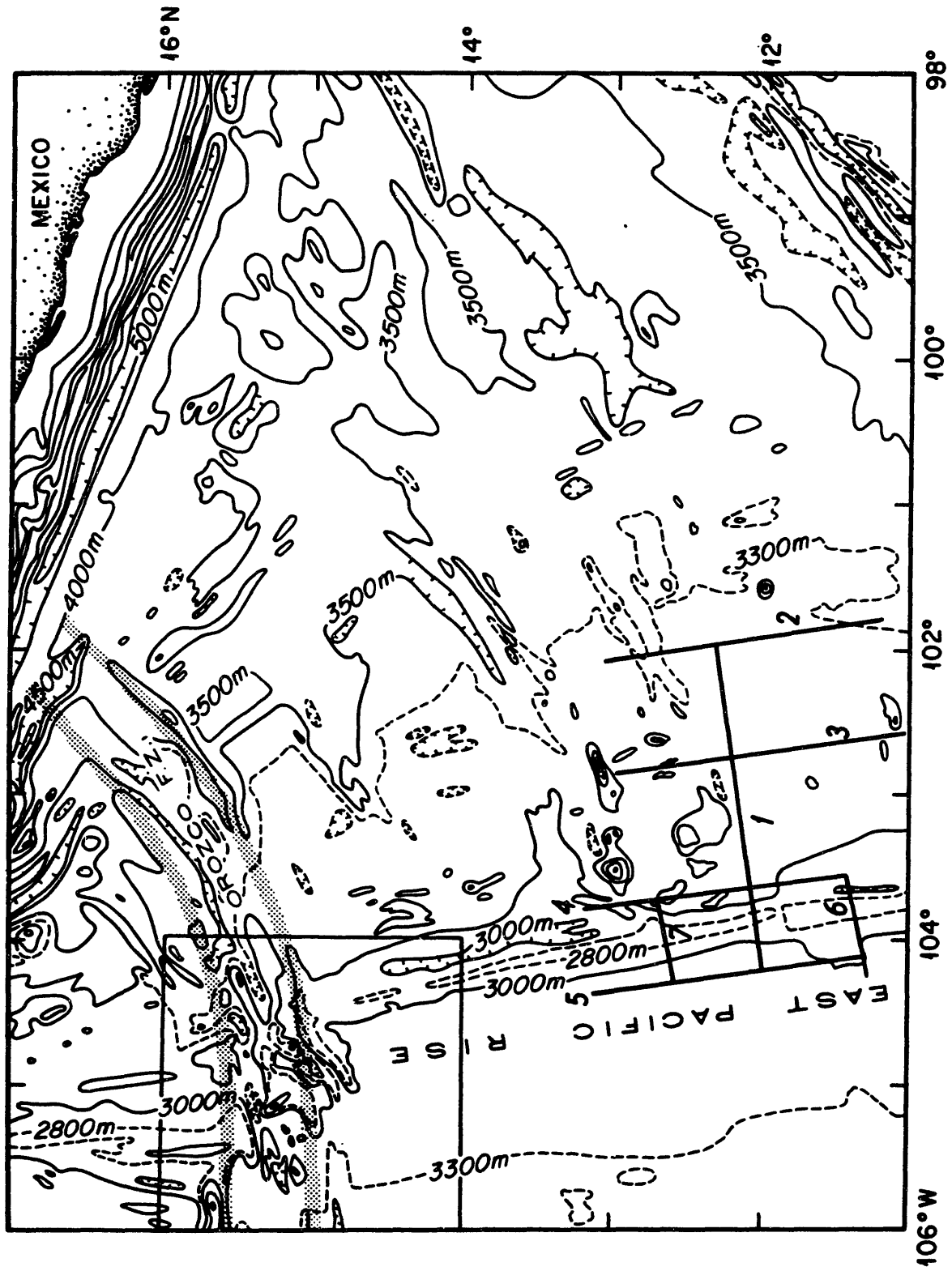


Figure 2

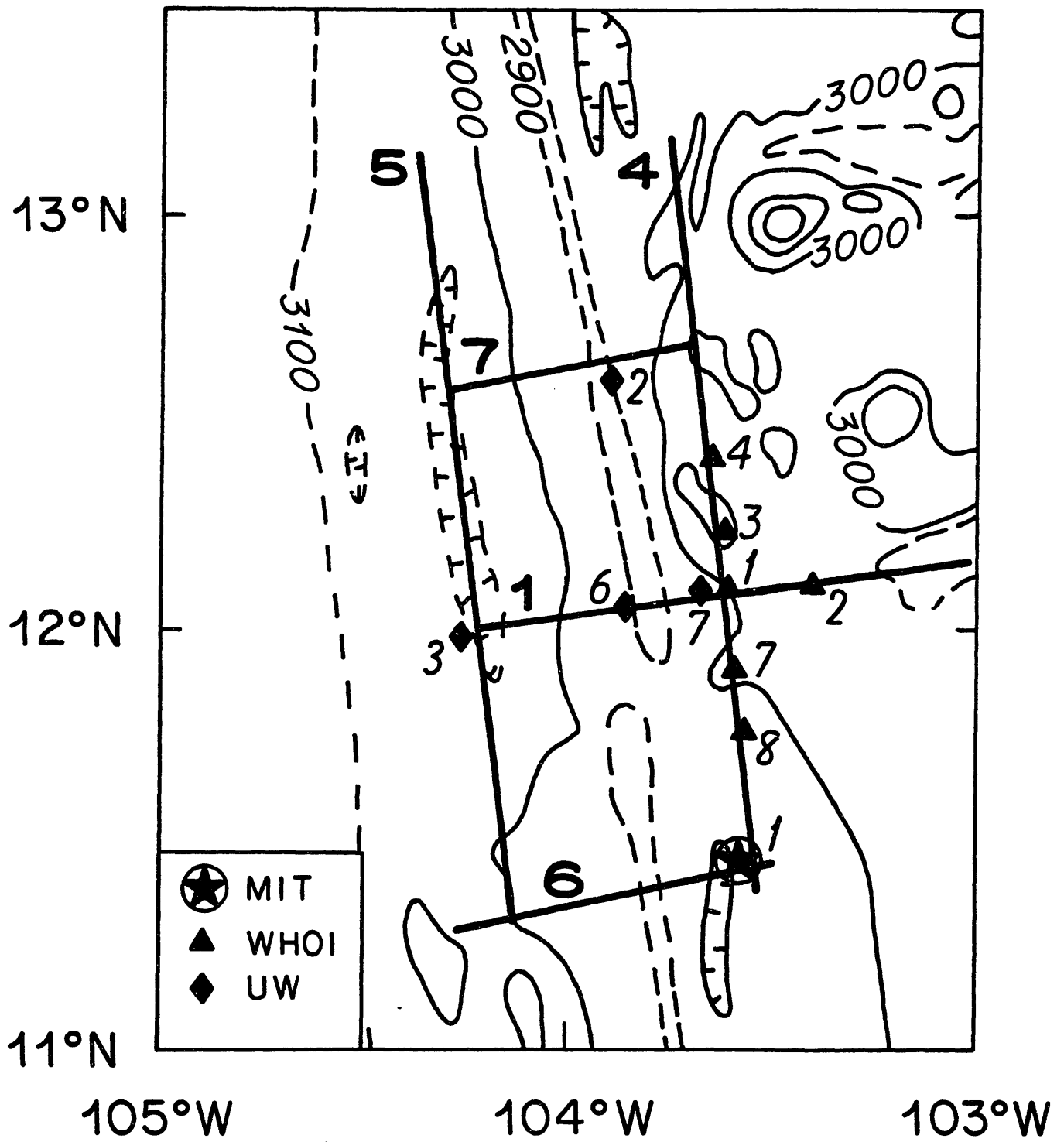


Figure 3

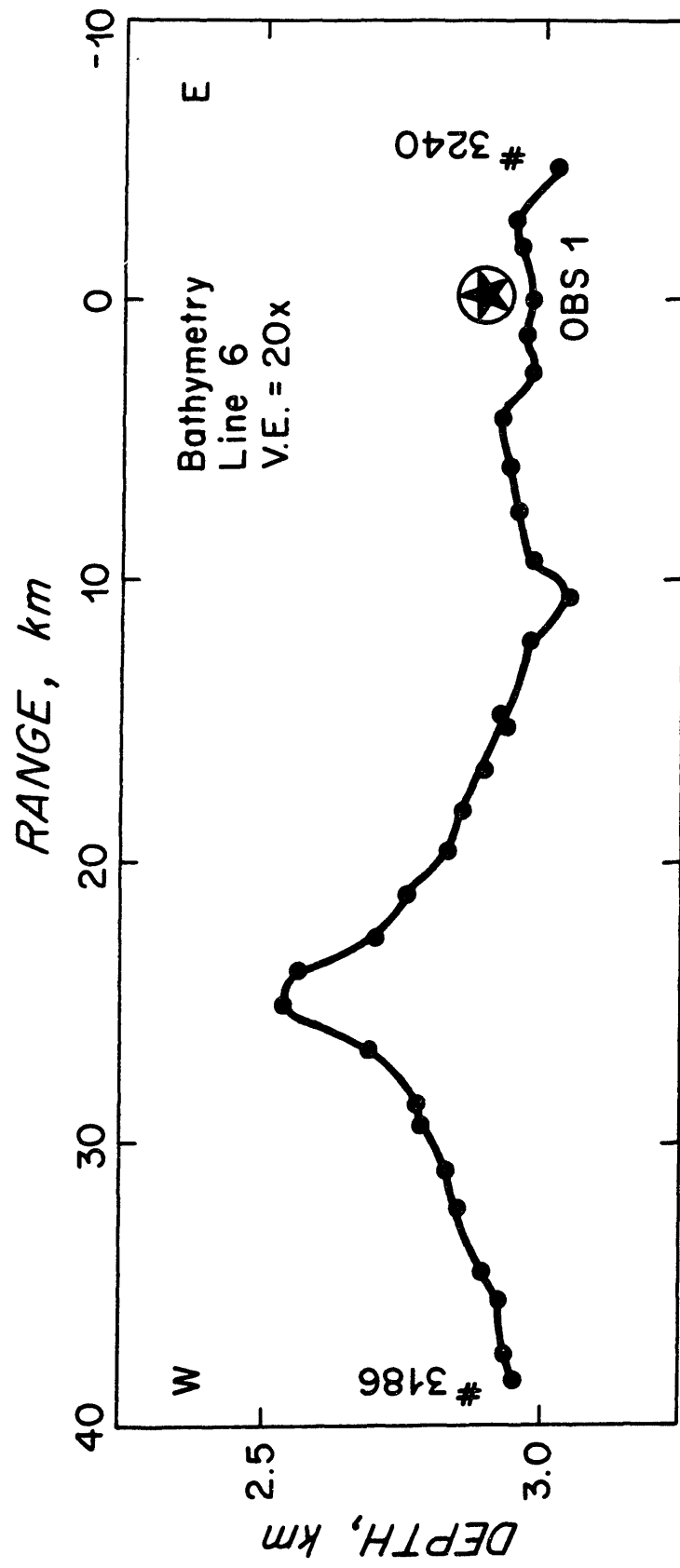


Figure 4a

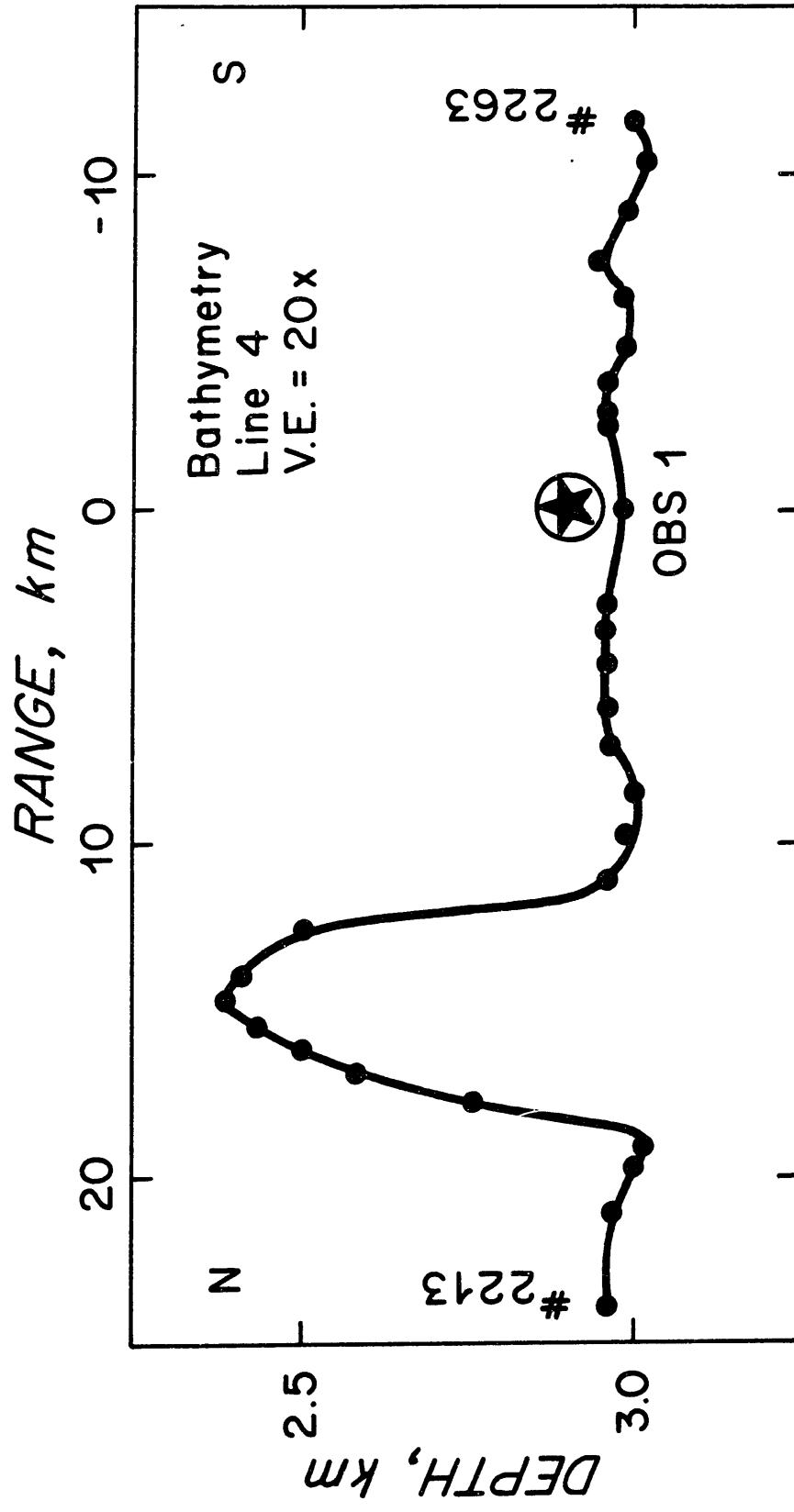


Figure 4b

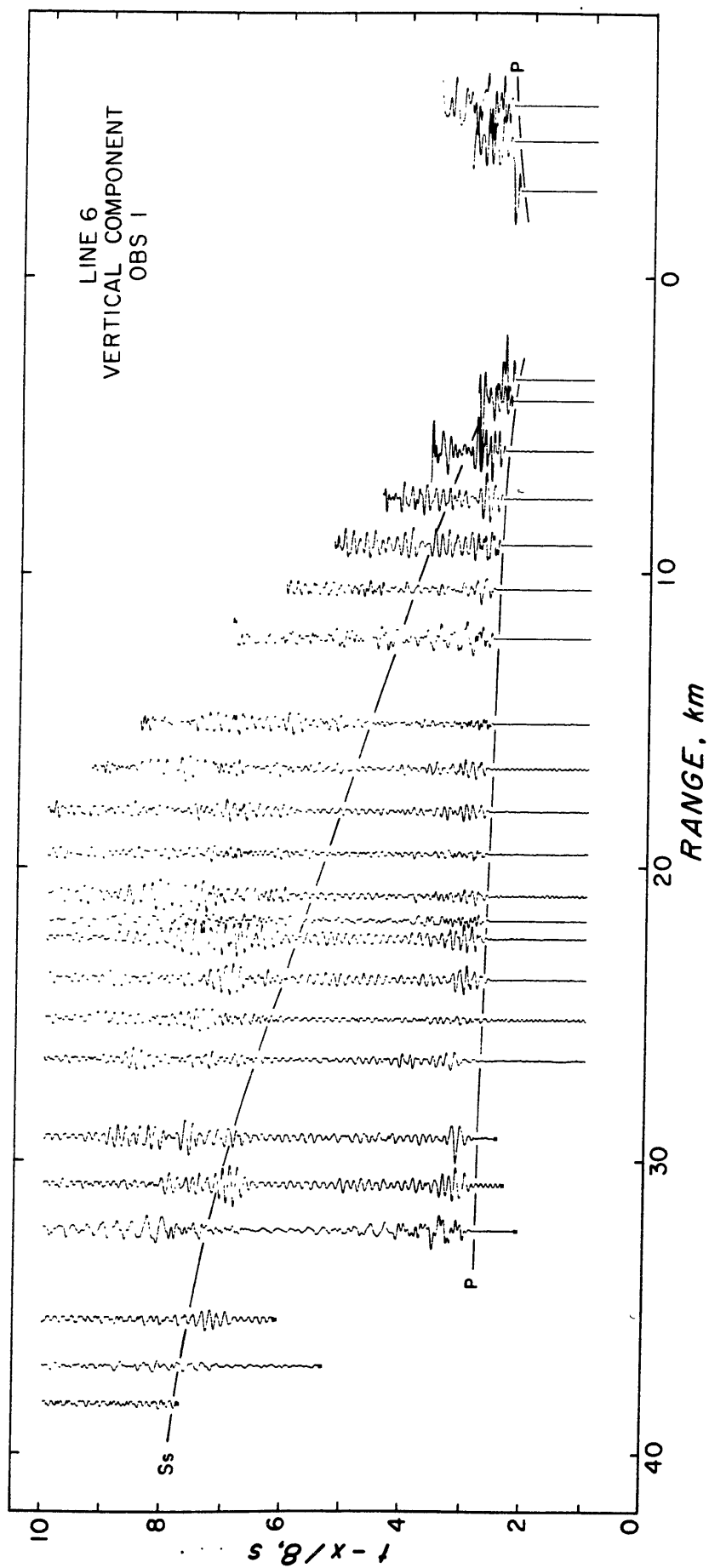


Figure 5a

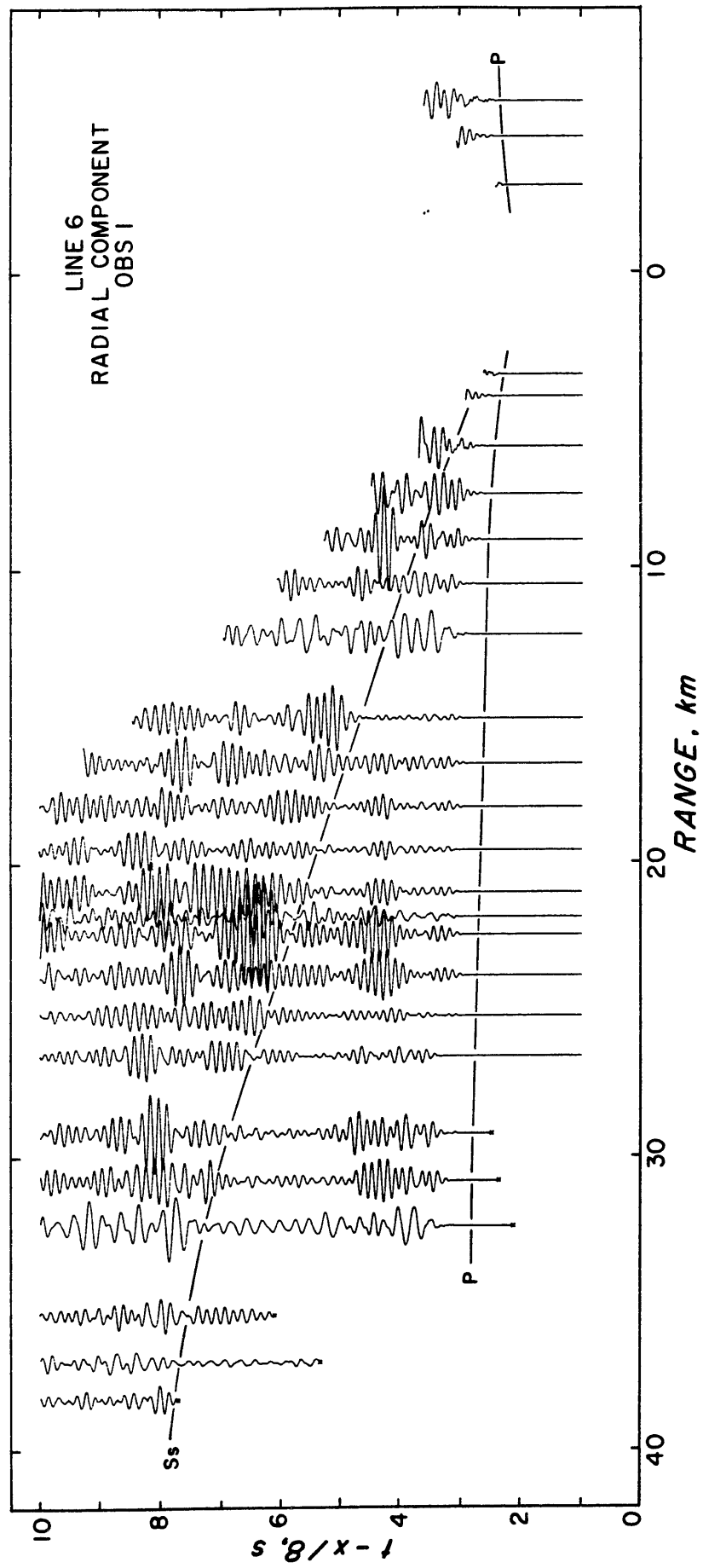


Figure 5b

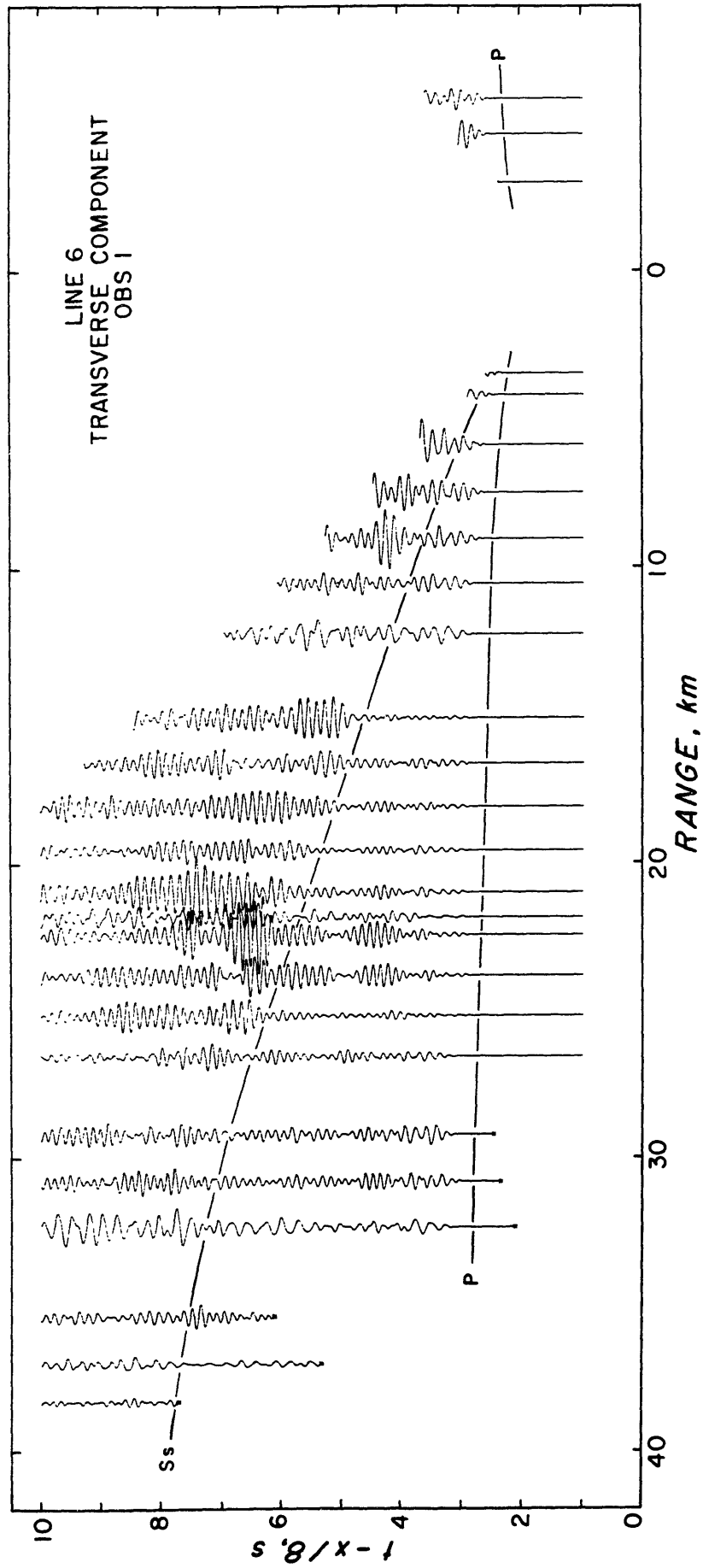


Figure 5c

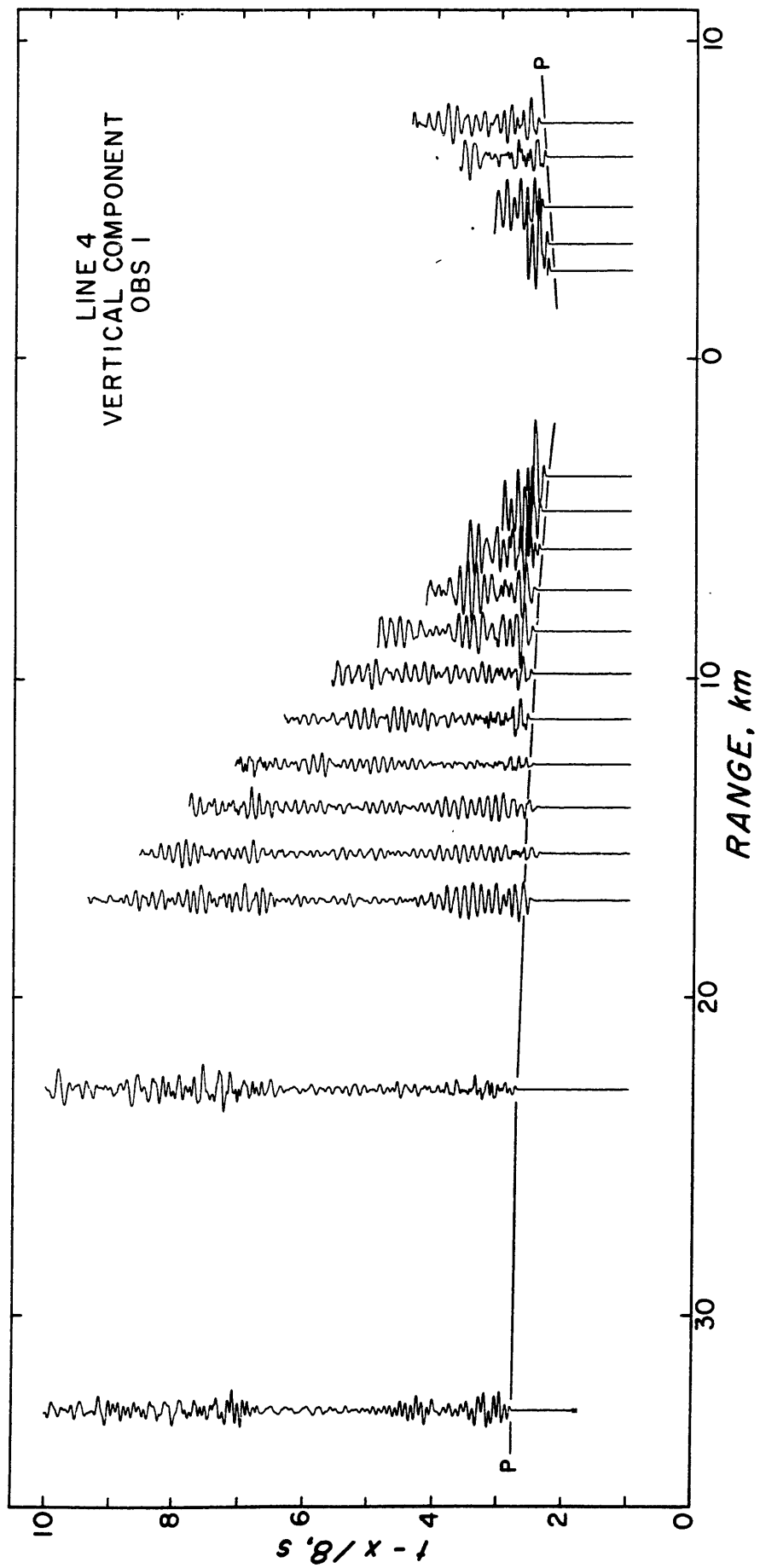


Figure 6a

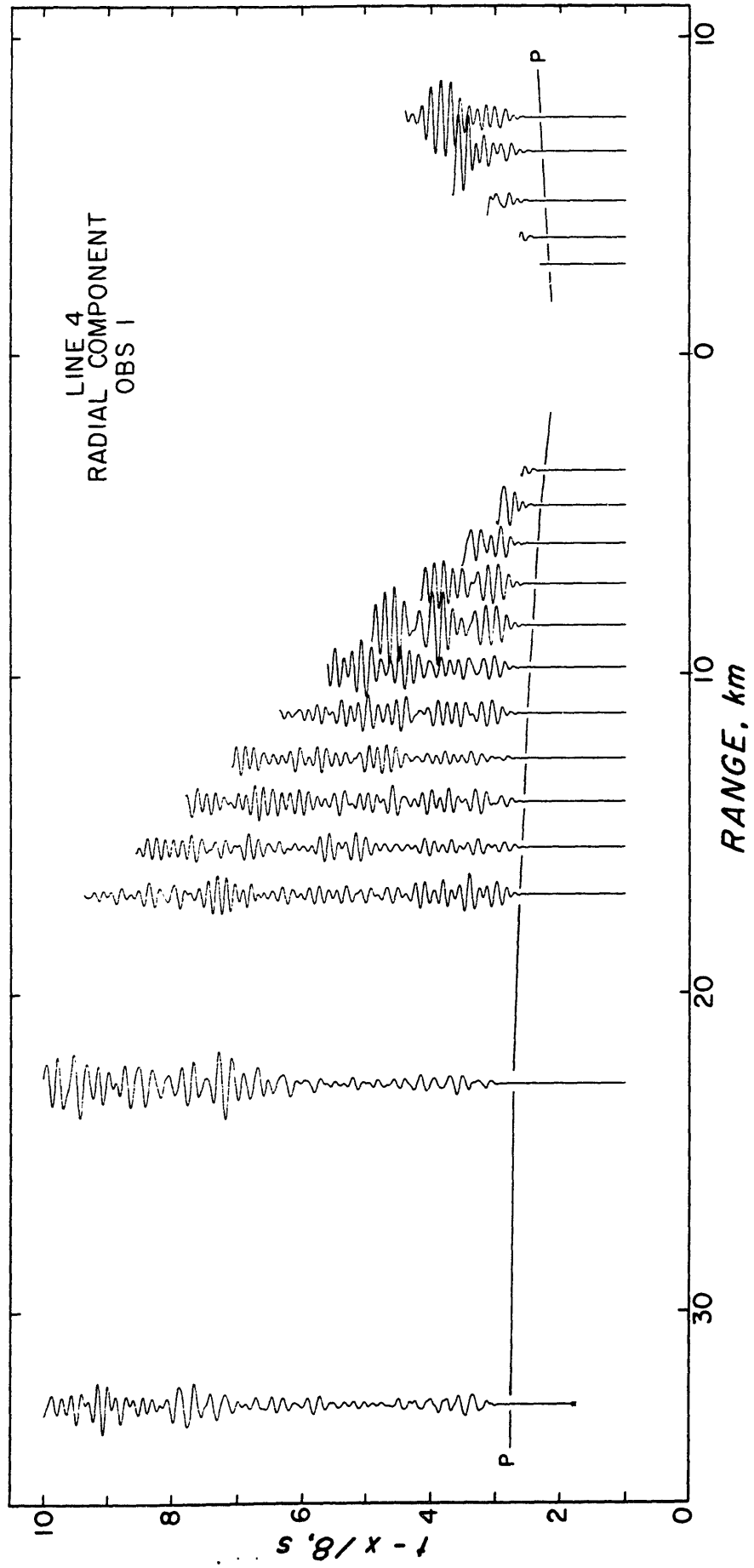


Figure 6b

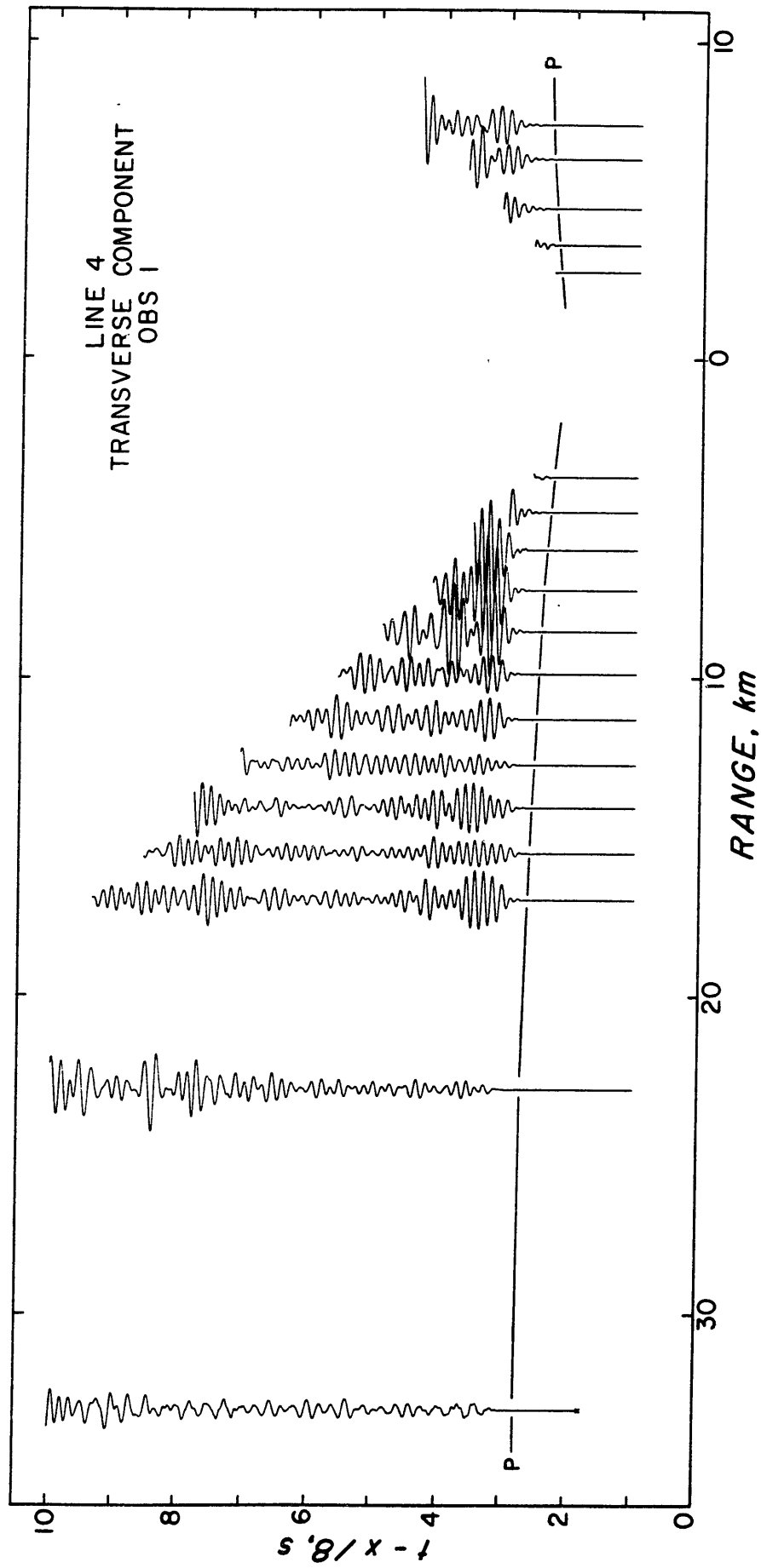


Figure 6c

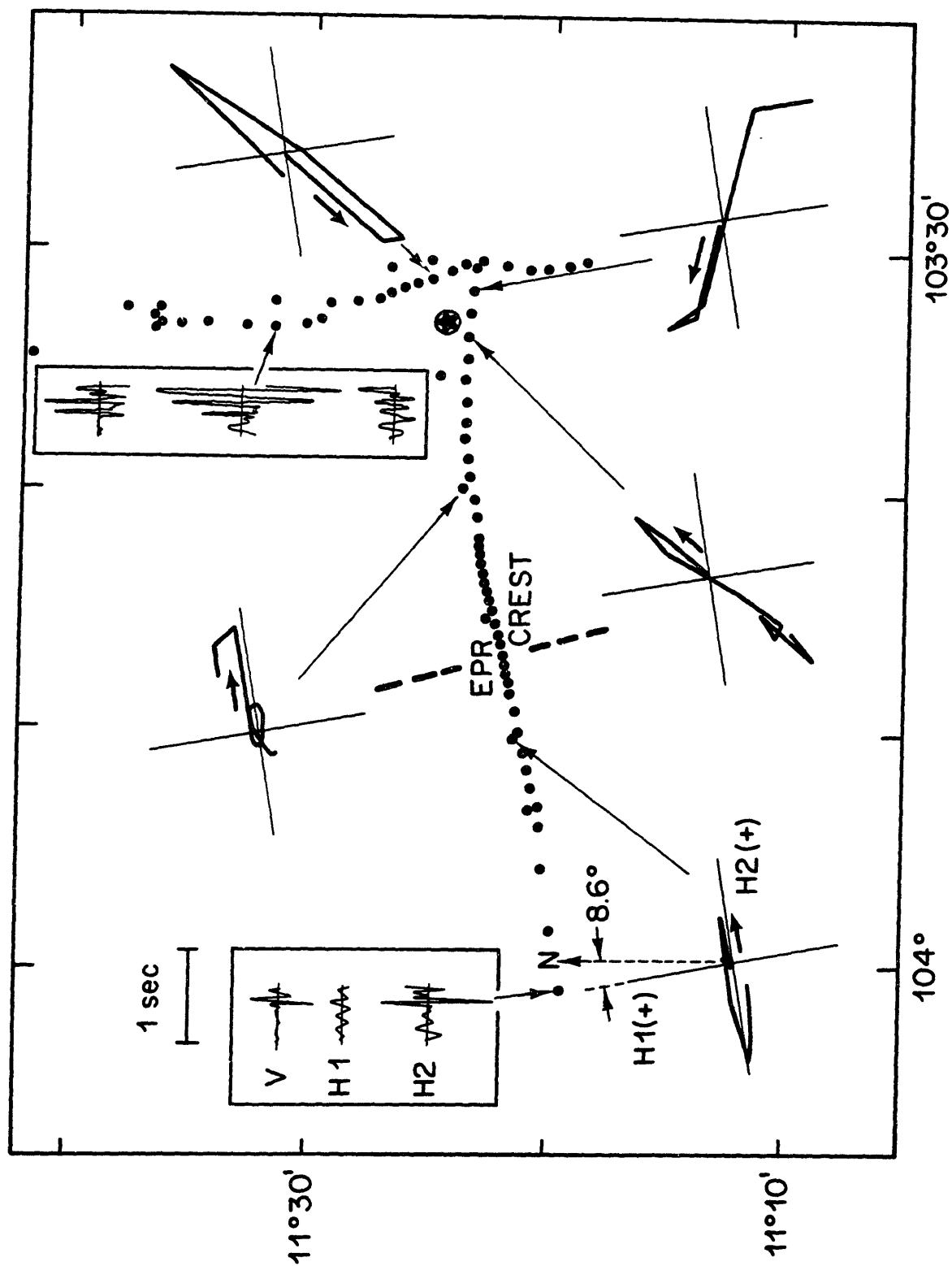


Figure 7

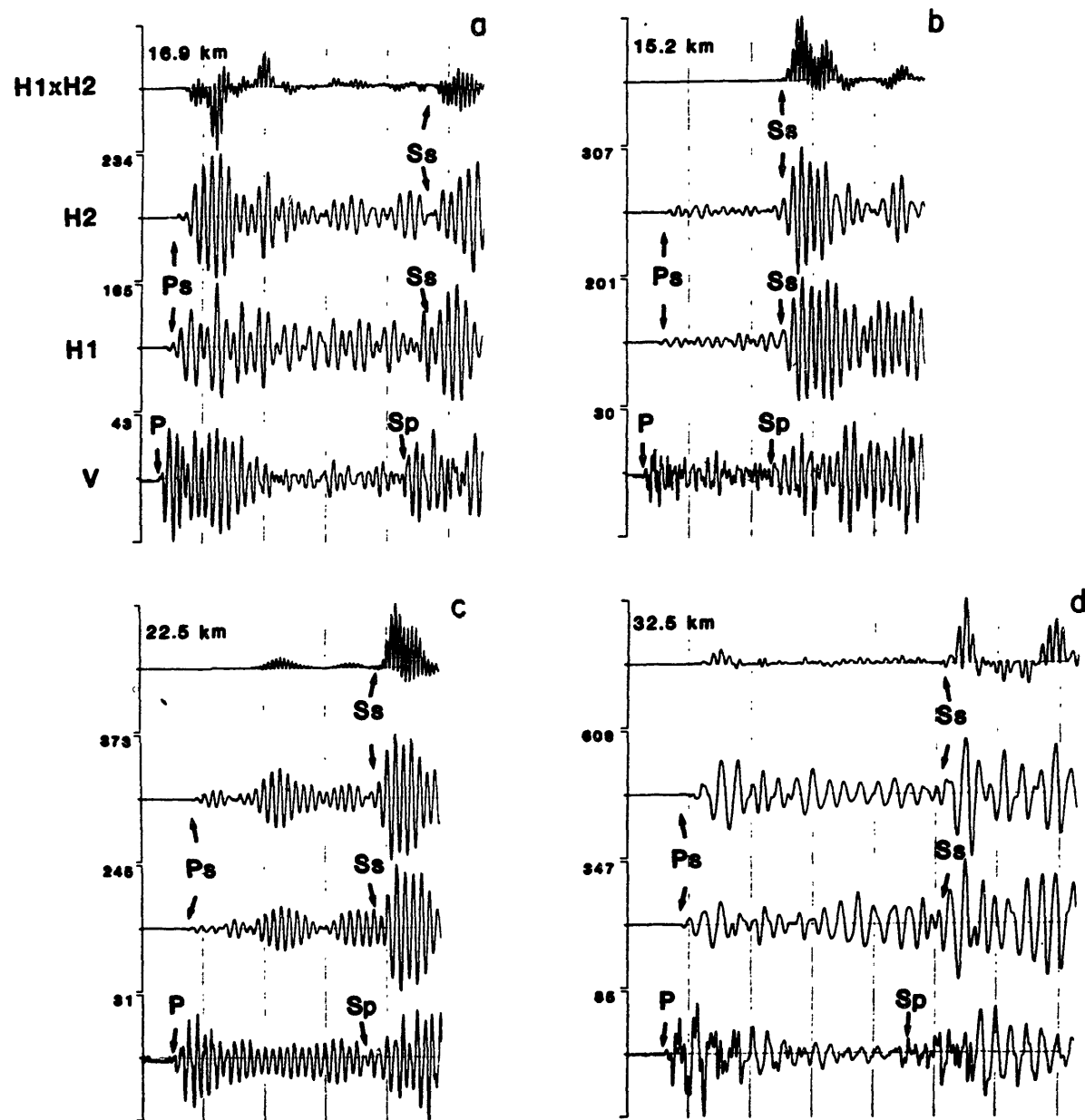


Figure 8

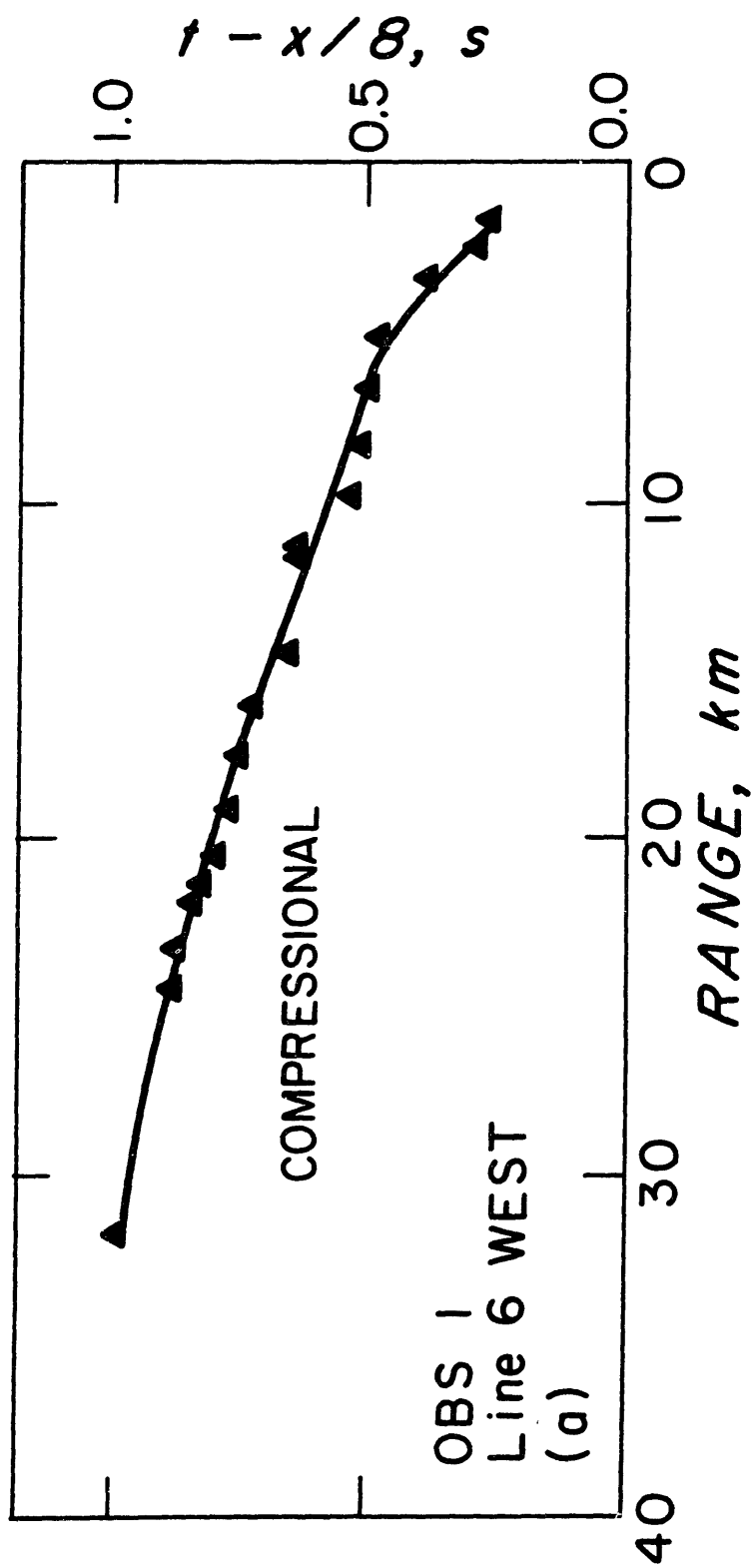


Figure 9a

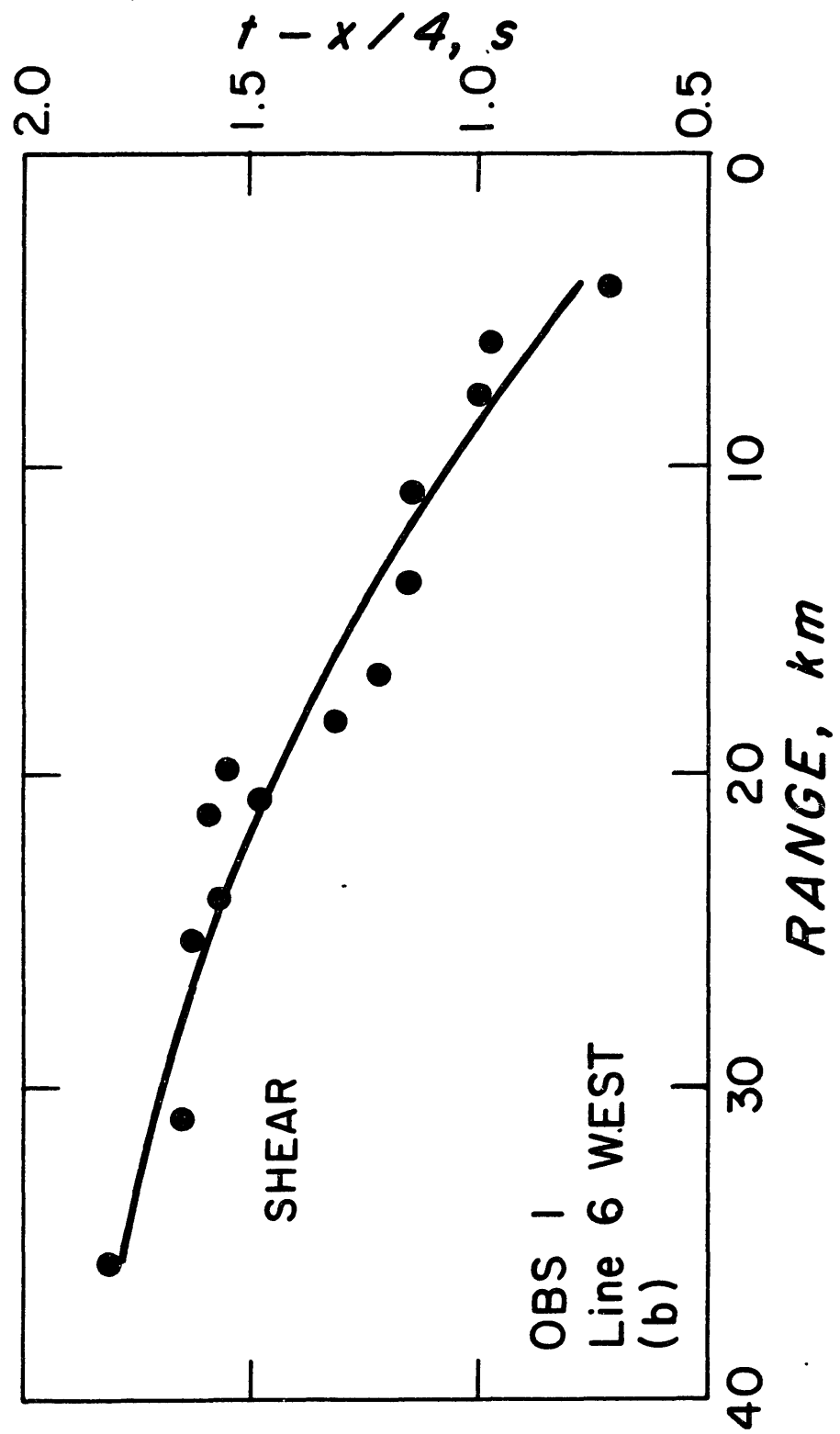


Figure 9b

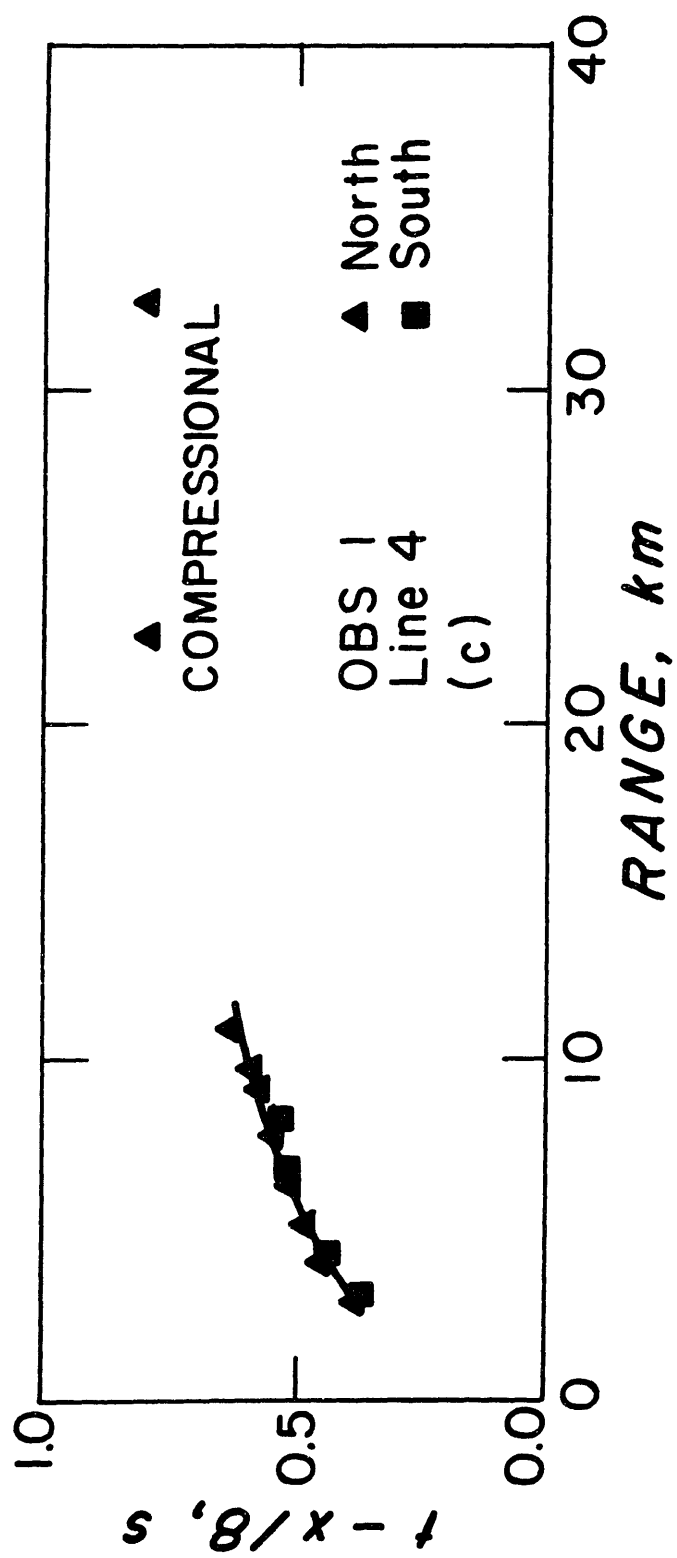


Figure 9c

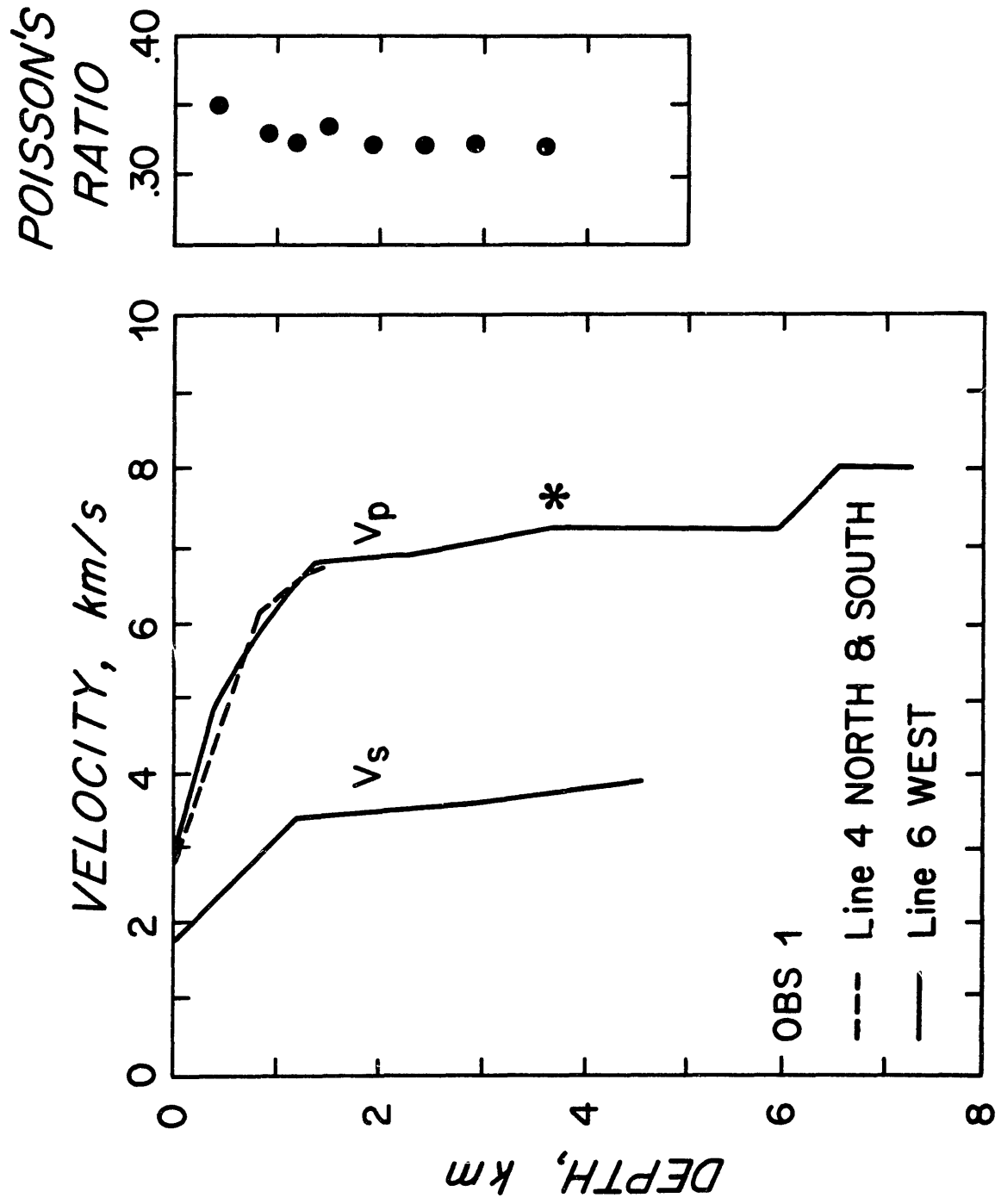


Figure 10

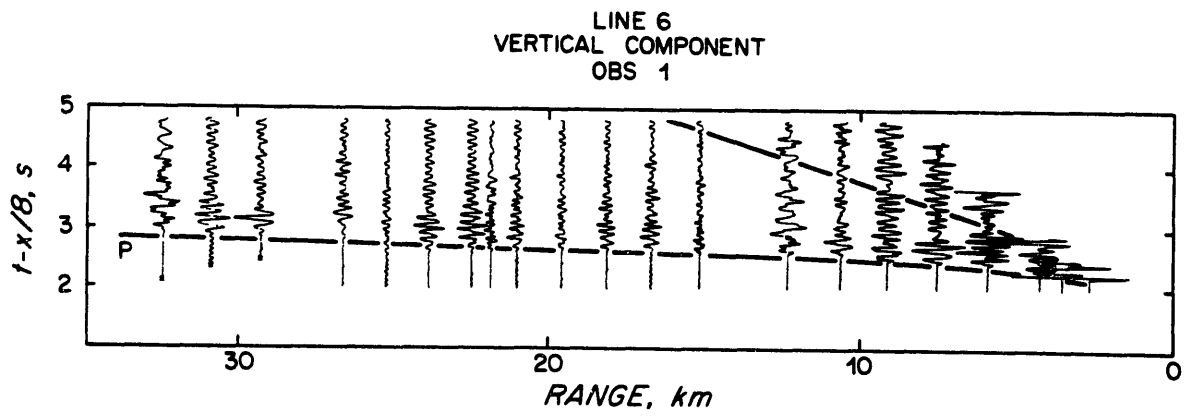
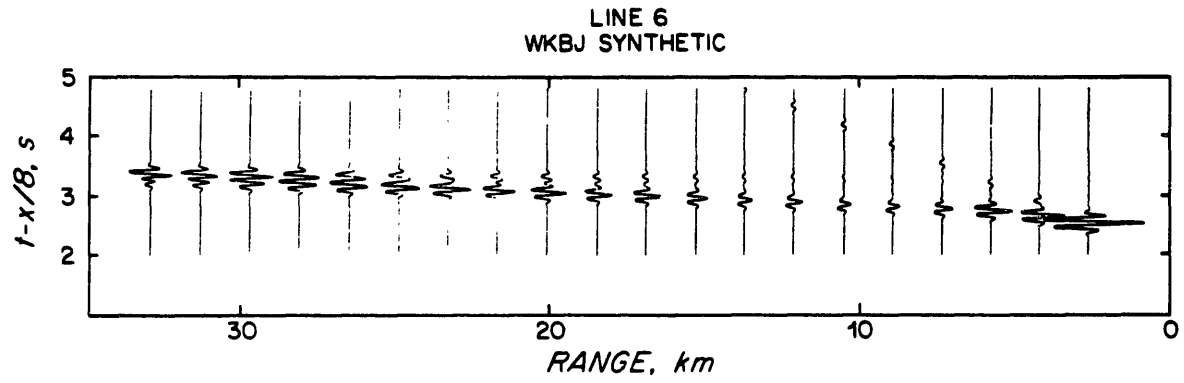


Figure 11

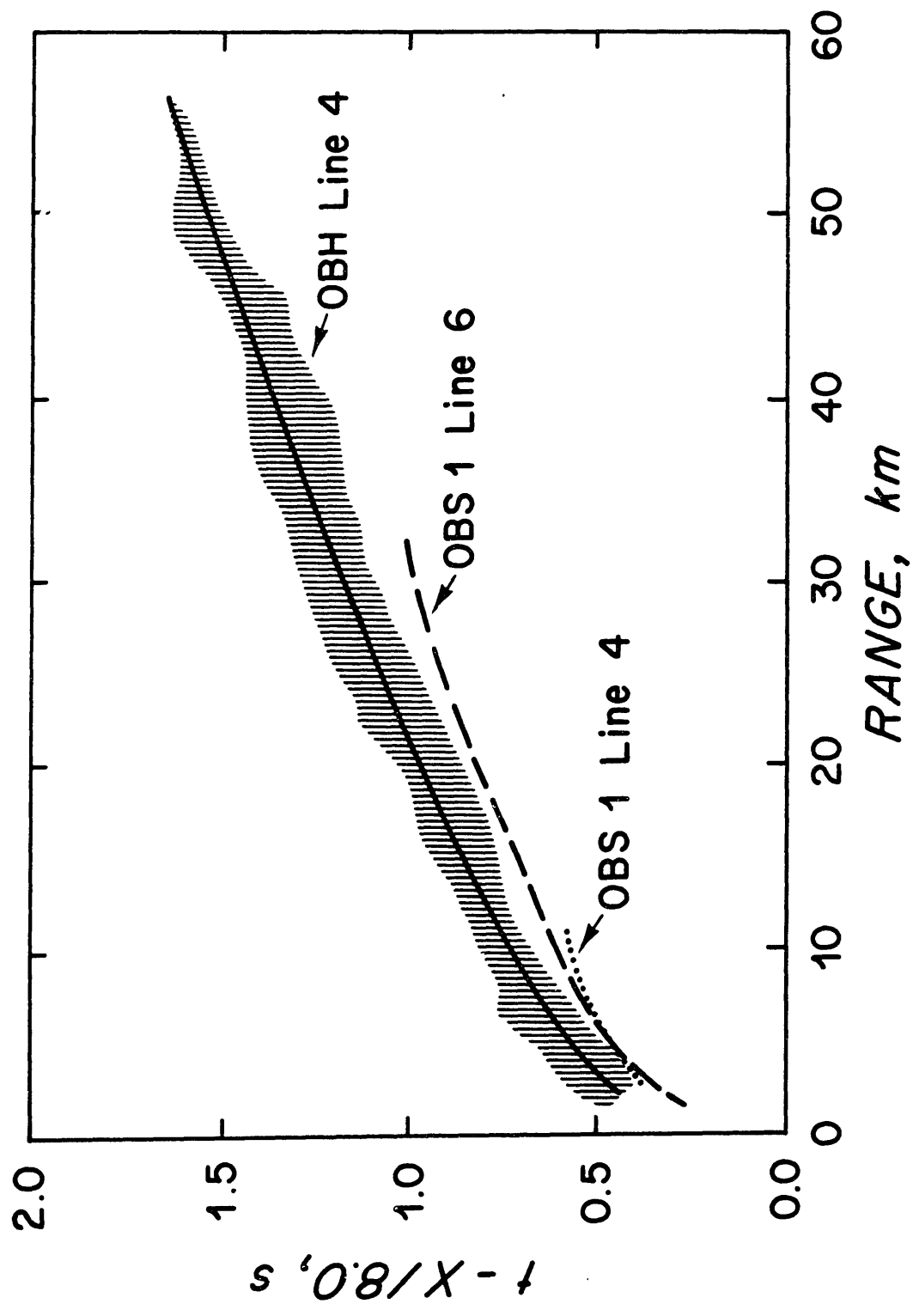


Figure 12

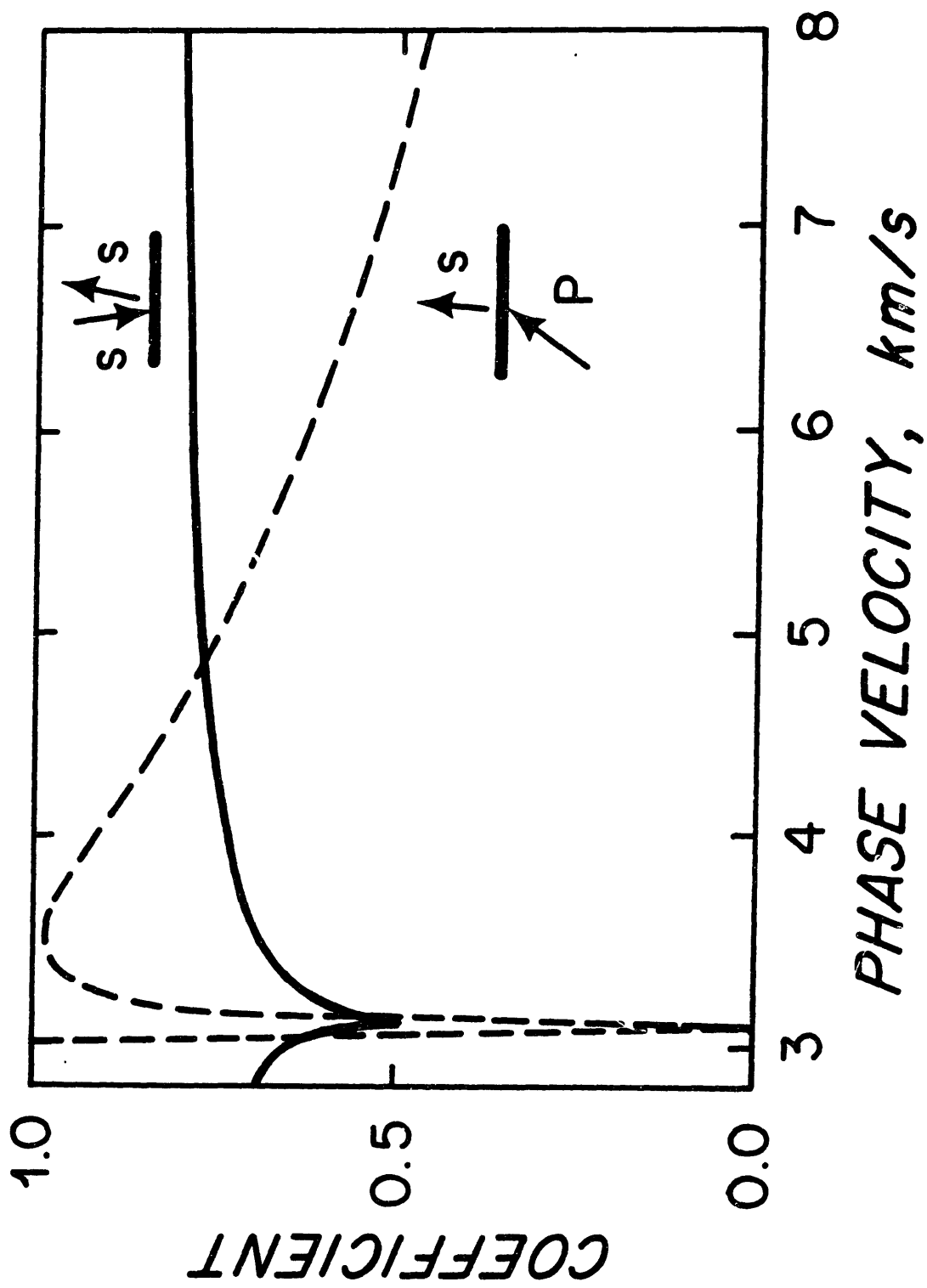


Figure 13

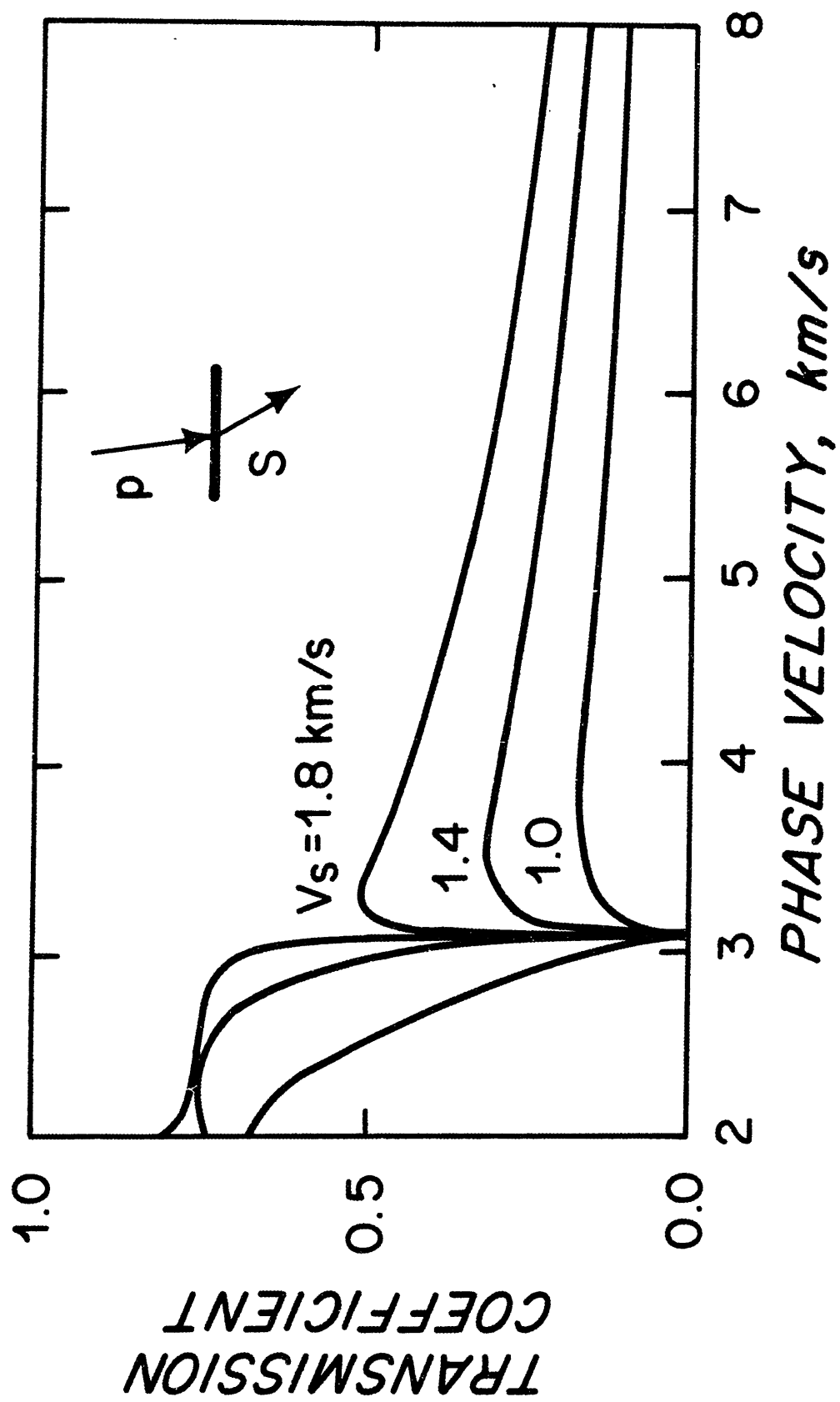


Figure 14

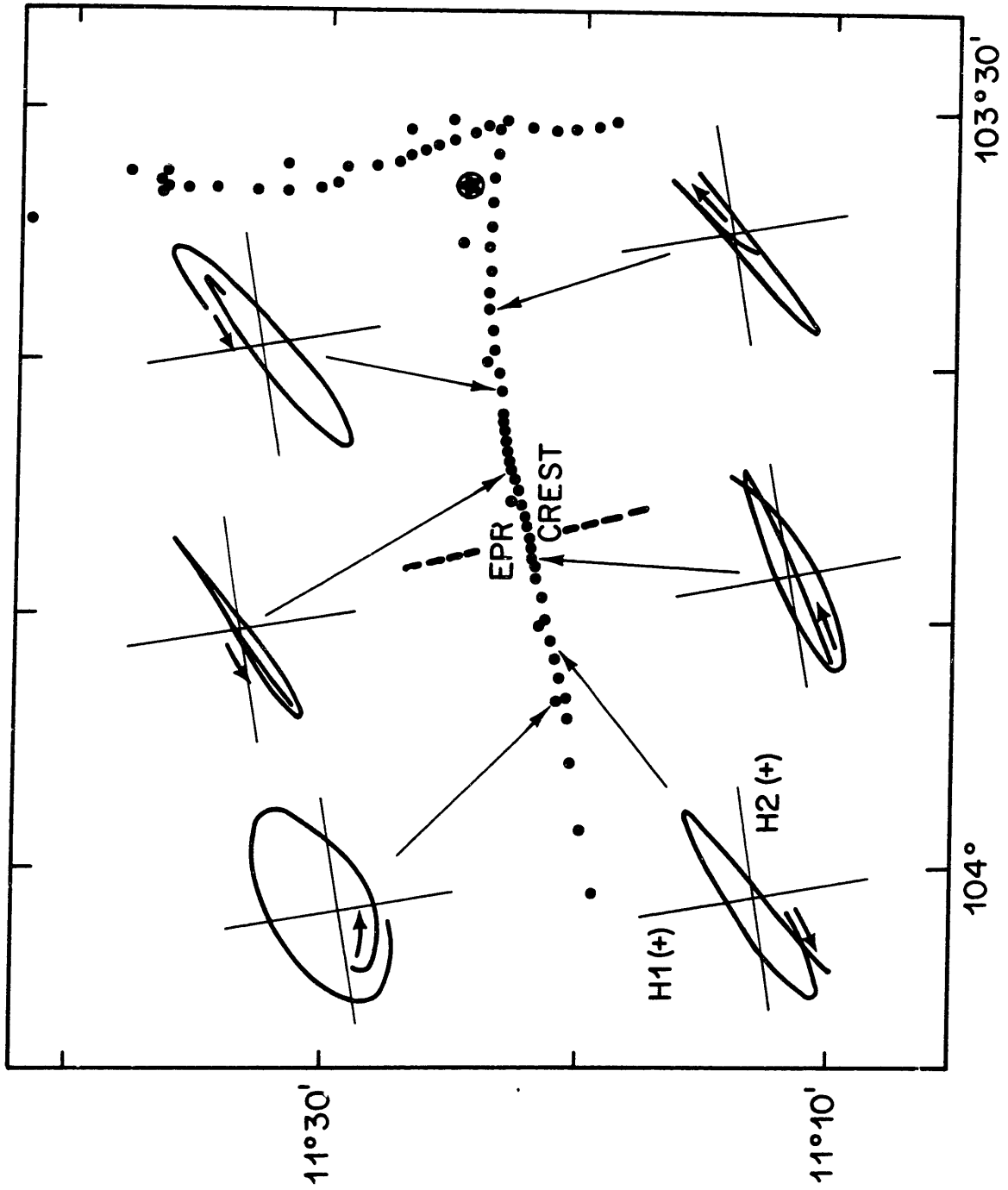


Figure 15

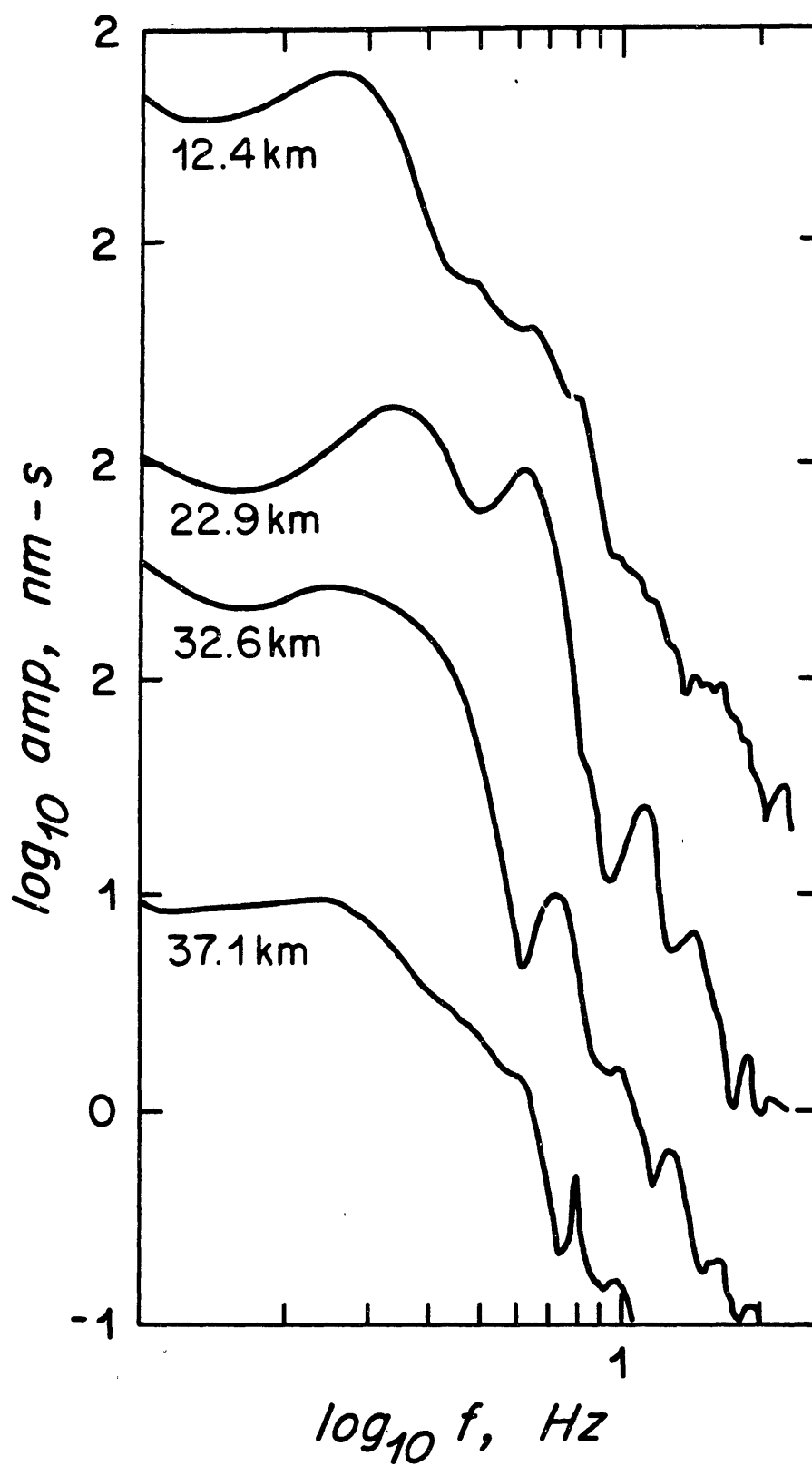


Figure 16

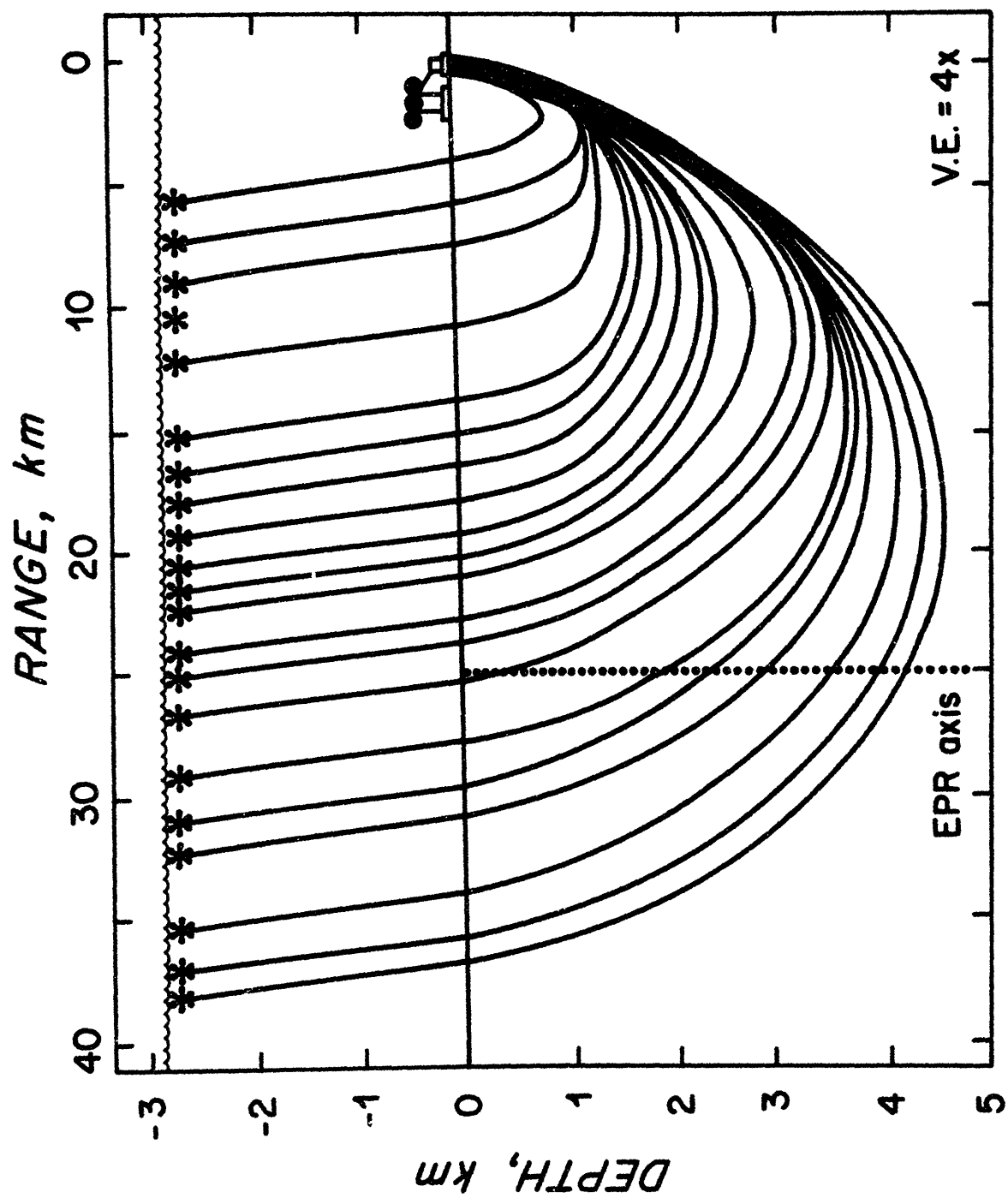


Figure 17

Chapter 3

STRUCTURE AND VARIABILITY OF OCEANIC CRUST ON THE FLANKS OF THE EAST PACIFIC RISE BETWEEN 11° AND 13°N

INTRODUCTION

Two fundamental problems faced by marine earth scientists center around the manner in which oceanic lithosphere is formed at a spreading ridge and the nature of the processes which control the evolution of the crustal structure as it moves away from a ridge. One key to solving these problems lies in our ability to assess accurately the degree to which the structure of oceanic lithosphere varies laterally along the strike of a mid-ocean ridge. Only with an enhanced understanding of the structure and structural heterogeneity beneath a representative set of spreading ridges can we begin to formulate a more general model for the processes of crustal accretion and alteration. As a means to this end, we analyze a large set of explosion refraction data recorded along strike on the East Pacific Rise with a particular emphasis on assessing the degree of structural variability in the young oceanic crust.

Strong constraints on the structure of young lithosphere have been provided by a large number of independent seismic refraction experiments carried out in different tectonic settings. During the last two decades, improvements in experimental and instrumental design and in travel-time inversion and waveform analysis techniques have provided increasingly more complex pictures of velocity structure within oceanic lithosphere. However, variations in instrumentation, experimental design, and interpretation techniques employed in the past make it difficult to

interpret compilations of different marine refraction data sets in terms of structural variability. The many such compilations of the historical refraction data set that have been made [e.g., Spudich and Orcutt, 1980a] have uncovered few systematic structural trends. As a result, our understanding of the processes of accretion and alteration have not benefitted substantially from definitions of structural variability that follow from these compilations.

There is undoubtedly some degree of variability in the seismic structure of ocean crust. However, of great importance both to the assessment of our state of knowledge concerning oceanic crustal structure, as well as to the understanding of the results presented here, is an awareness of the large difference between the very broad range of scales on which we can suppose this variability exists and the relatively narrow band of scales truly resolvable from travel-time and amplitude information collected during conventional seismic refraction experiments.

The large data set presented in this chapter has been interpreted specifically with the goal of quantifying the degree of structural heterogeneity in young oceanic crust that can be resolved by a conventional marine experiment. The refraction experiment was located over a 200 km long segment of 0.5 m.y. old crust on the eastern flank of the East Pacific Rise between the Orozco and Clipperton Fracture Zones and provides a data set of over 500 seismograms recorded by five identical Woods Hole Oceanographic

Institution ocean-bottom hydrophone (OBH) instruments [Koelsch and Purdy, 1979]. We use experimental design and uncertainties in the data as constraints on the spatial scales applicable to this study. We then examine the velocity structure of the young crust, first using travel-time data alone and later incorporating amplitude information to better resolve crustal structure and any variability in that structure over the length of the refraction line.

SEISMIC SCALE

Geologic evidence from submersible observations [e.g., Ballard et al., 1975], analysis of drilling cores [e.g., Aumento et al., 1977], and ophiolite studies [e.g., Casey et al., 1981] indicate that variations on a scale of centimeters to hundreds of meters exist within what is typically called normal oceanic crust. Refraction experiments, however, resolve structural variability on a scale significantly different from that seen in the geologic observations. It is necessary to understand this "seismic scale" before we can analyze our refraction data set, assess the degree of heterogeneity within our experiment, and compare our assessment with those inferred from other experiments.

The seismic scale defines the lateral and vertical extent of both the velocity structure and the variability of that structure resolvable from a particular refraction data set. The lower bound of the seismic scale is determined by the data bandwidth and the lateral data density. Propagating

seismic energy essentially averages material properties over seismic wavelengths so heterogeneities smaller than a wavelength cannot be resolved. For a given velocity-depth structure, lateral shot spacing controls the vertical separation between ray paths and thus sets a limit upon the detail with which one can examine velocity variation with depth under a single refraction line. The accuracy with which a velocity-depth function at any spatial scale can be computed is of course ultimately constrained by uncertainties in the data.

Higher resolution of the velocity structure beneath a single refraction line can be attained by using both amplitude and travel-time information rather than by the use of travel times alone [e.g., Spudich and Orcutt, 1980a]. For example, though the existence of a velocity transition zone can be recognized in the travel-time data, the shape and depth of the transition cannot usually be resolved due to the innate complexity of the travel-time curve, the inadequate shot spacing and the data uncertainty. The amplitude signature of a transition zone is, in comparison, more precisely defined by the data and the thickness and depth of the velocity transition directly affect the distance over which the amplitude event is observed and the range at which it occurs. Amplitude data act as powerful constraints on the existence, shape, and depth of steep velocity gradients in the lithosphere.

An assessment of the degree of lateral heterogeneity in

a region can be made by comparing travel-time and amplitude data or the computed velocity-depth functions from two or more refraction lines. The upper bound of the seismic scale is, then, determined by the number and lateral separation of the refraction profiles being compared. The total horizontal range over which energy from the different refraction profiles samples crust at a given depth defines the range over which variability at that depth can be assessed. The distance between regions of crust sampled under two adjacent refraction lines defines the spatial resolution of the variability determination. A region can be assessed as being laterally heterogeneous on a seismic scale only if the data from two or more independent refraction profiles can be shown to differ outside their estimated uncertainty.

THE EXPERIMENT

Project ROSE (Riviera Ocean Seismic Experiment) was a multi-institutional marine seismic investigation conducted between 11° and 16° N on the East Pacific Rise (Figure 1) in 1979 [Ewing and Meyer, 1982]. As part of the first phase of that experiment, a 200 km seismic refraction line, line 4, was shot parallel to and 25 km east of the rise crest. We focus here on the data recorded by five Woods Hole Oceanographic Institution ocean-bottom hydrophones [Koelsch and Purdy, 1979] deployed along this rise-flanking line. Figure 2 shows the locations of line 4 and of the OBHs (numbers 1, 3, 4, 7, and 8) used for this study.

The spreading rate in the ROSE area is 104 mm/yr (full-rate) with an azimuth of N80°E [Mammerickx and Klitgord, 1982]. Line 4 is located within the Brunhes polarity epoch and samples 0.5 m.y. old crust. The average water depth beneath line 4 is about 2900 m (Figure 3). The major bathymetric feature along the line is a 600 m near-axis seamount located south of OBH 8. The closest major fracture zone is the Clipperton which lies 100 km to the south. A fracture zone inferred from a small offset in magnetic lineations [Klitgord and Mammerickx, 1982] near OBH 7 has no obvious bathymetric signature. Using Seabeam bathymetry, Macdonald and Fox [1983] detected an overlapping spreading center at the intersection of this fracture zone and the rise crest (12°N latitude) 25 km west of OBH 7.

A total of 277 alternating 2 and 11 kg charges were shot from the R/V THOMAS G. THOMPSON at 0.75 km intervals. Because the dominant frequency of the data is near 8 Hz (i.e., the bubble pulse frequency of the explosions), the lower bound of the seismic scale is about 0.5 km in the upper crust and 1.0 km at the base of the crust. The product of the experimental phase of this study was ten seismic refraction profiles (five split spreads) collected by the five OBHs. Two typical record sections are shown in Figure 4. Two important features of these data are the high amplitude arrivals near 30 km range characteristic of the oceanic Moho triplication, and the lack of energy at phase velocities close to 8 km/s that would be generally associated

with oceanic upper mantle. These ten refraction profiles give effectively ten determinations of the phase velocity at a given depth over a 200 km long strip of 0.5 m.y. old Pacific crust.

DATA REDUCTION AND UNCERTAINTY

A rigorous evaluation of the velocity structure and the magnitude of the lateral variations in that structure should include a realistic estimation of the errors associated with each step of the data reduction process. Though summaries of errors commonly related to refraction data corrections have been discussed elsewhere [e.g., Kennett and Orcutt, 1976], we take the time here to assess the uncertainties which can be attributed specifically to our data set and to the assumptions implicit in reducing our data to a form from which arrival times can be analyzed. Though our estimation of uncertainties is necessarily somewhat subjective, we consider the errors examined below to approximate 95% confidence bounds on the corrections and measurements being discussed.

Shot instant and timing corrections were applied to all records. Errors associated with instrumental corrections such as these are usually ± 0.02 s [Purdy, 1982a]. Shot-receiver ranges were determined from water wave arrival times picked from a high frequency data channel. The resulting relative shot locations are accurate to about 10 m. The arrival time errors resulting from range determination uncertainties are greatest for low phase velocity arrivals

from the shallow crust, but are still small in relation to other reduction errors.

We were able to pick 479 first arrivals from the collection of seismograms recorded on our ten refraction profiles. Almost all of the shots detonated from a range of 40 km or less are included in that data set. The only shots excluded from this group were those detonated directly over the seamount (located 60 km south of OBH 1, Fig. 3), those for which the first refracted arrival was masked by the direct water wave, and those that failed to detonate properly.

Assuming that picking errors are approximately normally distributed, Kennett and Orcutt [1976] suggest that the standard deviations of travel-time measurements generally range from 0.05 to 0.15 s depending on the shot-receiver offset. Their 95% confidence level (two standard deviations) is at best 0.10 s for close shots in a typical refraction experiment. The total picking error is affected by both the limited bandwidth of the shot-receiver system [Bracewell, 1965] and a seismologist's inability to consistently pick the first motion from a seismogram. Human picking error was evaluated by repicking selected portions of the data set. The total uncertainty involved in measuring first arrivals varied from shot to shot, but was generally estimated at 0.03 s for close shots to 0.10 s for distant shots.

Errors in making topographic corrections are highly dependent on the assumptions under which the corrections are

made. A water path correction [Purdy, 1982b] was applied because of the mild topography and sufficient data density. This correction requires no assumptions concerning the velocity structure of the shallow crust. Using water depth under the shot and apparent slowness from the observed travel-time curve, the water path correction does not in fact correct for the seafloor topography, but effectively places the shot where the ray from that shot to an OBH intersects the seabed. Accurate determination of the apparent phase velocities presents the largest source of uncertainty in the application of a water path correction.

For example, given a phase velocity near 4 km/s and a water depth of 3 km, an error in the calculation of the apparent phase velocity of about 0.2 km/s produces an effective error in the water path correction of about ± 0.02 s [see Figure 2, Purdy, 1982b]. Errors dependent on the apparent phase velocity are less severe at high phase velocities. However, uncertainties in the water depth at the ray entry point bring the accuracy of the correction to within ± 0.01 s for high phase velocity arrivals (>6 km/s) and ± 0.03 s for low phase velocity arrivals (<5 km/s).

One final source of uncertainty should strictly be associated with the interpretation phase and not the reduction phase of this study. These errors are due to departures of the real crust from the assumptions used to interpret the travel time data. Because velocities in the upper crust near a mid-ocean ridge are probably controlled by

porosity and crack geometry [e.g., Spudich and Orcutt, 1980a], a likely first order model for the velocity structure is one in which the iso-velocity contours parallel the bathymetry [Purdy, 1982a]. Implicit in most travel time analysis techniques is the assumption that the crust is laterally homogeneous. To estimate the possible uncertainty associated with this discrepancy, Purdy [Figure 18, 1982a] compared water path corrected travel-time data from the ROSE area to a travel-time curve derived by ray tracing through a model crust in which velocity contours followed observed seafloor topography. The two travel-time curves were within ± 0.02 s of each other with errors being worse near the most severe topography. The relief along line 4 is of similar magnitude to that in Purdy's [1982a] study. We therefore accept ± 0.02 s as the "error" due to the incorrect assumption of lateral homogeneity. This uncertainty itself is of course insignificant compared with the total of the other uncertainties; therefore, the detection of near surface heterogeneity of this type is beyond the resolving power of travel-time information alone.

A list of all errors discussed above for three different ranges is shown on Table 1. It is difficult, however, to assess the cumulative uncertainty due to the four contributing error sources because of our incomplete understanding of the character (i.e., probability density function) of each individual error. We instead attempt to place bounds on the total uncertainty associated with the

travel-time measurements. If we assume that the errors are normally distributed, our error estimates (95% confidence bounds) would represent a value equal to two standard deviations from the most likely travel time. With the additional assumptions that the errors (E_i) are random and independent, we can estimate the total error by:

$$E_{T0} = \sqrt{\sum_{i=1}^4 E_i^2}$$

We know, however, that the errors are neither independent nor random. Because the first arrival picks, the water path correction, and the effects of assumptions about lateral homogeneity in our interpretation are interrelated, the associated uncertainties are likewise dependent. Any dependence among the variables increases the accumulated error [Mosteller and Rourke, 1973]; thus E_{T0} represents only a minimum estimate of the total uncertainty.

In the worst of all possible situations, all of our estimated errors would affect a measurement in the same sense such that the travel time would be the actual travel time, plus or minus an error equal to:

$$E_{Tp} = \sum_{i=1}^4 E_i$$

The bottom of Table 1 lists our estimates of the minimum and maximum 95% confidence bounds on the travel-time measurements. Our most optimistic assessment of the error (E_{T0}) at any range is 0.05 s and the most pessimistic estimate (E_{Tp}) is 0.11 s.

TRAVEL TIME INTERPRETATION

The ten travel-time curves in Figure 5 show the water path corrected first-arrival travel times as a function of range. The lack of typical upper mantle phase velocities (8 km/s) on any of the plots in this figure suggests that many of the picked arrival times at ranges greater than 30 km are most probably supercritical Moho reflections. It is, therefore, possible to infer little about the total crustal thickness or the velocity of the upper mantle from an interpretation of these travel times alone.

To assess the structure and heterogeneity of the upper crust, we follow the procedure used by Purdy [1983]. Instead of inverting ten individual travel-time curves to derive ten separate velocity profiles (and then discussing the significance of their differences), we first compare the total data set by combining the 479 first arrival times from all the curves into one travel time versus distance plot. Figure 6 shows the merged arrival times before the application of a water path correction compared with the merged set of ten corrected travel-time curves. We next fit a piecewise parabolic spline, constrained to have a negative second derivative, through the merged set of corrected travel times. The error bars in Figure 6 denote the most "optimistic" and most "pessimistic" total uncertainty in our data (Table 1). The fit to all the data shown has a root mean square deviation 0.05 s, which equals the smallest possible assigned 95% confidence bound (Table 1). We thus

conclude that, to the limit of resolution of our travel-time data alone, the velocity structure of 0.5 m.y. old crust is homogeneous in this region of the Pacific.

The curve fit through the merged, corrected travel times can be used as a constraint on the mean velocity structure of the upper crust beneath line 4. We follow the work of Ewing and Purdy [1982] to determine a suitable linear velocity gradient for the uppermost crust and then invert directly the mean travel-time fit in intercept-slowness (τ - p) space for a velocity structure which increases monotonically with depth. The resulting upper crustal velocity structure is depicted in Figure 7. This model for young Pacific ocean crust has a velocity gradient of 4.4 s^{-1} in the top 0.8 km and a smooth transition to a gradient of 0.2 s^{-1} at a depth of 2.5 km. The velocity at the seafloor is 2.1 km/s and at a depth of 3.8 km is 7.0 km/s.

Travel-time analyses of other refraction data sets collected during project ROSE provide additional information on the structure and variability of young lithosphere in a single tectonic province. A study near line 4 by Purdy [1982a] suggests that a 20 km section (line 2E2W) of shallow crust (uppermost 2 km) parallel to the spreading direction is homogeneous on a seismic scale within the limits of resolution of the data. In addition, the velocity-depth function for line 2E2W is in close agreement with the structure under line 4 determined in this study.

Our travel-time inversion for a mean velocity structure

along line 4, with the exception of the upper 250 m, also fits within the bounds designated by Gettrust et al. [1982] for structure under the same line. The two studies differ in the methods used to calculate the upper crustal velocity. Also the velocity structure derived from travel-time data collected to the south is insignificantly different from that determined from data recorded to the north, thus supporting our conclusion that the crustal structure along strike in this region is homogeneous on the seismic scale defined by the travel-time data.

The principal difference between the crustal structure determined here and that derived in Chapter 2 for crust immediately to the south on lines 4 and 6 is that the compressional velocity structure of the uppermost crust is between 0.5 and 1.0 km/s higher in the latter region. They suggest the difference between the observed velocity structures may be attributable to the lower bulk porosity of the basement to the south. Possible contributors to the lower porosity there may be basement of inherently low porosity because of larger scale lava flows, or basement on which a sediment layer acts as a capping blanket to hydrothermal circulation thus accelerating the cementation of cracks and pores by hydrothermal mineralization [Stephen and Harding, 1983].

Results of the travel-time analysis presented here support a conclusion derived from a similar study on 140 m.y. old Atlantic crust [Purdy, 1983]: Variability in the

processes that control crustal formation and its subsequent alteration is small on the scale defined by refraction travel-time data. Additional information can be drawn from a comparison of a set of 183 merged first-arrival travel times from Purdy's [1983] study of 140 m.y. old Atlantic crust with the travel times examined in this study for 0.5 m.y. old Pacific crust (Figure 8). Travel time through a thick sediment blanket has been subtracted from the travel times of the data collected on the older crust to facilitate a comparison of the structures of the igneous crust beneath the two study areas. A consistent difference of about 0.2 s between the mean travel-time curves fit through the two data sets suggests that the velocity of the shallow crust is greater in the older Atlantic lithosphere. In fact, the major differences in the crustal structures for young Pacific and old Atlantic crust (Figure 7, Purdy [1983]) are the velocity at the top of the basement (2.1 and 5.0 km/s, respectively) and the velocity gradient in the upper 0.5 km of basement (4.4 and 1.1 s⁻¹, respectively). This comparison is in agreement with the results of Houtz and Ewing [1976] indicating that an increase in the velocity of the uppermost portion of layer 2 represents perhaps the most significant change in the evolution of oceanic crust with age.

AMPLITUDE INTERPRETATION

As mentioned earlier, the use of amplitude information in addition to first-arrival travel times effectively

decreases the lower bound of the seismic scale by substantially enhancing our ability to resolve details in the velocity structure of oceanic lithosphere under a refraction line. Therefore, comparing amplitude characteristics of phases as they appear on different refraction sections can lend additional insight into the degree of structural variability in the crust and upper mantle on a regional scale.

As is true for travel-time data, there is uncertainty involved when comparing waveform information as a means of assessing heterogeneity beneath the seafloor. Many factors other than the velocity-depth profile within the lithosphere can affect the observed amplitude patterns. Focussing and defocussing of seismic energy by bathymetric features can influence the amplitude characteristics of individual seismograms [Purdy, 1982a]. Misfired charges and variability in their size and detonation depth can also hinder a direct comparison of any two refraction profiles. The uniformity of both the explosive sources and the five receivers in this experiment improves our confidence in our ability to compare different record sections.

Though it is difficult to evaluate heterogeneity in terms of quantifiable uncertainties in amplitude-range relationships, we can compare major amplitude patterns on all profiles and qualitatively assess observed similarities and differences. Because of the previously mentioned effects of topography and source characteristics on recorded waveforms, interpretations of a velocity structure in a region based on

the amplitude information from a single refraction section could prove erroneous. We therefore model only those characteristics common to two or more profiles to provide information about the mean lower crustal and upper mantle structure and to further constrain the upper crustal structure. Figure 9 shows a typical record section (11 kg charges north of OBH 7) compared to the WKBJ [Chapman, 1978] synthetic profile which best fits the general characteristics of all ten observed refraction sections.

To constrain the velocity structure from a depth of 4 km below the seafloor to the upper mantle, we matched four general waveform patterns appearing at ranges greater than 25 km on the 10 refraction sections. (1) The strength of super-critical Moho reflections (PmP) relative to the lower crustal refractions (P₃) and the observation of PmP out to ranges as great as 60 km suggests that the vertical velocity gradient in the lowermost crust is near zero. (2) The PmP phase is most energetic from a range of 25 km to about 50 km. Fixing the Moho transition zone to be 1.0 km thick best accounts for the appearance of PmP in this interval. A thicker transition zone would focus PmP into a smaller range window. (3) The interaction of PmP and P₃ phases between 25 and 40 km allows a determination of total crustal thickness. The region of high amplitudes resulting from PmP-P₃ interaction is best matched with a total crustal thickness of 5.5 km. This is in close agreement with other crustal thickness determinations made in the ROSE area by Lewis and

Garmany [1982] and in Chapter 2. (4) The absence of strong arrivals at upper mantle velocities makes it difficult to conclude much about the structure of the upper mantle, but does indicate that the velocity gradient in the upper mantle is near zero or possibly negative. We assume a velocity of 8 km/s at the top of the mantle based on Lewis and Garmany's [1982] determination of P_n velocities along strike near line 4. Figure 7 shows the velocity structure of the crust and upper mantle under line 4 determined from both the first arrival travel time analysis and the modelling of amplitudes on ten seismic refraction profiles.

Of the above four amplitude characteristics, (1) and (4) are consistent throughout the ten refraction sections. This indicates that the velocity gradients in the lower crust and upper mantle are similarly near zero on a seismic scale along the entire 200 km line. However, because the character of patterns (2) and (3) vary somewhat within the range bounds placed on them in the preceding paragraph, we may be observing the effects of heterogeneity in the shape and depth of the Moho. We do not know how much of the observed variability is due to topographic focussing and defocussing or irregular source characteristics. Because upper mantle refractions are not observed, the velocity and velocity gradient in the upper mantle are not well constrained and thus a thorough investigation of the degree of structural heterogeneity at and beneath the Moho is not possible. We assert that the structure and depth of the Moho shown in

Figure 7 represents a well constrained mean for the region of line 4.

Based mainly on comparisons of amplitude relationships seen on two refraction profiles from line 4, Gettrust et al. [1982] concluded that the Moho may vary in depth along the rise-flanking line. They proposed a difference in the depth and character of the Moho over a 40 km distance based on a 5 km difference in the location of an amplitude increase near a range of 30 km. Whether this is within the resolution of their data is unclear. We agree with their conjecture that the seamount on the southern part of the line and other topographic perturbations might have influenced their amplitude patterns.

Two amplitude patterns ubiquitous throughout the data set provide information about the structure of the upper crust that is complementary to that already obtained from the first-arrival travel time interpretation: (1) A first order discontinuity (velocity changing from 4 to 6 km/s) at a depth of 1 km such as that suggested by Lewis and Garmany [1982] would be expected to produce high-amplitude, wide-angle reflections asymptotically approaching the velocity of the crust above the discontinuity (4 km/s). Because we see no such phase, we suggest that a more gradual velocity transition is likely. (2) The complete lack of converted shear energy on any of the ten profiles implies that there are no sharp compressional and shear velocity contrasts at or just beneath the seafloor. The high Poisson's ratio and low compressional velocity which probably characterizes the young oceanic basement [Spudich and Orcutt, 1980b] reduces the

efficiency of p to S conversion (lower case letter represents travel of that phase in the sediment or water layer) at the seafloor to a sufficiently low level to explain the lack of identifiable shear wave phases [White and Stephen, 1980]. It is interesting to note, however, that the data recorded by a three-component ocean-bottom seismometer (OBS 1 on Figure 2) on line 4 contained some converted shear energy on its radial component [Chapter 2]. It is possible that conversion from p to S at the sediment/basement contact is energetic enough to be recorded as arriving shear energy on an OBS horizontal geophone, but that the conversion from p to S to p necessary to be recorded on an OBH is not energetic enough in the young crust along line 4.

Six of the ten refraction sections exhibit a region of high amplitude first arrivals in the 5 to 10 km range, a feature that has been attributed to the transition from layer 2 to layer 3 [Spudich and Orcutt, 1980a,b]. A visual comparison of the seismic profiles suggests that this amplitude feature and thus the layer 2 - layer 3 transition zone (which was not defined at all by the travel-time interpretation) are perhaps the most variable features along the 200 km strike line. The signature of the transition zone is characterized on the refraction sections by low amplitudes at close range which build to a maximum with increasing range then decrease again, all within a range window 3 to 5 km in width.

To better examine the layer 2 - layer 3 transition we

compute the power in the first 0.2 s of each seismogram on all ten seismic sections. To achieve maximum resolution, both 2 and 11 kg shots were used. The seismograms are scaled by charge weight and also linearly by range. The power-versus-range plots in Figure 10 show that the amplitude patterns fall into four groups. The plots in the first group represent lines displaying no peak within 15 km range. The data recorded by OBH 1 from the south and shown in Figure 4a typify this group of lines. The remaining three groups exhibit distinct peaks at ranges of 6, 8, and 10 km, respectively, and each group is represented by power curves from two refraction lines. The maximum power within the peak is a factor of 1.5 to 3.0 greater than the power at greater or lesser ranges. The refraction section shown in Figure 4b (OBH 4 south), for example, has a peak in its power curve at 10 km.

We proceed to model the layer 2 - layer 3 transition by perturbing the basic upper-crustal structure obtained in the travel-time study and using WKBJ techniques [Chapman, 1978] to compute synthetic seismograms of arrivals out to 20 km range. To compare the synthetic seismograms with the data, we next calculate the power in the first 0.2 s of each synthetic seismogram in a manner identical to that previously applied to the refraction data. A most important constraint on the models is that the travel-time curves computed from the synthetic seismograms fit within the maximum confidence bounds assigned to the travel-time data (Figure 6).

The lack of an amplitude build-up in the first group of plots in Figure 10 can be attributed to an upper-crustal velocity gradient that continually decreases with depth. The structure derived from the travel-time analysis, for example, would adequately explain the absence of an amplitude peak in the 5 to 10 km range. The peak observed in each of the other three groups of power versus range curves require that a steep velocity gradient (characterizing the layer 2-layer 3 transition zone) be sandwiched above and below by zones of near-zero velocity gradient. The velocity contrast across the transition zone must be large (> 1 km/s) to match the observed magnitude of the peaks. The range at which the maximum power is reached appears strongly related to the depth at the base of the transition zone.

Layer 2 may be regarded as being made up of three units; mean velocities through each of these units show a similarity to velocities in the three layer solution for layer 2 proposed by Houtz and Ewing [1976]. We use their terminology, layer 2A, 2B, and 2C, to represent the steep gradient layer at the top of the crust, the underlying region of near-zero velocity gradient and the layer 2 - layer 3 transition zone, respectively. According to this nomenclature the four refraction profiles exhibiting no amplitude peak at close range sample crust with no resolvable layer 2B.

The constraints imposed by the travel times and amplitudes greatly narrow the bounds of possible structures which can be responsible for the amplitude patterns observed at ranges from

5 to 10 km. There are, however, many structures within these bounds that can satisfy the observations. We attempt to develop sets of simple, reasonable models in which there is a logical progression in the shapes of the travel-time curves corresponding to the observed amplitude peaks at ranges of 6, 8, and 10 km. A structure in which crustal velocity gradients decrease with depth and thus produces no layer 2 - layer 3 amplitude event becomes a member of this set at either end of the structural progression. The two sets of models shown in Figure 11 illustrate the range of structures that can fit both the travel-time and spectral-power constraints. In both sets the velocity structure of the upper 0.7 km of crust (layer 2) is similar to that derived from the travel-time analysis. Models in which the layer 2C thickness and gradient vary as well as those in which both quantities remain constant can generate amplitude patterns that are in good agreement with the observations. In the model sets shown in Figure 11, and in all other sets generated to match the data, the depths to the base of layer 2C (and thus the total thickness of layer 2) necessary to create amplitude peaks at 6, 8, and 10 km range are within 0.1 km of 1.60, 1.95, and 2.20 km beneath the seafloor, respectively. The thicknesses of layer 2 along lines without an amplitude peak are more difficult to infer because of the subtlety of the transition from layer 2 to layer 3 velocities required by this amplitude pattern.

The synthetic travel-time curves and power curves for the model set shown in Figure 11a are displayed in Figure 12 and for

the set in Figure 11b in Figure 13. In the first model set, a layer 2C with a thickness of 0.6 km and a gradient of 2.6 s^{-1} is placed increasingly deeper to produce amplitude events at greater range. Transition zones thinner than a seismic wavelength (0.5 to 0.7 km) produce late-arriving phases reflecting from the steep transition zone gradient which arrive with amplitudes that are strong relative to the first-arriving energy. The amplitudes of these reflections become especially pronounced when the transition area is located deeper than 2 km. Though some refraction profiles (see Figure 4) exhibit energy that might be reflections from layer 2C, the phenomenon is not pronounced or consistent on a shot-to-shot basis throughout the refraction sections collected along line 4. We therefore suggest that the thickness of the layer 2 - layer 3 transition regions sampled under line 4 cannot be much thinner than those shown in Figure 11a.

In the model set of Figure 11b, the thickness and gradient of layer 2C vary. A thicker layer 2 - layer 3 transition zone corresponds to a thinner layer of constant velocity material above the zone. A difference between this model set and that shown in Figure 11a is that there is a relative paucity of constant-velocity crust above layer 2C which allows more energy to arrive at ranges between the OBH and the amplitude event. The product of such a structure is a peak that is generally less sharp than those computed from the set of Figure 11a. Models with a layer 2B thickness of less than 0.2 km (equal to the thinnest layer 2B shown in Figure 11b) produce only small

peaks in the power-versus-range functions that are within the noise of the data in Figure 10.

Another possible set of models might be one in which the different amplitude patterns in the 5 to 10 km range are due to variability in the velocity of the upper 0.7 km of crust, and not in layers 2B and 2C. However, the first order effect of simultaneously holding the structure of the layer 2 - layer 3 transition zone constant and varying the shallow crustal velocities would be to increase or decrease all of the resulting arrival times, including those of energy refracting through the transition region. With models such as these, it would be difficult to match variations in the range of the observed amplitude patterns. Also, more extreme perturbations to the structure of the lower crust would be required to keep the travel-time curves computed from the models within the bounds of uncertainty assigned to the observed travel times. We therefore have considered only sets of structures in which the uppermost 0.6 to 0.7 km of crust is identical to that derived from the travel-time inversion.

As the summary of the two model sets in Table 2 indicates, it is impossible with these data to correlate the individual thicknesses of layer 2A, 2B or 2C with the location of layer 2 - layer 3 amplitude peaks. In the first model set where layer 2C thickness remains constant, the range at which the peak amplitude occurs increases as the thickness of layer 2B increases. In the second model set, as the range at which peak amplitudes occur decreases, the layer

2B thickness increases and the transition layer 2C thickness decreases. Given that our assumptions about the velocities in the upper 0.7 km of layer 2 are correct the amplitudes can, however, provide at least two constraints on the thicknesses of seismic units in the upper crust: (1) an absence of the layer 2 - layer 3 amplitude peak implies a layer 2 in which there is no resolvable thickness of a constant velocity layer (layer 2B) and, (2) if the amplitude peak is present, the event moves farther out in range as the total thickness of layer 2 increases.

Just as important as the above two constraints on the structures associated with particular amplitude patterns is the evidence from the 5 to 10 km amplitude patterns that the 0.5 m.y. old crust under line 4 is definably heterogeneous on a seismic scale. The total thickness of layer 2 under the rise-flanking profiles which display a layer 2 - layer 3 amplitude peak vary by 0.5 to 0.8 km over lateral distances of tens of kilometers, depending on the assumed model. The variability of layer 2 thicknesses along line 4 could conceivably be greater than this range if it were possible to better define the structure under the lines exhibiting no amplitude peak. Thus some variability in the process of crustal accretion is suggested by the observed variability of amplitude patterns and the implied heterogeneity in structure of the upper crust.

Because of the magnitude of the uncertainties assigned to the travel-time data and the consistency of those data

within the error bounds, we concluded previously that a travel-time analysis alone could resolve no heterogeneity within the crust under the rise-flanking line. A re-examination of the travel-time curves in light of the amplitude study confirms our previous assertions. Despite the fact that the power-versus-range curves within each of the four groups in Figure 8 are very similar in shape, the travel-time curves (Figure 4) corresponding to the members of each group do not always bear a close resemblance to each other. Furthermore, the "kink" in the synthetic travel-time curves (Figures 10 and 11) representative of the layer 2 - layer 3 transition is difficult to resolve in any of the ten refraction sections of Figure 5. We can conclude from these observations that the error bounds placed on the first-arrival travel times are realistic and that, indeed, travel-time analysis resolves structure on a scale larger than that observed when amplitude information is included as well.

DISCUSSION

It is, of course, impossible to define unambiguously the geologic structure of a region of oceanic crust from the compressional velocity structure alone. As a result, many authors [e.g., Fox et al., 1973; Christensen and Salisbury, 1975; Spudich and Orcutt, 1980b] have further constrained oceanic crustal geology by comparing the velocity structures inferred from travel-time and amplitude analyses with the structure of ophiolite complexes and the velocities measured in laboratory samples from ophiolites, dredge hauls, and

drill cores. Though there is no universal agreement as to whether ophiolite complexes were formed at spreading centers representative of those currently active, drilling at Deep Sea Drilling Project (DSDP) hole 504B has confirmed that the upper units of a complete ophiolite section (sediments, metamorphosed extrusives, metamorphosed sheeted dikes) are represented in the upper 1350 m of 6.2 m.y. old Pacific crust [Anderson et al., 1982].

We consider the possible effects of lithology, metamorphism, and porosity on the velocity structure of the crust along line 4 based on the results at hole 504B and studies of ophiolite complexes. We are especially interested in the geologic analogs of seismic layers 2A, 2B, 2C, and 3 and whether or not the magnitude and scale of variability seen in the seismic models of this study agree with field observations of structural variability seen in the drilling results and ophiolite complexes.

DSDP hole 504B is thus far the only site where we can look in situ at the structure down to the sheeted dike units and compare them to what is seen on a seismic scale in marine refraction experiments. Of the 1075 m of oceanic basement drilled at 504B, the upper 575 m is composed of pillow lavas, breccias, and small flows, the lower 290 m is primarily in the form of dikes, and the intervening 210 m is a transition zone between the two units [Anderson et al., 1982]. Sonic logs from the hole provide velocity measurements of altered extrusive and dike units on a scale of about 1 m. Figure 14

shows the compressional velocity sonic log obtained from hole 504B compared to the observed lithostratigraphy. Though fluctuations in the data due to small-scale heterogeneities, irregular crack geometry, and disturbance of pore pressure and fluid content related to drilling complicate the measured velocity structure, several trends can be seen in the data. Steep velocity gradients in the upper crust take the average compressional velocity from about 4 km/s at the top of the basement to about 5 km/s at 200 m into the basement. From this depth to the top of the extrusive dike transition zone, the mean velocity fluctuates around 5 km/s. Within the extrusive-dike transition region, velocities rise to values over 6.0 km/s and fluctuations in the sonic log decrease, thus suggesting that the dikes are not only of higher velocity material but also are more homogeneous on the scale sampled by the logging tool. Compressional velocities within the pure dikes continue to increase with depth to the base of the hole where a velocity of about 6.3 km/s is reached. The pattern of velocity gradients indicated by the sonic log from the top to bottom--steep, low, intermediate--is exactly that required by the amplitude modelling on line 4.

The comparison of our inferred seismic velocities to the velocities of altered and unaltered ophiolite samples measured on a laboratory scale provides another means by which we can constrain the geology of the crust along line 4. Because some alteration observed in ophiolite complexes is probably related to post-obduction processes, one must be

careful when correlating sample velocities with velocities measured in situ by refraction and logging techniques.

Salisbury and Christensen [1978] measured samples from the Bay of Islands ophiolite at pressures and temperatures thought appropriate to the environment in which they once existed. Velocities of metabasalts (extrusives) ranged from 5.6 to 6.5 km/s, metadolerites (sheeted dikes) and metagabbros ranged between 6.6 and 7.0 km/s, and pyroxene gabbros gave velocities between 6.9 and 7.3 km/s. Additionally, the crustal gradient structure suggested by the measured ophiolite samples (shown in Figure 5, Salisbury and Christensen [1978]) is similar to that determined in our study and seen in the hole 504B sonic log [Anderson et al., 1982]. The two steepest velocity gradients in the upper crustal units of the Bay of Islands complex occur at the top of the extrusives and between the base of the extrusives and the upper section of sheeted dikes. The velocity remains fairly constant as a function of depth throughout the sheeted dike samples before a third steep velocity gradient occurs starting in the metagabbroic rocks and continuing into the top of the unaltered pyroxene gabbros.

The pattern of gradients in bulk porosity versus depth calculated from apparent resistivity measurements in DSDP hole 504B by Becker et al. [1982] (Figure 14) also may be related to the velocity-depth gradients seen in the hole 504B sonic log the Bay of Island ophiolite, and the velocity structure determined for line 4 in this chapter. Porosity

decreases rapidly in the upper 300 m of extrusives from about 13% to 8%. The porosity in the next 300 m of extrusives remains nearly constant. Within the 210 m geologic transition zone, porosity drops to less than 3% and continues to drop slowly to 1% in the lower 200 m of dikes sampled by the resistivity log.

The velocity structure of oceanic crust is probably strongly related to the porosity of its different geologic units. Spudich and Orcutt [1980a] examined the possible effects that the volume and shape of cracks and fissures have on the velocities of extrusive basalts. They concluded that though laboratory samples of Mid-Atlantic Ridge basalts with 8% porosity have an average velocity of 5.9 km/s [Hyndman and Drury, 1976], the velocity of the same material with 20% porosity could fall anywhere between 2.0 and 5.1 km/s depending on the assumed crack geometry. The porosity of near-surface extrusives may actually be greater than 20% in young oceanic crust [Hyndman and Drury, 1976]. Velocities increase rapidly within young crust due to a decrease in porosity as a function of depth [Spudich and Orcutt, 1980a]. This effect can be seen directly in the sonic log data and apparent porosity computations from hole 504B [Anderson et al., 1982; Becker et al., 1982]. It can also be inferred from our data where layer 2 velocities increase from 2.1 km/s at the surface to 5.6 km/s at a depth of about 0.7 km. The difference between the travel-time curves of the 140 m.y. old Atlantic crust [Purdy, 1983] and the 0.5 m.y. old Pacific

crust both compared in Figure 8 suggest that the steep velocity gradients found at the top of a young oceanic crustal section are made more gradual by the closing of cracks and pores due to chemical alteration and precipitation in the upper crust [e.g. Houtz and Ewing, 1976; Spudich and Orcutt, 1980a,b] as a section of oceanic lithosphere ages.

The decrease in porosity and accompanying increase in velocity as functions of depth in the extrusives may be primarily related to three different processes: 1) An increase in the confining pressure of about 0.2 kb in the upper 0.8 km of crust acts to close an increasing percentage of cracks and pores with increasing depth and thus contributes to the steep velocity gradients observed. The decrease in the number of large pillows and flows and the increase in the amount of brecciation observed in borehole televiewer records down hole 504B may represent the compressive effects of increasing overburden in the extrusives [Anderson et al., 1982]. In an environment controlled partially by a relationship between porosity and confining pressure, there may be some critical depth at which the increasing overburden no longer has a significant effect on the porosity and velocity of single rock unit. Because the existence of any significant pore pressure will tend to hold pore spaces open, the pressure at this critical depth is probably less than the pressure (1kb or about 2.8 km below the seafloor) predicted by Walsh [1965] at which most elongate cracks are thought to close in dry rock. Thus if the extrusives are

thicker than the critical depth, the velocity in the rock below this critical depth will no longer be controlled by changing crack and pore volume related to increasing overburden. 2) The results from DSDP hole 504B suggest a strong correlation between decreasing porosity and an increasing amount of hydrothermal mineralization with depth in the extrusives [Anderson et al., 1982; Becker et al., 1982]. Toward the base of the extrusives (from 200 to 500 m) where porosity gradients are small (Figure 14), a large percentage of cracks and pores are filled by alteration products. This interval is also a region where velocities in the extrusives reach a maximum. 3) A final contributor to the decrease in porosity with depth in the extrusives may be related to draining of a portion of fresh lava downward into previously extruded units near the rise axes. The product of this "flow back" might be an extrusive section in which older pillows and flows are increasingly saturated as a function of depth with material that has been drained from the surface. The CYAMEX Scientific Team [1981], in dives at 21°N on the East Pacific Rise, noted the presence of drained lava lakes in that region. This suggests that at least some fraction of erupted lava flows down into the subsurface to both decrease the bulk porosity and increase the velocity in the underlying units.

Based on the above discussion and the results of the amplitude and travel-time studies of line 4 refraction data, we suggest a geologic model for the crust under the rise-

flanking line similar to that shown in Figure 15. This model is not based solely on determinations of compressional wave velocities but most importantly on comparisons of the velocity gradients inferred from the line 4 data with the velocity and porosity gradients measured in DSDP hole 504B and in ophiolite samples.

In this model for the upper oceanic crust, layer 2 is composed mainly of extrusives and sheeted dikes which act and interact on a seismic scale so as to produce two regions of high velocity gradients. The uppermost region of high velocity gradients (4.4 s^{-1}) from the surface down to a depth of about 0.7 km is layer 2A. It is characterized by extruded pillow lavas and rubble in which the velocity increases as a function of decreasing porosity. Below this depth, the amplitude patterns for shots in the 5 to 10 km range on six of the ten refraction lines require a region, layer 2B, in which the velocity is nearly constant. The low velocity gradient in layer 2B suggests that the porosity in this layer has approached a constant and thus has ceased to be a factor controlling velocity changes with depth in the extrusives. The near-constant porosity in layer 2B may represent a minimum porosity for extrusives in an oceanic environment. Layer 2B then defines a region where confining pressure, hydrothermal mineralization, and extrusive flow back have closed a large fraction of the non-grain boundary porosity. This minimum is probably close to the porosities measured by Becker et al. [1982] in the near-constant porosity section of extrusives in

hole 504B (~ 8%) and those measured in laboratory samples of Mid-Atlantic Ridge basalts by Hyndman and Drury [1976] (~ 8%). Because velocities in the zero-gradient layer of crust in our models (5.3 to 5.6 km/s) are constrained to be close to those of low porosity ophiolitic [Salisbury and Christensen, 1978] and oceanic [Hyndman and Drury, 1976] basalts measured in the laboratory and are also within the range measured in 8% porosity basalts in hole 504B, it is likely that layer 2A and 2B are primarily extrusive in origin and basaltic in composition. The remaining four lines where no amplitude peaks are observed at short range sample thin extrusive sections with no resolvable layer 2B. All three primary features which control porosity in the extrusives--overburden, hydrothermal mineralization, and lava flow back--all act to decrease porosity with depth. Thus if the extrusives are not thick enough for the porosity to reach some minimum it is probable that the extrusives would never develop a region of near constant velocity and porosity within its base. One might then expect that rays arriving from shots at 5 to 10 km range on lines exhibiting no amplitude peak (OBH 4 shot from the north, 3 south, 1 south, 8 north) propagate through crust that is topographically elevated since such areas may have accumulated the least amount of extruded volcanics [e.g. CYAMEX Scientific Team, 1981]. Examination of Figure 3 indicates that arrivals from shots at short range on the aforementioned lines propagate beneath topographic highs and that they may, in fact, sample thin volcanic layers emplaced

on previously elevated regions of newly forming crust.

Progressing downward through the crust, the next seismological unit is layer 2C (layer 2-layer 3 transition zone.) We suggest that this region is a contact between the extrusives and the underlying sheeted dikes, which is complex on the scale viewed in drilling cores and logs and in ophiolites, gradational on a seismic scale, and variable on all scales. At the Bay of Islands ophiolite complex, where there is 100-km wide segment of what is thought to be oceanic lithosphere, the extrusive-dike contact intertwines on predominantly two scales [e.g. Casey et al., 1981; Rosencrantz et al., 1982]. Extrusives grade into 1 m thick dikes by an increase in the percentage of dikes with depth. The scale across which the transition takes place varies from a few hundred meters to several meters over lateral distances of several kilometers. On a larger scale, the extrusive-dike contact itself undulates vertically with amplitudes and wavelengths of up to several hundred meters. When viewed on the scale sampled by seismic waves, this contact could represent the layer 2 - layer 3 transition zone, similar in thickness and variability to that seen along line 4, formed by both the interfingering of extrusives and dikes and the undulating shape of the contact between the two lithologic units. The velocity within layer 2C increases as a function of depth due to a combination of five factors: 1) a decreasing percentage of basaltic extrusives; 2) increasing percentage of diabase dikes (laboratory v_p = 6.6 to 6.9 km/s

[Salisbury and Christensen, 1978]; 3) the presence of some higher velocity metabasalts (V_p up to 6.5 km/s 4) a decreasing volume of brecciated dikes ($V_p \sim 6.3$ km/s); and 5) a general decrease in porosity (Figure 14) resulting from both an infilling of cracks and pores by hydrothermal mineralization and the transition from extrusive volcanics ($\sim 8\%$ porosity) to sheeted dikes with a smaller volume of cracks and pores ($\sim 1\%$). The substantial drop in fracture intensity and degree of alternation observed in hole 504B across the extrusive-dike transition suggested to Becker et al. [1982] that the dikes may possess inherently low porosity and that the efficiency of hydrothermal circulation in the dike units was probably insignificant in the region of the drill hole.

Geologic evidence from ophiolite sections [Geotimes, 1972] and drilling results [Anderson et al., 1982] suggests that a region of horizontally continuous sheeted dikes should be present beneath the complex extrusive-dike transition zone. If, then, a region of sheeted dikes, laterally continuous on a seismic scale, exists in the crust along line 4, it is probably located in the top portion of layer 3 beneath the seismic transition zone (layer 2C). The velocities determined in this study for the upper 1 km of layer 3 were found to be between 6.8 and 6.9 km/s and thus are comparable to velocities of sheeted dike samples measured in the laboratory [Salisbury and Christensen, 1978]. Also, the similarity between the gentle vertical velocity gradients

measured in uppermost layer 3, to those inferred from Salisbury and Christensen's [1978] measurements of ophiolite samples further strengthens the argument that the top of layer 3 is primarily composed of dike material.

At the Bay of Islands ophiolite complex [Casey et al., 1981] the transition from dikes to isotropic gabbro is gradational and varies in thickness by several hundred meters over lateral distances of several kilometers. The velocity contrast across the transition zone is probably small (< 0.3 km/s, Salisbury and Christensen [1978]) and, therefore would be expected to produce only a minor amplitude peak even if the geometry of the zone were similar to that of the extrusive-dike transition. A small peak in more than half of the power-versus-range curves in Figure 10 between the ranges of 14 and 21 km may be the signature of a velocity transition zone within layer 3. This transition region might represent a change from sheeted dikes to massive gabbros, but could instead be related to a change in metamorphic facies. Because these amplitude features are small and are usually defined on only one or two seismograms, we have not formally modelled them. However, we can conclude from the variation in the range of the small amplitude peaks that whatever process is responsible for the structure within layer 3 is variable over the length of line 4.

SUMMARY AND CONCLUSIONS

We have reported an analysis of refraction data recorded by five ocean-bottom hydrophones on a line shot parallel to

and 25 km east of the East Pacific Rise between 11° and 13°N latitude. Because of the uniformity of instrumentation and the large quantity of clearly recorded data (over 500 seismograms), this experiment was ideally suited for the study of the structure and structural variability along strike on 0.5 m.y. old Pacific crust. The nature of seismic refraction data allow us to view the crust on two different spatial scales. Both scales are dependent on the experimental design and uncertainties in the information used to resolve structure within the crust.

On the basis of only first-arrival times, the crust along the 200 km long rise-flanking line is homogeneous. An inversion of 479 travel times yields a well-constrained mean crustal structure along line 4 characterized by velocity gradients that are steep (4.4 s^{-1}) in uppermost layer 2 and that decrease continuously with increasing depth. The vertical velocity gradient was determined to be near zero in layer 3. By viewing this young crust at the scale defined by the first arrival time data and by comparing this study to similar studies carried out in other regions, we can examine gross structural variability that may be related to fundamental large-scale processes such as spreading rate, tectonic province, or crustal age. For example, a comparison of our results with those derived from an identical analysis of 140 m.y. old Atlantic lithosphere [Purdy, 1983] indicates that the major difference between the two segments of the uppermost 3 km of crust is that the velocity gradient in the

upper 0.5 km of old Atlantic basement is only 1.1 s^{-1} .

A study of the distribution of amplitude with range permits the determination of structure in significantly greater detail and leads to a number of important conclusions:

1) The use of amplitude information allows the resolution of structural details within the young crust beneath line 4 that were not observed in the travel-time data alone.

2) Six of the ten refraction profiles contain a prominent and consistent pattern of high-amplitude first arrivals in the range 5 to 10 km. Systematic changes in the location of the amplitude patterns in this range window can be related to lateral changes in the structure of layer 2.

3) The simplest explanation of the amplitude peak in this range window is a layer 2 consisting of three components: i) an uppermost layer (2A) less than 1 km in thickness that is characterized by the highest ($\sim 4 \text{ s}^{-1}$) vertical velocity gradients, underlain by ii) about 0.5 km of very low gradient or even constant velocity material that composes layer 2B, and finally iii) a layer 2C that has an intermediate gradient and is the transition from layer 2 velocities to the low gradients ($< 0.1 \text{ s}^{-1}$) and velocities near 6.8 km/s that characterize oceanic layer 3.

4) Forward modelling of this amplitude peak using synthetic seismograms suggests that the thickness of layer 2 varies by about 0.7 km over lateral distances of tens of kilometers along the 200 km line. The only systematic pattern to these changes is a correlation between small

topographic highs and thin layer 2B.

5) It is well known that compressional wave velocities alone cannot be unambiguously related to rock type or structural fabric. However, an important additional constraint provided by the data presented here is the interrelationships of the velocity gradients that are well determined by the amplitude distributions.

6) A comparison of the velocities, velocity and porosity gradients, and variability inferred from these data with those seen in DSDP hole 504B [Anderson et al., 1982] and in ophiolite complexes [e.g., Casey et al., 1981] suggests that layers 2A and 2B are extrusive in origin, the material at the top of layer 3 consists of horizontally continuous sheeted dikes, and layer 2C is a region of interfingering and undulations of dikes and extrusives on vertical and lateral scales of meters to hundreds of meters. Another amplitude peak, which varies in range from 14 to 21 km, may signal a transition from sheeted dikes to the underlying isotropic gabbros.

7) The presence and variability of these two different amplitude peaks provide direct evidence that the along-strike structure of oceanic crust contains vertical changes in structure of several hundred meters that are definable by conventional seismic refraction experiments when interpreted in the careful manner described here. The magnitude and consistency of the amplitude peak between 5 and 10 km range and the clear relationship of that peak to velocity gradients

within the upper crust suggest that the layer 2 - layer 3 transition may serve as a useful and mappable characteristic of oceanic crust that can be sampled by drilling and related to seismic results. Further studies are needed to better resolve the lateral structural changes responsible for the heterogeneity seen along the line examined in this study so that any systematic relationship between these changes and along-strike variations in the accretion processes can be defined and understood.

REFERENCES

- Anderson, R.N., J. Honnorez, K. Becker, A.C. Adamson, J.C. Alt, R. Emmermann, P.D. Kempton, H. Kinoshita, C. Laverne, M.J. Mottl, and R.L. Newmark, DSDP hole 504B, the first reference section over 1 km through layer 2 of the oceanic crust, Nature, 300, 589-594, 1982.
- Aumento, F., W.G. Melson et al., Initial Reports of the Deep Sea Drilling Project, Vol. 37, Washington, D.C., U.S. Government Printing Office, 1977.
- Ballard, R.D., W.B. Bryan, J.R. Heirtzler, G. Keller, J.G. Moore and Tj Van Andel, Manned submersible operations in the FAMOUS area: Mid-Atlantic Ridge, Science, 190, 103-108, 1975.
- Becker, K., R.P. Von Herzen, T.J.G. Francis, R.N. Anderson, J. Honnorez, A.C. Adamson, J.C. Alt, R. Emmermann, P.D. Kempton, H. Kinoshita, C. Laverne, M.J. Mottl, and R.L. Newmark, In situ electric resistivity and bulk porosity of the oceanic crust, Costa Rica Rift, Nature, 300, 594-598, 1982.
- Bracewell, R., The Fourier Transform and Its Applications, p. 160, New York, McGraw-Hill, 1965.
- Bratt, S.R., and S.C. Solomon, Compressional and shear wave structure of East Pacific Rise at 11°20'N: constraints from three-component ocean-bottom seismometer data, J. Geophys. Res., in press, 1984.
- Cann, J.R., A model for oceanic crustal structure developed, Geophys. J. R. Astron. Soc., 39, 169-187, 1974.

- Casey, J.F., J.F. Dewey, P.J. Fox, J.A. Karson and E. Rosencrantz, Heterogeneous nature of oceanic crust and upper mantle: A perspective from the Bay of Islands ophiolite complex, in The Sea, vol. 7, edited by C. Emiliani, pp. 305-338, New York, Wiley-Interscience, New York, 1981.
- Chapman, C.H., A new method for computing synthetic seismograms, Geophys. J. Roy. Astron. Soc., 54, 481-518, 1978.
- Christensen, N.I., and M.H. Salisbury, Structure and constitution of the lower oceanic crust, Rev. Geophys. Space Phys., 13, 57-86, 1975.
- CYAMEX Scientific Team, First Manned submersible dives on the East Pacific Rise at 21°N (Project RITA): General results, Mar. Geophys. Res., 4, 345-379, 1981.
- Ewing, J.R., and R.P. Meyer, Rivera Ocean Seismic Experiment (ROSE) overview, J. Geophys. Res., 87, 8345-8357, 1982.
- Ewing, J. and G.M. Purdy, Upper crustal velocity structure in the ROSE area of the East Pacific Rise, J. Geophys. Res., 87, 8397-8402, 1982.
- Fox, P.J., E. Schreiber and J.J. Peterson, The geology of the oceanic crust: Compressional wave velocities of oceanic rocks, J. Geophys. Res., 78, 5155-5179, 1973.
- Fuchs, K., and G. Muller, Computation of synthetic seismograms with the reflectivity method and comparison with observations, Geophys. J. Roy. Astron. Soc., 23, 417-433, 1971.

- Geotimes, Penrose field conference on ophiolites, 17, 24-25, 1972.
- Gettrust, J.F., K. Furukawa, and W.B. Kempner, Variation in young oceanic crust and upper mantle structure, J. Geophys. Res., 87, 8435-8445, 1982.
- Houtz, R.E., and J. Ewing, Upper crustal structure as a function of plate age, J. Geophys. Res., 81, 2490-2498, 1976.
- Hyndman, R.D., and M.J. Drury, The physical properties of oceanic basement rocks from deep drilling on the Mid-Atlantic Ridge, J. Geophys. Res., 81, 4042-4060, 1976.
- Hughes, D.S., and C. Maurette, Variation of elastic wave velocities in basic igneous rocks with pressure and temperature, Geophysics, 22, 23-31, 1957.
- Kennett, B.L.N. and J.A. Orcutt, A comparison of travel time inversions - Marine refraction profiles, J. Geophys. Res., 81, 4061-4070, 1976.
- Klitgord, K., and J. Mammerrickx, Northern East Pacific Rise: Magnetic anomaly and bathymetric framework, J. Geophys. Res., 87, 6725-6750, 1982.
- Koelsch, D.E., and G.M. Purdy, An ocean-bottom hydrophone instrument for seismic refraction experiments in the deep ocean, Mar. Geophys. Res., 4, 115-125, 1979.
- Lewis, B.T.R., and J.D. Garmany, Constraints on the structure of the East Pacific Rise from seismic refraction data, J. Geophys. Res., 87, 8417-8425, 1982.

- Mammerickx, J., and K.D. Klitgord, Northern East Pacific Rise: Evolution from 2.5 m.y.B.P. to the present, J. Geophys. Res., 87, 6751-6766, 1982.
- Macdonald, K.C., and P.J. Fox, Overlapping spreading centers: New accretion geometry on the East Pacific Rise, Nature, 301, 55-58, 1983.
- Mosteller, F., and R.E.K. Rourke, Sturdy Statistics, p. 294, Addison-Wesley, Reading, MA, 1973.
- Purdy, G.M., The variability in seismic structure of Layer 2 near the East Pacific Rise at 12°N, J. Geophys. Res., 87, 8403-8416, 1982a.
- Purdy, G.M., The correction for the travel time effects of seafloor topography in the interpretation of marine seismic data, J. Geophys. Res., 87, 8289-8396, 1982b.
- Purdy, G.M., The seismic structure of 140 m.y. old crust in the Western Central Atlantic Ocean, Geophys. J. Roy. Astron. Soc., 72, 115-138, 1983.
- Raith, R.W., The crustal rocks, in M.N. Hill, ed., The Sea, vol. 3, New York, Interscience, 85-102, 1963.
- Rosencrantz, E., Formation of uppermost oceanic crust, Tectonophysics, 1, 471-494, 1982.
- Salisbury, M.H., and N.I. Christensen, The seismic velocity of a traverse through the Bay of Islands Ophiolite Complex, Newfoundland, an exposure of oceanic crust and upper mantle, J. Geophys. Res., 83, 805-817, 1978.

- Spudich, P., and J. Orcutt, A new look at seismic velocity structure of the oceanic crust, Rev. Geophys. Space Phys., 18, 627-645, 1980a.
- Spudich, P., and J. Orcutt, Petrology and porosity of an oceanic crustal site: Results from waveform modeling of seismic refraction data, J. Geophys. Res., 85, 1409-1433, 1980b.
- Stephen, R.A., and A.J. Harding, Travel-time analysis of borehole seismic data, J. Geophys. Res., 88, 8289-8298, 1983.
- Walsh, J.B., The effect of cracks on the compressibility of rock, J. Geophys. Res., 70, 381-389, 1965.
- White, R.S., and R.A. Stephen, Compressional to shear wave conversion in oceanic crust, Geophys. J. R. Astron. Soc., 63, 547-565, 1980.
- Whitmarsh, R.B., Seismic refraction studies of the upper igneous crust in the North Atlantic and porosity estimates for layer 2, Earth Planet. Sci. Lett., 37, 451-464, 1978.

FIGURE CAPTIONS

- Figure 1. Geographic location of Project ROSE. The numbered lines at the bottom of the figure are the seismic refraction lines from the first phase. The box in the upper left encloses the microearthquake study region.
- Figure 2. Locations of refraction lines and instruments discussed in this chapter in relation to the local bathymetry [after Klitgord and Mammerickx, 1982]. In addition to the locations of 5 Woods Hole (WHOI) instruments, locations for instruments from the Massachusetts Institute of Technology (MIT) and the University of Washington (UW) are shown. The interval between the solid contours is 500 m.
- Figure 3. Bathymetric profile along refraction line 4 as a function of range from the OBH 1. The locations of all five OBHs used in this study are also shown.
- Figure 4. Two sample record sections from the Woods Hole OBHs on line 4. All travel times have been reduced using a velocity of 8 km/s. The amplitudes have been scaled linearly with range and adjusted for charge weight. Each seismogram stops at the arrival time of the water wave. (a) OBH 1 south. (b) OBH 4 south.
- Figure 5. Travel-time curves from the ten refraction sections used in this study. The arrival times

have been reduced by 8 km/s. Note the absence of phase velocities near 8 km/s. Arrivals picked at ranges greater than 35 km may be supercritical Moho refractions.

Figure 6. Merged set of 479 travel times versus range for refracted compressional waves. The arrival times have been reduced by 8 km/s. Travel times in the upper plot show the scatter in the data before application of a water-path correction [Purdy, 1982a]. Travel times in the lower plot have been water-path corrected. The spline fits the 479 data points with a root mean square deviation of 0.05s. The error bars mark the upper and lower bounds on the uncertainties in the travel-time data.

Figure 7. Compressional velocity structure above a depth of 3.8 km was derived from an inversion of the spline fit to the merged travel times in Figure 6. Below this depth, the structure was derived using amplitude information from ranges between 20 and 60 km.

Figure 8. Comparison of the P-wave travel-time data collected by 5 OBHs on line 4 with the merged set of travel times collected on 140 m.y. old Atlantic crust and examined by Purdy [1983]. Both sets of data were identically processed. Travel time through the thick sedimentary layer in the

140 m.y. old crust has been removed. The difference between the sets of travel times from the two studies is probably due to lower velocity in the young upper crust of this experiment.

Figure 9. WKBJ synthetic seismic section calculated for the crustal structure (shown in Figure 8) for line 4. The source time function is an exponentially damped cosine function, $\cos(2\pi ft)\exp[-(2\pi ft/\gamma)^2]$, where $f = 8$ Hz and $\gamma = 5$. The scaling of seismograms with range is the same as in Figure 4. The synthetic section is compared to data collected by OBS 7 from shots to the north.

Figure 10. Power in the first 0.2s after the first refracted arrivals as a function of range from the OBH. The plots are grouped by curve shape: (a) lines 8N (OBH 8 shot from the north), 1S, 3N, 4N have no power peak; (b) lines 8S and 3S have a power peak at 6 km range; (c) lines 7N and 1N have a power peak at 8 km range; (d) lines 7S and 4S have a power peak at 10 km range. Note also the secondary peak in some of the curves between the ranges of 14 and 21 km.

Figure 11. Two of the sets of simple crustal structure models found to fit the travel-time and amplitude data at short range. (a) Model set in which the layer 2 - layer 3 transition zone thickness and gradient remain constant and the depth to the base of the

transition zone varies. (b). Model set in which transition zone thickness, gradient and depth vary. In both sets, the greater the depth to the base of the transition zone, the greater the range at which the amplitude peak is observed. Each structure is designated by a letter corresponding to the model set (a or b) and a number corresponding to the range (6, 8, or 10 km) at which the amplitude peak resulting from this structure is observed.

Figure 12. Power and travel-time curves generated from the models in the set shown in Figure 11a. Dashed curve outlines the cloud formed by the merged first-arrival times from line 4. Labelling scheme is the same as that in Figure 11.

Figure 13. Power and travel-time curves generated from the models in the set shown in Figure 11b. Dashed curve outlines the cloud formed by the merged first-arrival times from line 4. Labelling scheme is the same as that in Figure 11.

Figure 14. Results from DSDP hole 504B. Shown from left to right are, geologic section of region drilled, shear and compressional velocity sonic logs [Anderson et al.,1982], and apparent bulk porosity as a function of depth [Becker et al.,1982].

Figure 15. Generalized velocity structure of layers 2 and 3 and inferred geologic interpretation based on

the travel-time and amplitude analyses presented here, results from DSDP hole 504B [Anderson et al., 1982; Becker et al., 1982], and studies of ophiolite complexes [e.g., Salisbury and Christensen, 1978; Casey et al., 1981].

Table 1. Estimates of uncertainties (in s) in travel times arising from uncertainties in the data reduction process.

	Range, km:	5	20	35
Source:				
Instrument corrections		0.02	0.02	0.02
Picking		0.03	0.04	0.06
Water path corrections		0.03	0.01	0.01
Assumption of lateral heterogeneity		0.02	0.02	0.02
Total:				
$\sqrt{\Sigma U^2}$		0.05	0.05	0.07
ΣU		0.10	0.09	0.11

Table 2. Summary of thicknesses and velocities of the seismic units within layer 2 modeled from amplitude patterns and travel-times on line 4.

Location of amplitude peak, km:	no peak	6	8	10	
	Thickness, km				Range of velocities in layer, km/s
Layer 2A	0.65 0.70	0.65 0.70	0.65 0.70	0.65 0.70	2.1 → 5.5
Layer 2B	0.0 0.0	0.45 0.60	0.70 0.40	0.95 0.20	5.5 → 5.5
Layer 2C	0.60 2.00	0.60 0.20	0.60 0.80	0.60 1.40	5.5 → 6.9
Layer 2 = 2A+2B+2C	1.25 2.70	1.70 1.50	1.95 1.90	2.20 2.30	2.1 → 6.9

The top row for each layer is from the model set in Figure 11a; the bottom row is from the model set in Figure 11b.

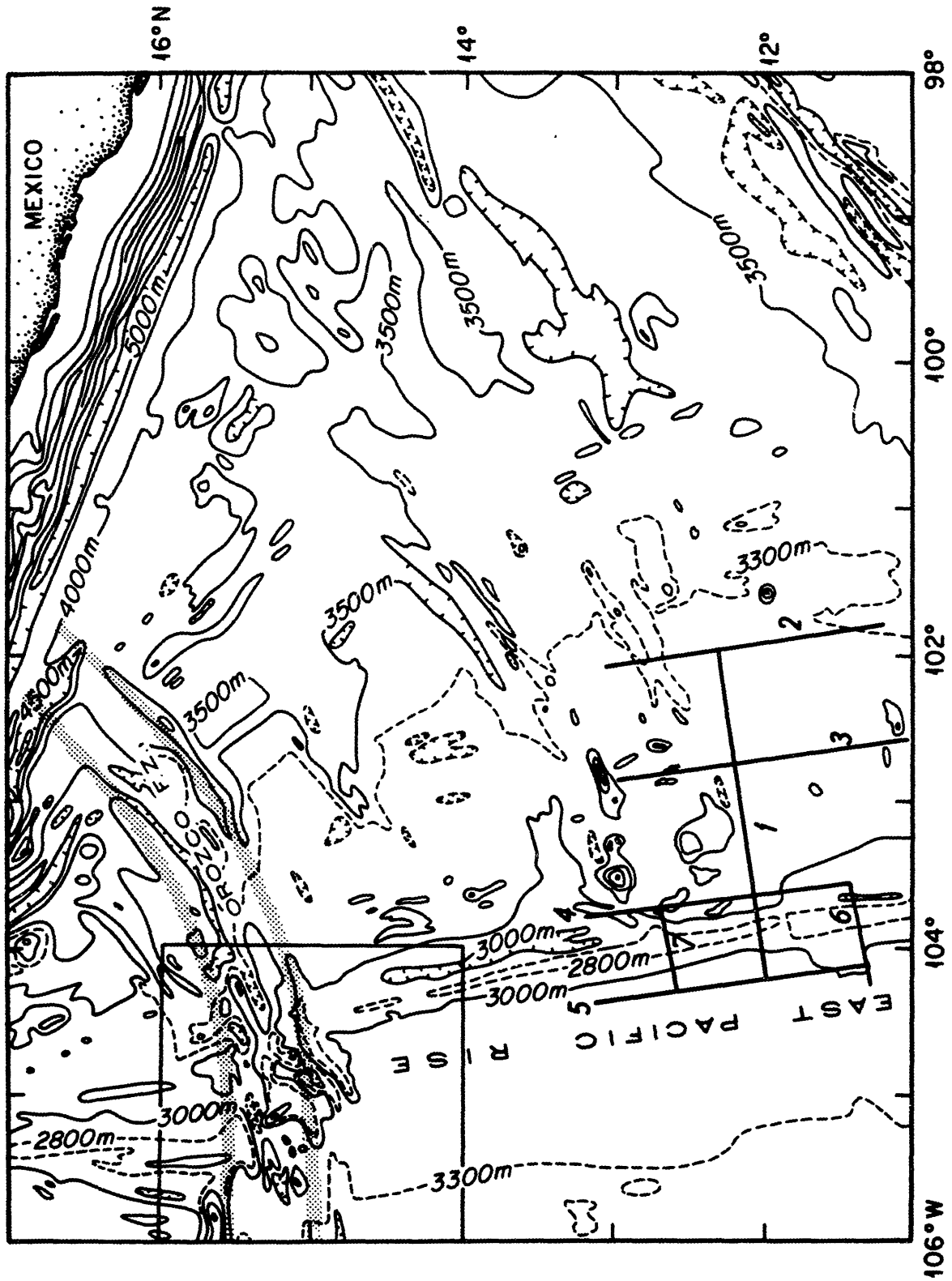


Figure 1

BATHYMETRY
LINE 4, V.E. = 50x

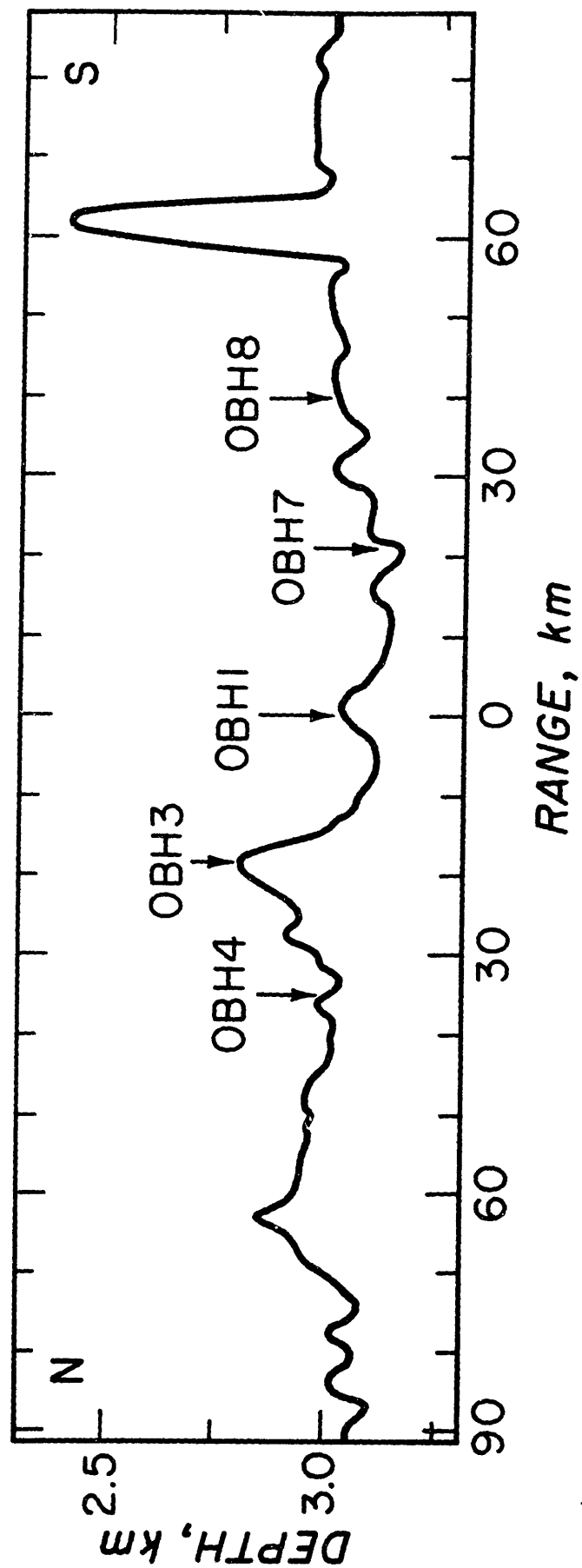


Figure 3

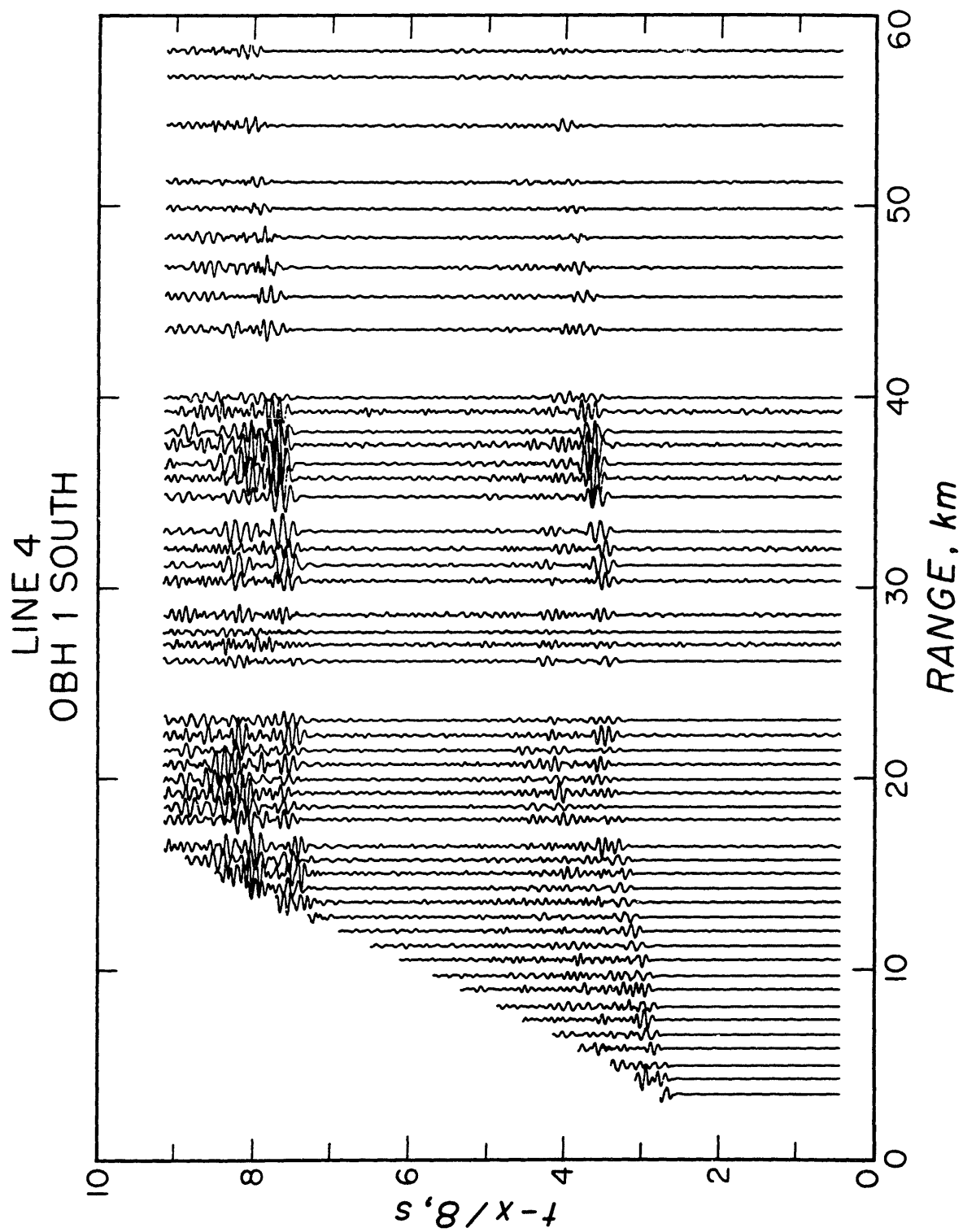


Figure 4a

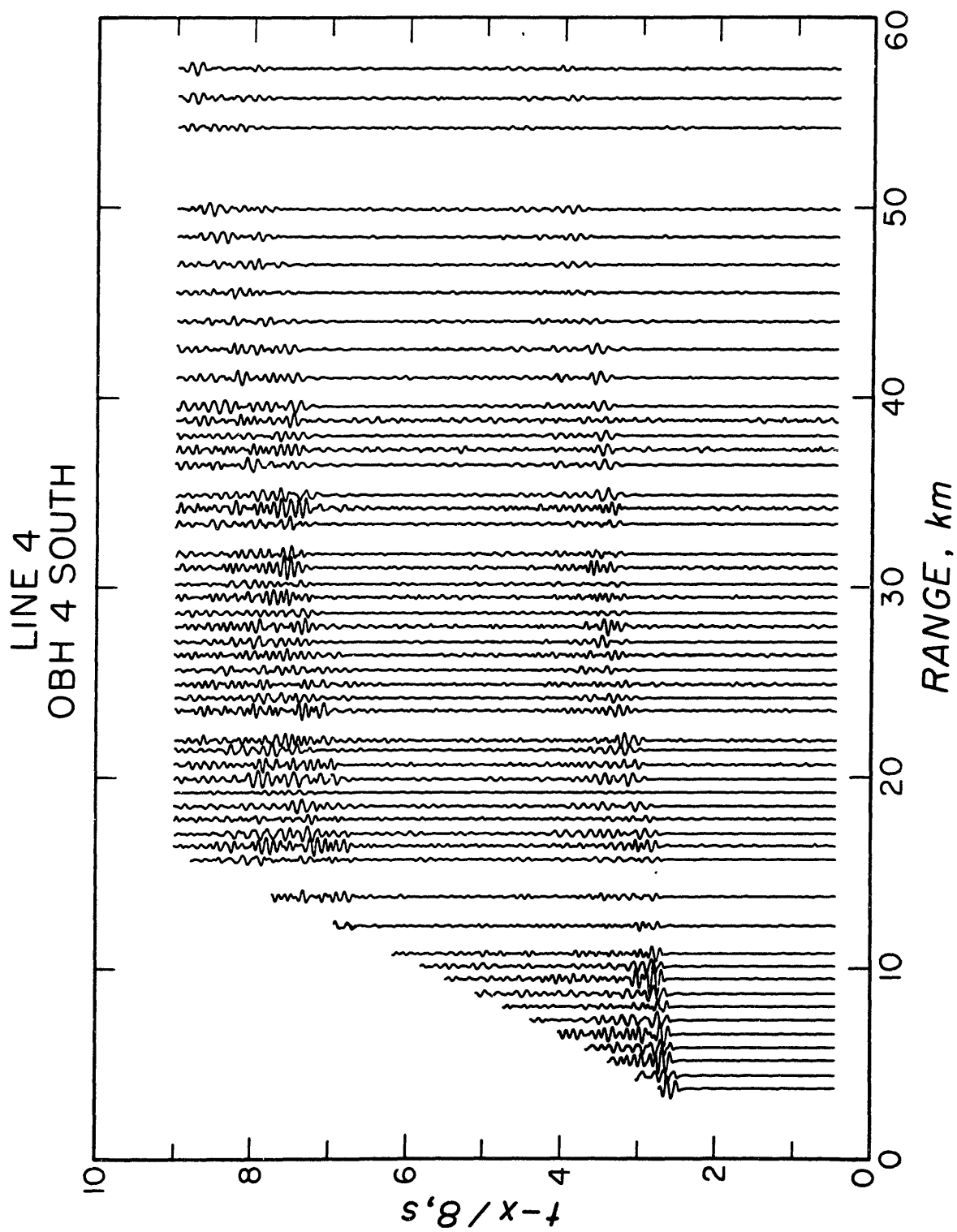


Figure 4b

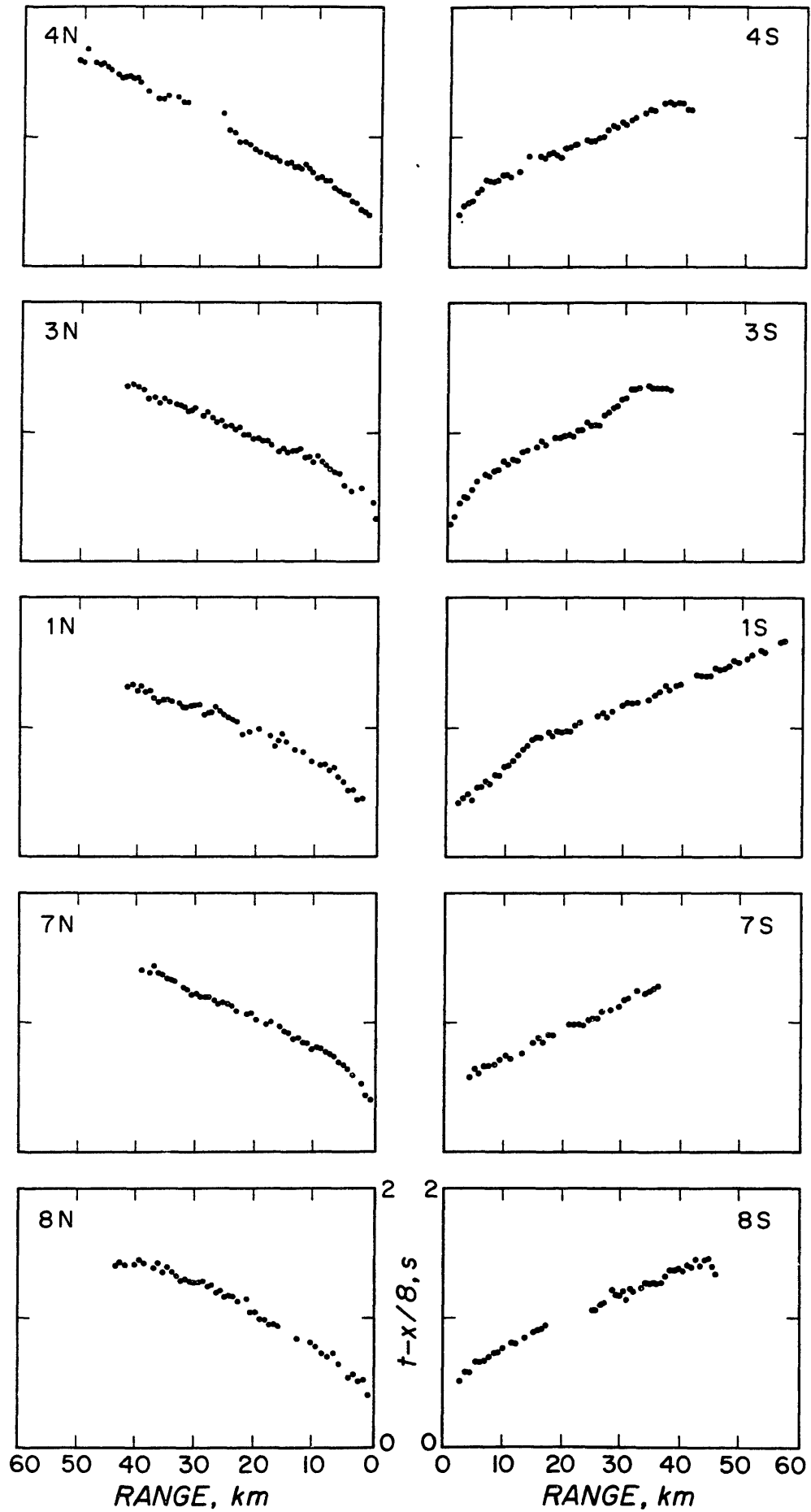


Figure 5

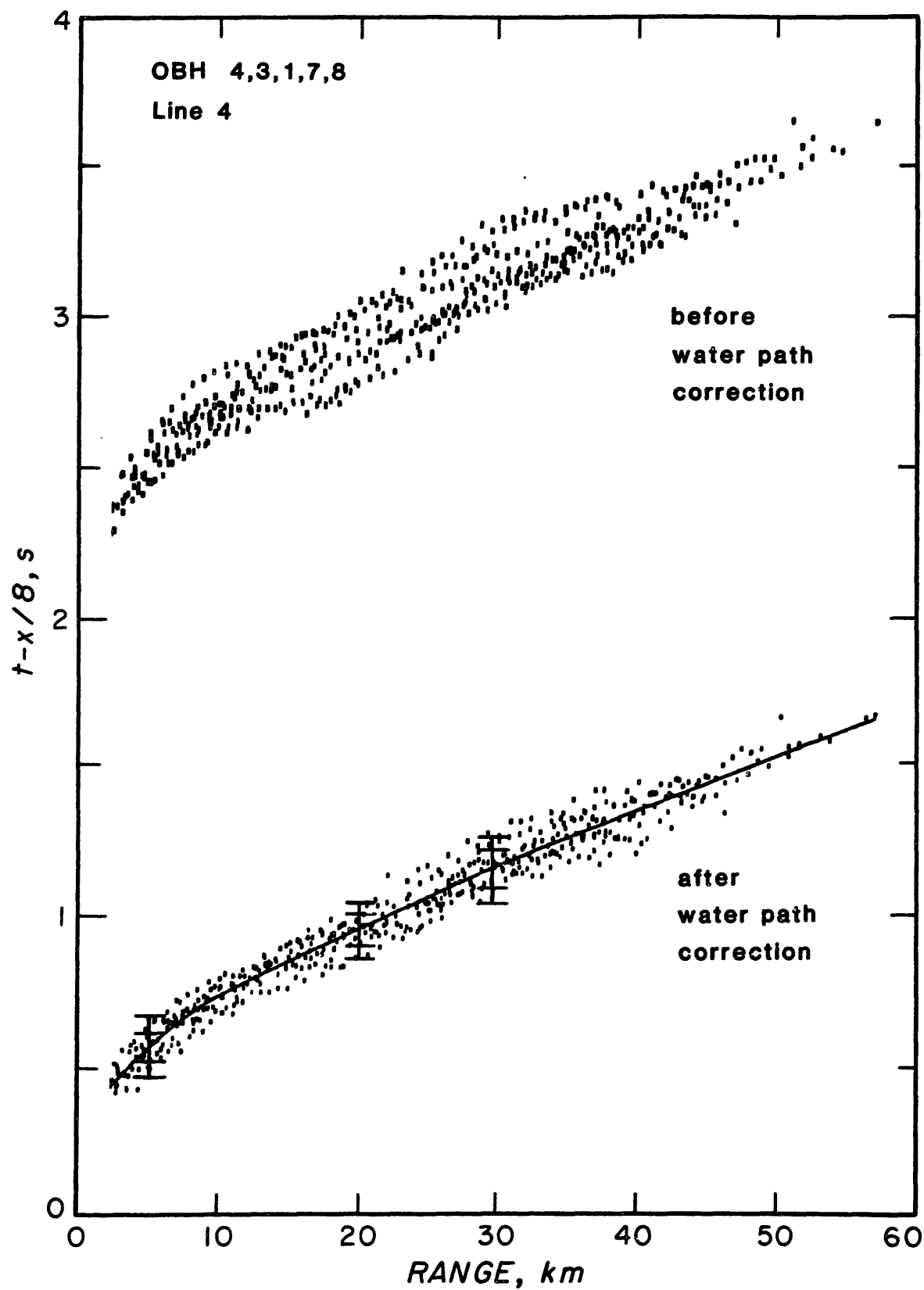


Figure 6

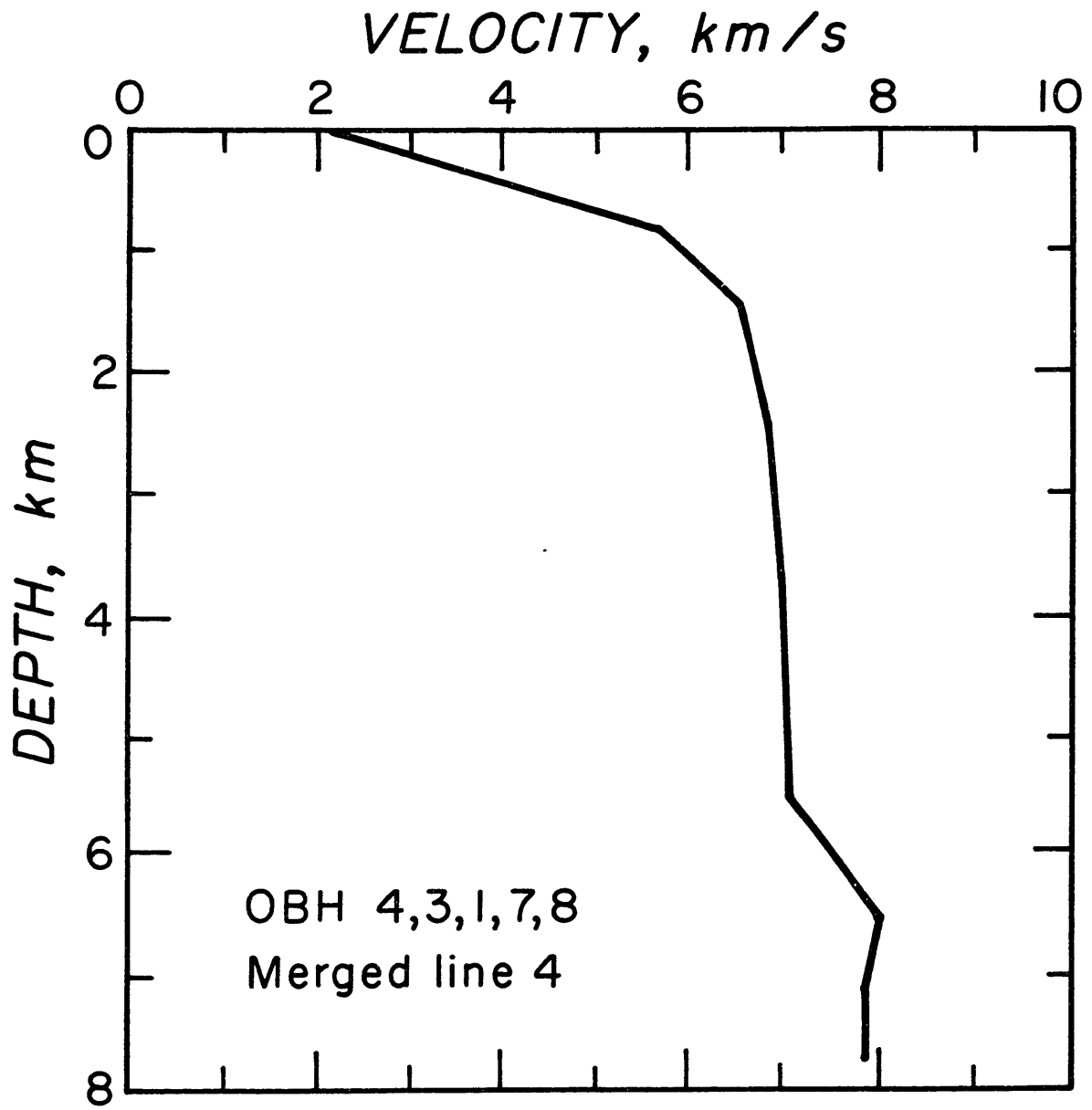


Figure 7

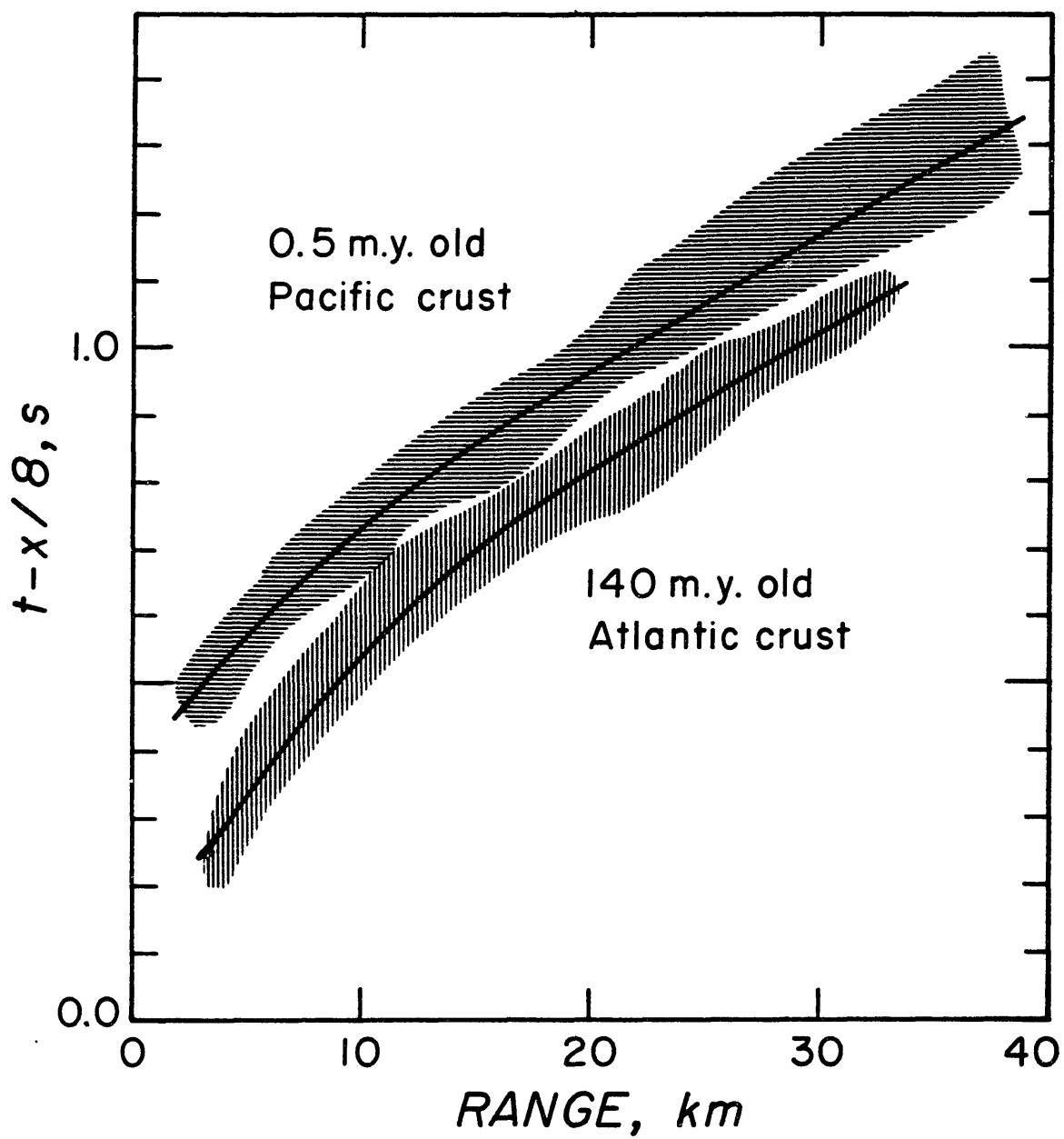


Figure 8

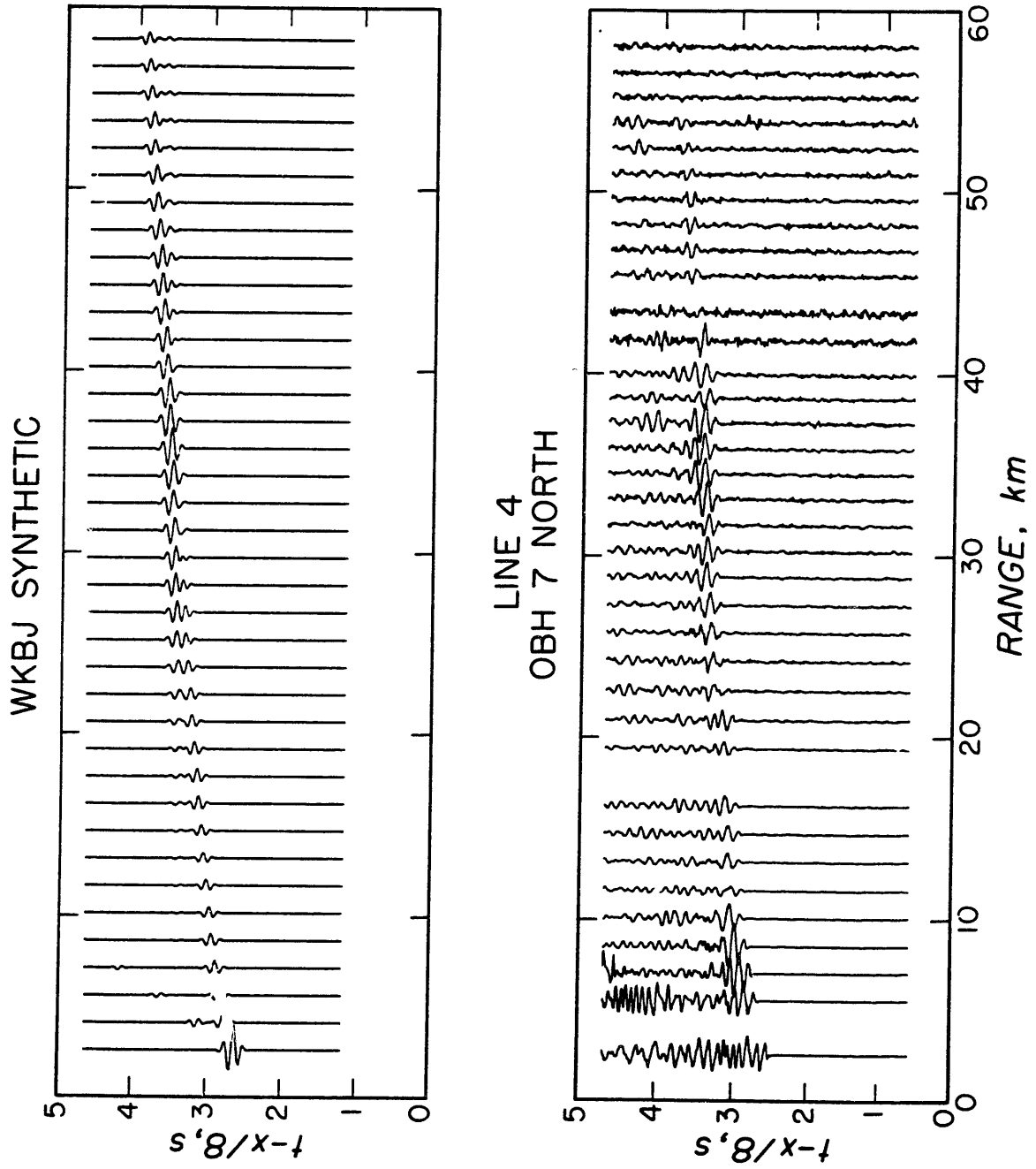


Figure 9

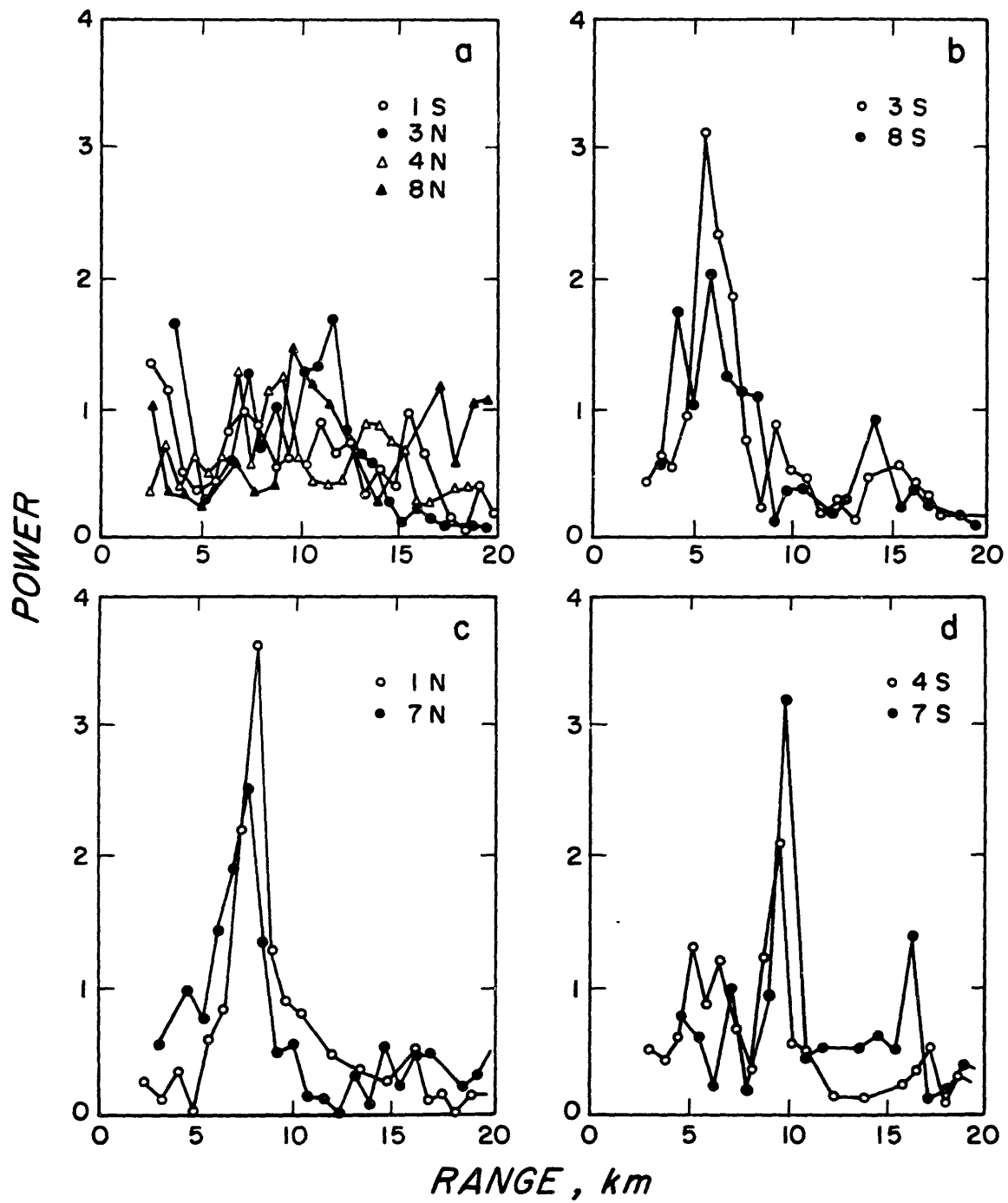


Figure 10

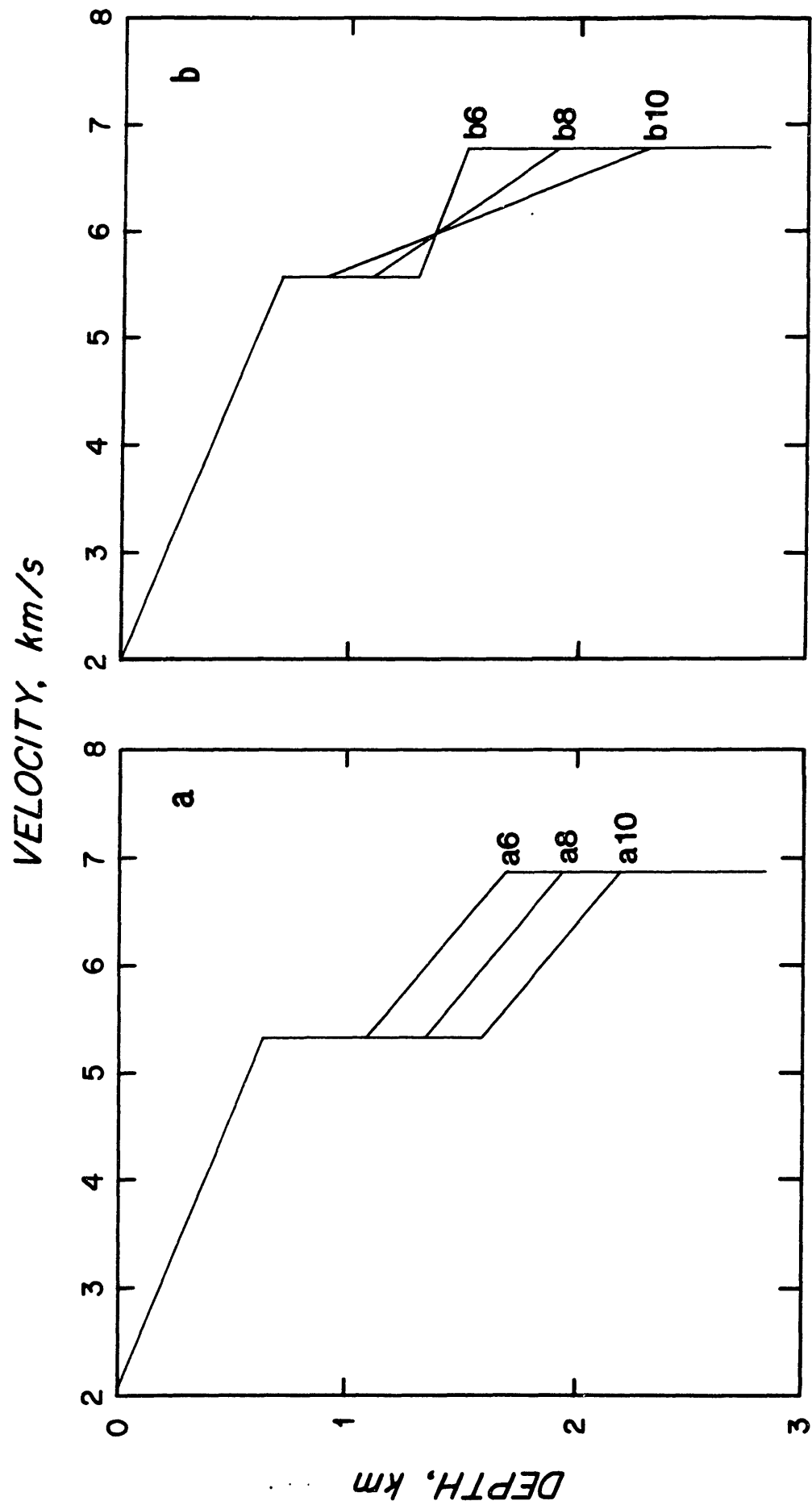


Figure 11

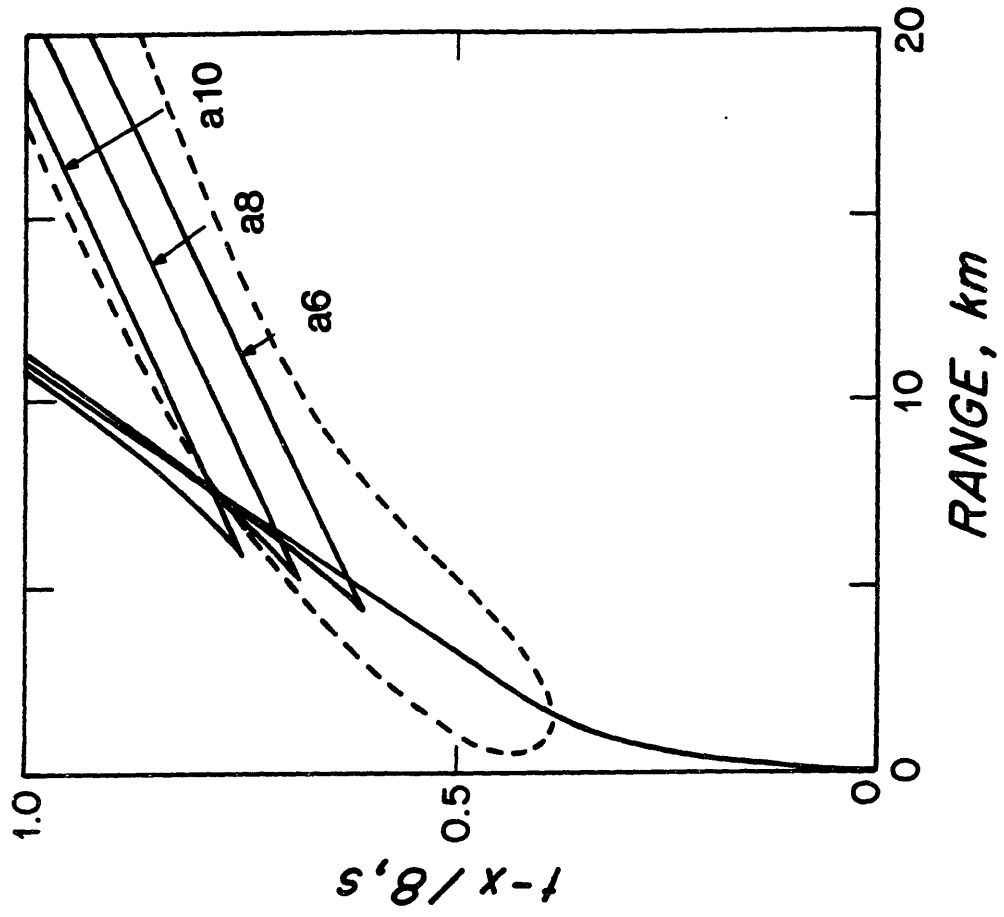
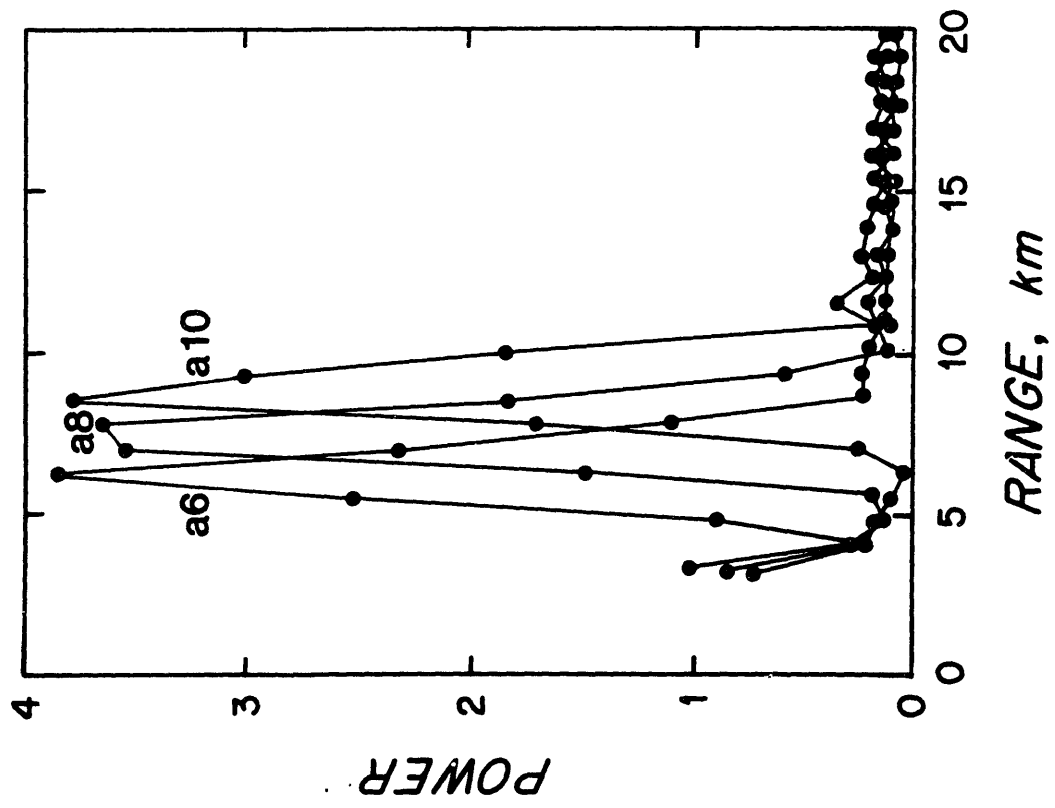


Figure 12

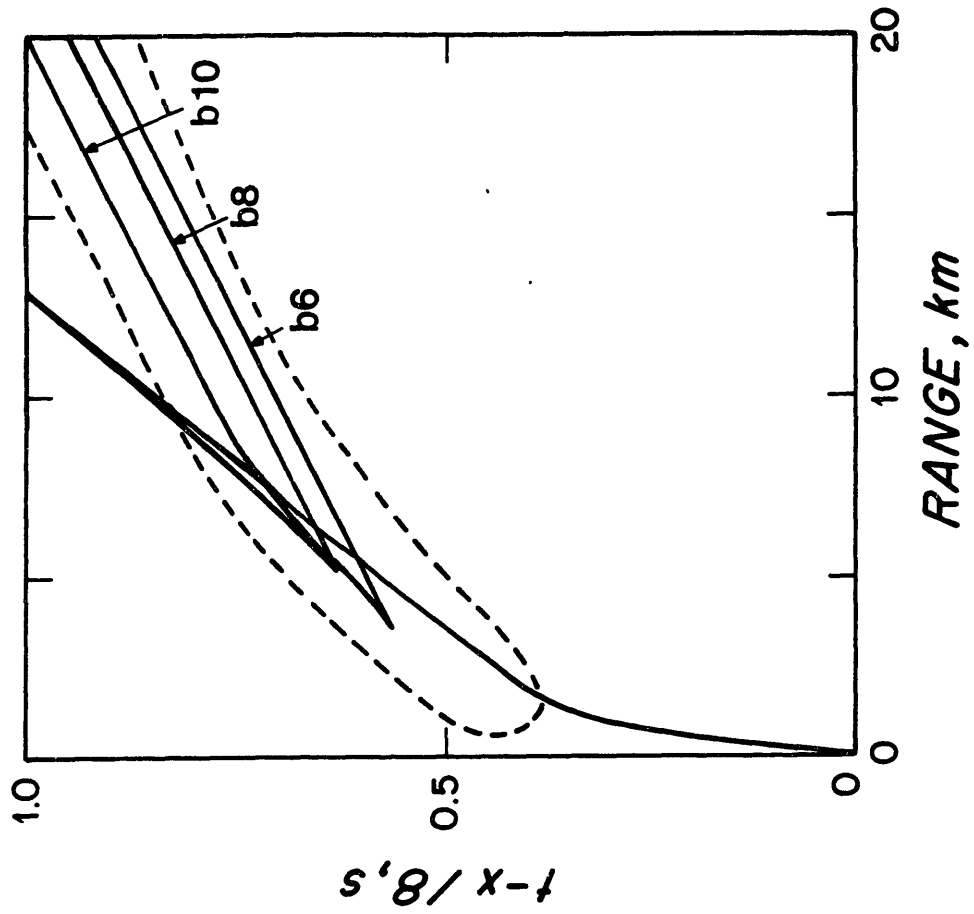
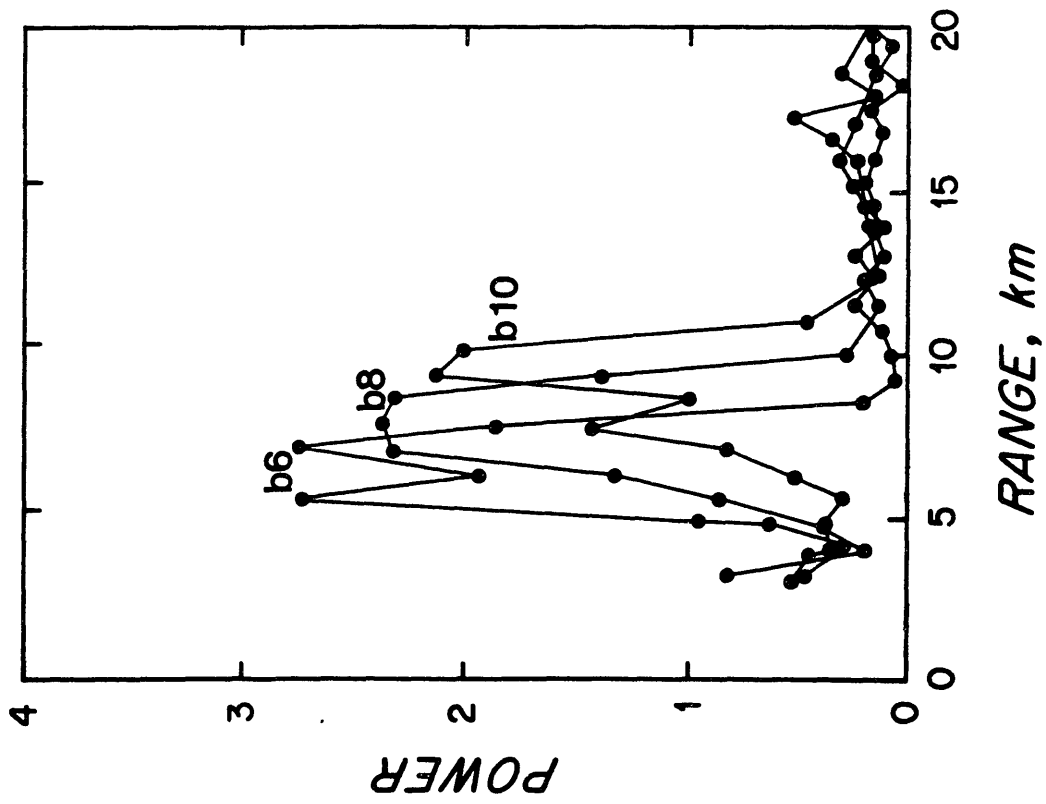


Figure 13

DSDP HOLE 504 B

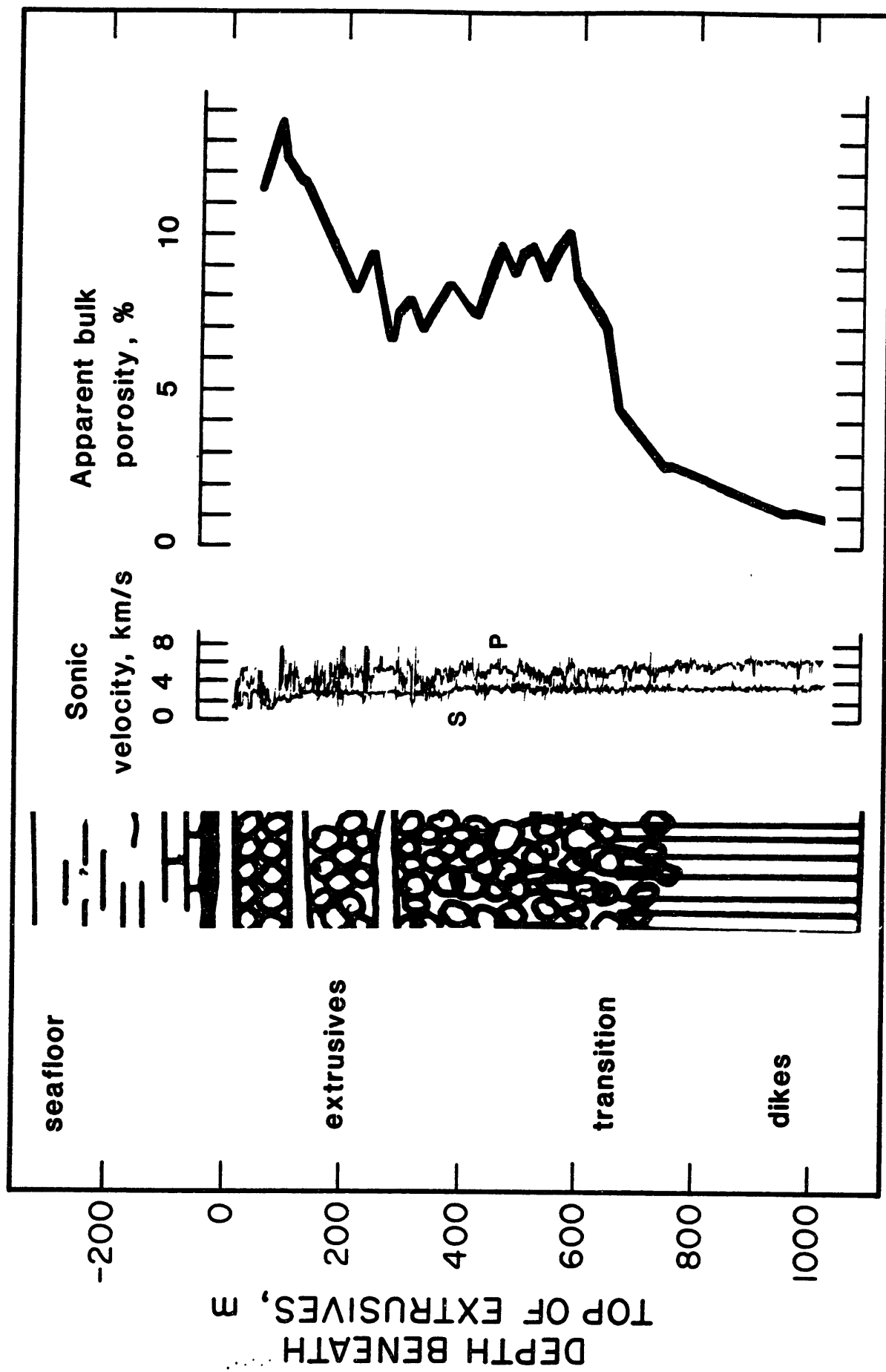


Figure 14

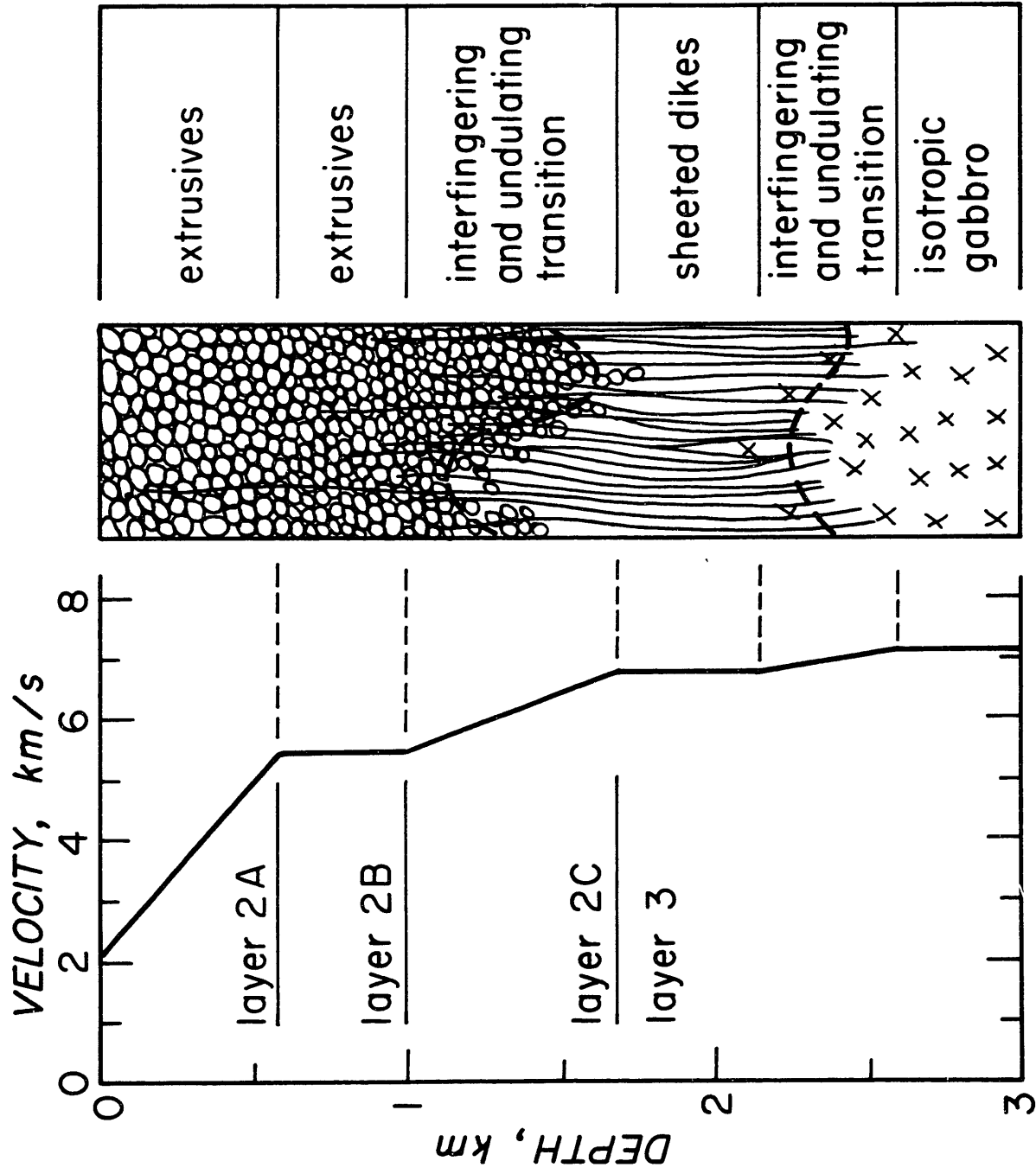


Figure 15

Chapter 4

THERMOELASTIC STRESS: HOW IMPORTANT AS A CAUSE OF EARTHQUAKES
IN YOUNG OCEANIC LITHOSPHERE?

INTRODUCTION

A potentially important contributor to the state of stress in oceanic lithosphere is thermoelastic or thermal stress. Thermal stress can develop in the elastic portion of young oceanic lithosphere because of differential cooling and contraction. It has been proposed as a major contributor to the formation of transform faults [Turcotte, 1974; Collette, 1974], the propagation of mid-plate island and seamount chains [Turcotte and Oxburgh, 1973], and the generation of earthquakes in older oceanic lithosphere [Sykes and Sbar, 1974]. In most studies of the state of stress in the lithosphere distant from plate boundaries, thermal stress generated during plate cooling has not been explicitly included in global stress models [e.g., Richardson et al., 1979; Fleitout and Froidevaux, 1983], in large part because of the expectation that thermal stress would tend to be relieved at young lithosphere ages and, once relieved, would not be renewed [e.g., Solomon et al., 1975].

Recent determinations of the mechanisms and focal depths of earthquakes in young oceanic lithosphere [Bergman et al., 1984; Wiens and Stein, 1984; Bergman and Solomon, 1984] have led us to reexamine the role of thermal stress. Specifically, the distribution of source mechanisms with depth and with distance from the ridge axis in oceanic lithosphere 2 to 35 m.y. old appears to be generally consistent with the hypothesis that many of these earthquakes represent the response of the brittle portion of the lithosphere to accumulated thermal stress

[Bergman and Solomon, 1984]. In this chapter, we present several simple models for the state of thermoelastic stress in young oceanic lithosphere, and we compare the predictions of these models with the characteristics of near-ridge earthquakes.

CHARACTERISTICS OF NEAR-RIDGE EARTHQUAKES

The principal source of data on the state of stress in young oceanic lithosphere consists of the source mechanisms and focal depths of near-ridge earthquakes. Detailed and comprehensive studies of the source mechanisms of such earthquakes have recently been completed by Bergman and Solomon [1984] and by Wiens and Stein [1984]. The data of Bergman and Solomon [1984], obtained from an inversion of long-period P and SH wave forms for 32 events from 1962-81 in oceanic lithosphere 2.5 to 35 m.y. old, are summarized in Figure 1. Individual data points are accurate to about ± 2 km in centroid depth and generally ± 2 m.y. in age; for a few events located near fracture zones or in poorly surveyed regions, the uncertainty in lithosphere age is greater [Bergman and Solomon, 1984].

On the basis of the results shown in Figure 1, several generalizations may be made. Most near-ridge events, and virtually all of the earthquakes with seismic moments greater than 10^{25} dyne-cm, occur in lithosphere less than 15 m.y. old. The majority of near-ridge events are located at mantle depths (greater than 6 to 7 km below the seafloor). While all styles of faulting are observed, there appears to be an approximate stratification to fault type. Thrust faulting is confined to

the uppermost 10 km; compressional axes both nearly parallel to and oblique to the local spreading direction are seen. Most of the strike-slip mechanisms occur at intermediate depths within the seismically active portion of young oceanic lithosphere. The deepest earthquakes at any age display predominantly normal faulting mechanisms; tensional axes are either oblique or perpendicular to the direction of spreading. The centroid depths of the normal faulting events appear to be bounded by the depth to an isotherm of about 800°C. A large fraction of near-ridge earthquakes during the last 20 years, particularly of events with normal faulting mechanisms, has occurred in the central Indian Ocean. A high level of intraplate seismicity in the Indian Ocean in both old and young lithosphere may be attributable in part to enhanced levels of lithospheric stress associated with effect of the ongoing collision between India and Asia [Stein and Okal, 1978; Weissel et al., 1980; Bergman et al., 1984]. Generally similar distributions of mechanism type and centroid depth with age in the smaller data sets from the Atlantic and Pacific oceans, however, suggest that an explanation should be sought for the patterns displayed in Figure 1 that is not peculiar to any particular oceanic region.

SOURCES OF STRESS IN OCEANIC LITHOSPHERE

The tectonics of oceanic lithosphere are not the product of any single source of stress. Contributions to the total state of stress arise from such sources as shear tractions at the base of the lithosphere [e.g., Richter and Parsons, 1975;

Hager and O'Connell, 1979], latitudinal components of plate motions [Turcotte and Oxburgh, 1973], crustal thickness and lithospheric density variations [Artyushkov, 1973], lithospheric loading and unloading [Walcott, 1970; Watts and Cochran, 1974], and some combination of pull [McKenzie, 1969; Turcotte and Schubert, 1971] and resistance [Smith and Toksoz, 1972; Forsyth and Uyeda, 1975] at subduction zones and transform faults. Many of these processes are local in their effects and thus would not be expected to contribute to faulting in any regular pattern consistent from ocean to ocean.

Perhaps the most pervasive sources of stress in young oceanic lithosphere are those related to plate cooling. These include both thermal stress [Turcotte and Oxburgh, 1973; Turcotte, 1974] and the compression, or "ridge push," that results from the elevation of mid-ocean ridges and the lateral heterogeneity of lithospheric density [e.g., Hales, 1969; McKenzie, 1972; Dahlen, 1981]. Because the effects of these processes depend predominantly on the thermal structure and material properties of the plate, ridge push and thermoelastic stress should depend on age and on depth but should be more or less consistent from ocean to ocean.

The stress field resulting from the elevation of a mid-ocean ridge relative to adjacent cooler and denser lithosphere has been evaluated by Dahlen [1981] under the assumption that deviatoric stress in the lithosphere is the minimum necessary to support the lateral variations in density and bathymetry produced by cooling. For a two-dimensional model

of cooling lithosphere, the principal deviatoric stresses are compressional in the direction of spreading, extensional and equal in magnitude in the vertical direction, and zero parallel to the ridge. At any given age, the deviatoric stresses are maximum at the seafloor and decrease with depth. The surface deviatoric stress reaches about 200 bars at an age of 30 m.y. In contrast to thermal stress, the stress associated with ridge push is "renewable" [Bott and Kusznir, 1984]; i.e., it is not significantly relieved by brittle failure. We consider below the relative importance of thermal stress and ridge push in the near ridge environment.

PREVIOUS MODELS OF THERMAL STRESS

As young oceanic lithosphere spreads from the ridge, it cools non-uniformly and, as a result, is subjected to thermal stress. Previously published models of thermal stress either have been qualitative in nature [Collette, 1974; Sykes and Sbar, 1973; Epp and Suyenaga, 1978] or have been taken from simple analytic solutions based on restrictive assumptions [Turcotte and Oxburgh, 1973; Turcotte, 1974]. The most developed such model is that of Turcotte [1974]. In that model, the ridge is assumed to be two-dimensional and to have a prescribed zero-age temperature profile. The normal component of stress in the direction parallel to the ridge (σ_{yy}) is taken to be the only non-zero principal stress. Thermal stress at each point is assumed to be the result only of local thermal contraction due to cooling below an elastic "blocking temperature" [see also Turcotte, 1983]. The local stress in this model thus does not depend on the state of stress in surrounding material.

Turcotte [1974] considered the possibility that the product of Young's modulus and thermal expansion coefficient differ between crust and mantle. If so, a discontinuity is introduced at the Moho. As a result, the lithosphere acts as a bimetallic strip. In the absence of bending, the accumulated thermal stress σ_{yy} is everywhere extensional and can exceed 1 kbar at 2 m.y. age and 3 kbar at 20 m.y. As thermal stresses are relieved by fracture, the integral of σ_{yy} over depth will tend to zero, according to Turcotte [1974], placing the upper part of the elastic plate under horizontal compression and a lower part under horizontal extension (both in the y direction). The depth of the transition from compression to extension depends on the assumed temperature profile at the ridge and the adopted physical constants for crust and mantle material.

The simple thermal stress model of Turcotte [1974] is in reasonable qualitative agreement with the earthquake data of Figure 1, including the depth dependence of thrust and normal faulting mechanisms and the tendency of the P and T axes, respectively, for these earthquakes to be oblique or perpendicular to the spreading direction. Several of the assumptions made by Turcotte [1974], however, are not likely to be valid. The component of thermal stress in the direction of spreading cannot be neglected, except perhaps very near the ridge axis. To the extent that the upper lithosphere behaves elastically, the stress at each point must also depend on the deformation and state of stress in surrounding material. Further, the relatively high level of seismicity in young

oceanic lithosphere suggests that thermal stresses are not accumulated indefinitely but may rather be at least partly relieved by recurring earthquake activity. For these reasons, a more complete treatment of thermal stress in young oceanic lithosphere is desirable.

CALCULATION OF THERMAL STRESS

In this section, we describe a new approach to the estimation of thermal stress in oceanic lithosphere. The stress models are based on the elastic response to heterogeneous cooling of oceanic lithosphere. The basic model is that of a cooling elastic halfspace, though we also consider a simple modification to the assumed temperature field to account approximately for the effects of hydrothermal circulation. Various assumptions are made regarding an effective elastic "blocking temperature" [Turcotte, 1974, 1983] and the time scale for relief of thermal stress by fracture. A number of different thermal stress models resulting from these parameter variations are considered. For each model we compare the predicted state of stress with that indicated by the mechanisms and locations of earthquakes depicted in Figure 1. On the basis of these comparisons, we assess the contribution of thermoelastic stress to the tectonics in young oceanic lithosphere.

Temperature Structure

For most of the thermal stress models, the adopted temperature structure is that of a simple cooling halfspace [e.g., Turcotte and Oxburgh, 1967]:

$$T(x,z) = T_L \operatorname{erf} \left[\frac{z}{2} \left(\frac{V}{k|x|} \right)^{1/2} \right] \quad (1)$$

where erf is the error function, T_L is the temperature at the ridge axis and at large depth z , and k is the thermal diffusivity. The coordinate system is illustrated in Figure 2. The lithosphere is taken to spread in the $\pm x$ direction outward from the ridge at $x=0$ at the half rate v . Depth is relative to the seafloor, at which the temperature is a constant 0°C . The y direction is parallel to the ridge crest. Values for all adopted physical parameters are given in Table 1. Selected isotherms given by (1) are shown in Figure 1; this simple thermal structure has been shown to be consistent with seafloor bathymetry and heat flow for lithosphere up to 80 m.y. in age [Davis and Lister, 1974; Parsons and Sclater, 1977].

A complication of the simple cooling model represented by (1) and one that may significantly affect the resulting thermal stress field arises from hydrothermal circulation in the uppermost lithosphere. Hydrothermal circulation acts to cool a substantial fraction of the crust at or near the ridge axis [Lister, 1972, 1977], so that the zero-age temperature appropriate to the calculation of thermal stress is significantly cooler than (1) to some uncertain (and perhaps variable) depth. Limited data from the depths of ridge crest microearthquakes [Lilwall et al., 1978; Toomey et al., 1984] and from oxygen isotope measurements in ophiolite samples [Gregory and Taylor, 1981] suggest that hydrothermal circulation extends at least to the Moho for some ridge systems. Heat flow data indicate that hydrothermal circulation continues to maintain a low conductive gradient at the top of the lithosphere to ages as great as 80 m.y. [Anderson et al., 1977]. To approximate the effects of ridge-axis cooling by hydrothermal circulation, we

also consider as a simple alternative to the thermal structure given by equation (1) the following model: $T = 0^\circ\text{C}$ for $z < h$, and T is given by (1) with $z-h$ substituted for z in the argument of the error function for $z > h$. We have examined models in which $h = 5$ and $h = 10$ km. A similar thermal model has been employed by Lewis [1983] in his calculation of gravity anomalies over ridge structures.

Thermoelastic Displacement Potentials

To estimate thermal stress, we employ the method of thermoelastic displacement potentials developed originally by Goodier [1937]. The response of an infinite, elastic solid to a non-uniform temperature change $\Delta T(x,y,z)$ can be represented by a distribution of centers of contraction (or dilatation) of magnitude:

$$\theta = \frac{\alpha \Delta T (1+\nu)}{12\pi (1-\nu)} \quad (2)$$

where α is the volumetric coefficient of thermal expansion and ν is Poisson's ratio (Table 1). The solution to the thermal stress problem in a semi-infinite solid requires that the surface be free of tractions. To satisfy these boundary conditions, a center of contraction, a double force, and a doublet, each appropriately weighted, are applied at the image point $(x,y,-z)$ corresponding to every center of contraction (x,y,z) within the half-space [Mindlin and Cheng, 1950]. The displacement field $\underline{u}$ is given by

$$\underline{u} = -\nabla \phi_1 - \nabla_2 \phi_2 \quad (3)$$

where the thermoelastic displacement potentials:

$$\begin{aligned}\phi_1 &= \iiint_{V'} \frac{\theta}{R_1'} dx' dy' dz' \\ \phi_2 &= \iiint_{V'} \frac{\theta}{R_2'} dx' dy' dz'\end{aligned}\quad (4)$$

are integrals of θ over volume V in the half-space, weighted by the inverse of the distance scalars

$$\begin{aligned}R_1' &= [(x-x')^2 + (y-y')^2 + (z-z')^2]^{1/2} \\ R_2' &= [(x-x')^2 + (y-y')^2 + (z+z')^2]^{1/2}.\end{aligned}$$

The vector operator applied to ϕ_2 is

$$\nabla_2 \phi_2 = (3-4\nu)\nabla\phi_2 + 2\nabla\left(z \frac{\partial\phi_2}{\partial z}\right) - 4(1-\nu)\nabla^2(z\phi_2)\underline{e_k}$$

where ∇^2 is the scalar Laplacian and $\underline{e_k}$ is the unit vector in the z direction.

It may be seen from equation (3) that, given a temperature distribution for which the potential is known, the displacement field can be readily calculated by differentiation. To model the elastic response of cooling lithosphere, we divide the half-space into rectangular boxes (Figure 2) within which the temperature change ΔT is assumed to be constant and for which the potential may be determined. The total displacement field can then be computed by summing the displacements due to the temperature changes in individual boxes. While the thermoelastic displacement potential for a uniformly cooled box is known [MacMillan, 1930], the solution involves 36 terms containing various trigonometric and logarithmic functions. To reduce computation time, we approximate the potential of a

uniformly cooled box by that of a line source [MacMillan, 1930]. The potential at (x,y,z) due to a line source of strength $4ac\theta$ located at (x',y',z') is given by

$$\phi_1 = 4 a c \theta \ln \left[\frac{S_1 - (y-b)}{S_2 - (y+b)} \right] \quad (5)$$

where

$$\begin{aligned} S_1 &= [(x-x')^2 + (y-b)^2 + (z-z')^2]^{1/2} \\ S_2 &= [(x-x')^2 + (y+b)^2 + (z-z')^2]^{1/2} \end{aligned} \quad (6)$$

and where a , b , and c are the half-widths of the box represented by the line source in the x , y , and z directions, respectively (Figure 2). The corresponding expression for ϕ_2 differs only by the substitution of $(z+z')$ for $(z-z')$ in (6).

We have tested the accuracy of this approximation by comparing the displacement field produced by a cooling sphere embedded in a half-space [Mindlin and Cheng, 1950] with that predicted by the cooling of an embedded cube of equal volume assembled by the superposition of line sources (equation 5) representing boxes of various cross-sectional shapes. This procedure was repeated using different volumes and burial depths. The line-source approximation to the potential of a box was found to be best when the represented box has a nearly square cross-section with sides that are small relative to the distance (R_1) between the observation point and the box. Following these criteria, the calculated displacements (radial and vertical) induced by the cooling of an embedded cube of half-length L differ from those due to the cooling of an embedded sphere of volume $8L^3$ by about 1% at a distance of $0.1L$ from the center of either figure, 5% at a distance of $0.6L$, 5% at a distance of $2L$, and less than 1% at a distance of $2.5L$. On the basis of this

either figure, 5% at a distance of 0.6L, 5% at a distance of 2L, and less than 1% at a distance of 2.5L. On the basis of this test, we believe that the approximation to equation (4) is generally accurate to within a few percent for arbitrary two-dimensional temperature fields.

Thermal Stress

The thermal stress resulting from differential cooling is calculated from the displacement field (equation 3) using standard formulae [e.g., Timoshenko and Goodier, 1970, p. 456]. The normal strains and stresses are of the form:

$$\begin{aligned}\epsilon_{xx} &= \frac{\partial u}{\partial x}, \text{ etc.} \\ \sigma_{xx} &= - \frac{E}{(1+\nu)(1-2\nu)} [(1-\nu)\epsilon_{xx} + \nu(\epsilon_{yy} + \epsilon_{zz}) \\ &\quad - \frac{\alpha}{3} (1+\nu) \Delta T(x,y,z)] , \text{ etc.} \quad (7)\end{aligned}$$

where $\underline{u} = (u,v,w)$, E is Young's modulus (Table 1), and compressional stresses are positive. By restricting our consideration to the plane $y=0$ (Figure 2), σ_{xy} and σ_{yz} vanish and σ_{yy} is a principal stress. The shear strain and stress in the x - z plane are given by

$$\begin{aligned}\epsilon_{xz} &= \frac{1}{2} \left(\frac{\partial u}{\partial z} + \frac{\partial w}{\partial x} \right) \\ \sigma_{xz} &= - \frac{E}{1+\nu} \epsilon_{xz}\end{aligned} \quad (8)$$

Differentiation of the smooth displacement functions is accomplished numerically using the centered difference method. From (7) and (8), the magnitudes and directions of principal stresses may readily be calculated [e.g., Fung, 1977, pp. 91-108] and the principal deviatoric stresses σ_1 (greatest compressive stress), σ_2 , and σ_3 (greatest extensional stress) derived. The style of faulting expected in any location in the

thermal stress model follows from the stress deviators [Anderson, 1951]. Throughout the discussion below, all principal stresses are deviatoric stresses unless otherwise stated.

In order to compare the predicted thermal stress field with the near-ridge earthquake data of Figure 1, we make the assumption that thermal stress is generally relieved on time scales short compared to the age of the lithosphere. This relief of stress presumably occurs by brittle failure in the upper part of the lithosphere and by ductile flow in the lower part; the earthquakes depicted in Figure 1 are a primary manifestation of the brittle component of this stress relief process. In the terminology of Bott and Kuszniir [1984], thermal stress is non-renewable: once such stress is relieved, only the subsequently accumulated stress contributes to the future stress state. Relief of stress in the oceanic lithosphere undoubtedly occurs by several processes that operate on a variety of time scales. For simplicity, however, we parameterize this process by a single characteristic time Δt , independent of depth or lithosphere age, for relief of thermal stress. In this time interval, the temperature of a parcel of lithosphere (i.e., in a Lagrangian frame) changes by the amount

$$\Delta T(x,z,\Delta t) = T(x,z) - T(x-v \Delta t, z) \quad (9)$$

where $T(x,z)$ is taken from one of the thermal models described earlier. Equation (9) is independent of y , symmetric about x , and tends to zero with increasing x and z . The magnitude of ΔT

is approximately linear with Δt for locations distant from the ridge axis. Values of ΔT for the halfspace cooling model are given in Figure 3a for $\Delta t = 4$ m.y.

We calculate the thermal stress accumulated during the time interval Δt by substituting relation (9) into equations (2-3) and (5-8). With this approach, Δt may be regarded as a parameterized recurrence time for total relief of stress. In lithosphere younger than Δt , (9) is modified so that stress accumulation occurs over a time equal to the lithosphere age.

The thermal stress field computed from equations (2-3) and (5-9) will be valid only if the rocks subjected to the temperature change ΔT behave elastically over the entire temperature range. It is more likely, however, that at high temperatures stress will be rapidly relieved by ductile flow and thus will not be transmitted to surrounding regions. While we cannot account for this flow explicitly in the elastic model employed in this chapter, we can simulate its effect by postulating that material does not contribute to stress until it cools below an elastic blocking temperature T_e [Turcotte, 1974, 1983]. By this postulate, the magnitudes of $T(x,z)$ and $T(x-v\Delta t,z)$ in equation (9) are restricted not to exceed T_e , and the deepest non-zero contour of ΔT will follow closely the $T=T_e$ isotherm. We take T_e to be the isotherm approximately marking the base of the mechanically strong lithosphere. For an assumed stress difference of 500 bars, McNutt and Menard [1982] concluded that $T_e \approx 550^\circ\text{C}$ if laboratory deformation measurements can be extrapolated to geological strain rates. Contours of ΔT

for $\Delta t = 4$ m.y. and $T_e = 550^\circ\text{C}$ are shown in Figure 3b. The limitations of including a blocking temperature as a means of accounting for high-temperature anelastic behavior are discussed further below.

Assessment of Procedure

The procedures described above for the estimation of thermal stress involve several oversimplifications to the actual behavior of cooling oceanic lithosphere. A significant limitation of the model is the assumption that the oceanic lithosphere can be represented as a uniform elastic half-space. Because of the consequent continuity of displacements and stresses across the ridge axis, strains produced on one side of the ridge axis affect the stress field in the adjacent plate. A stress-free boundary or wedge along the ridge axis would be preferable, but cannot be readily accommodated within the scheme adopted here.

Two potentially important contributors to the lithospheric stress field which we have not included are variations in physical properties (e.g., between crust and mantle) and plate bending. As shown by Turcotte [1974], these effects can modify the thermal stress field in oceanic plates. Similarly, temperature- and depth-dependent rheology cannot be represented in a simple model such as that presented here, but the use of an elastic blocking temperature at least provides a rough approximation to the effects of stress relaxation by ductile flow. The time interval over which stress accumulates before failure is not likely to be uniform throughout the oceanic

lithosphere as is assumed in our model. Rather, the characteristic time for relief of stress must be a function of depth and age because of the dependence of mechanism and rate of deformation on temperature, effective pressure, strain rate, and state of stress [Brace and Kohlstedt, 1980]. Stress may accumulate over relatively long time intervals without failure in lithospheric regions where rates of cooling and strain are small and temperatures are low. The characteristic time for stress release may be considerably less near the ridge crest, where both temperatures and strain rates are relatively high and where at least the upper crust is likely to be cracked and porous and therefore weak [e.g., Lister, 1977; Goetze and Evans, 1979]. We return to this point below in the discussion of individual models.

The thermal stress models in this chapter do not include the effects of offsets of or modifications to the thermal structure at transform faults or fracture zones. Although such effects might be incorporated into the procedures used here by superposing cooling blocks in three dimensions to approximate actual distributions of seafloor ages, we do not consider such complications in this chapter. The models below are therefore appropriate only to regions far from fracture zones and transforms where temperature gradients in the y direction are negligible. A further concern is the possibility that transform faults and fracture zones are regions where the mechanical strength is low and the assumption of elasticity is invalid. While this possibility is likely for transforms, profiles of

bathymetry and gravity across inactive fracture zones indicate that the elastic lithosphere is essentially continuous across such features [Sandwell and Schubert, 1982].

The procedure for calculation of thermal stress adopted in this chapter has several advantages over earlier, simpler models. Whereas the models of Turcotte and Oxburgh [1973] and Turcotte [1974] consider the thermal stress due only to contraction at each point in isolation, the stress field determined here includes the contributions to stress from deformation in the surrounding medium. Furthermore, the method of thermoelastic displacement potentials enables us to consider the complete stress tensor without the need to make a priori assumptions about the magnitude of any particular stress component. Finally, it is straightforward to include a characteristic time for release of thermal stress. Through this characteristic time parameter we can also scale the predicted magnitude of thermal stress to other potential time-independent stress fields such as that arising from ridge push.

THERMAL STRESS MODELS

We now consider a suite of models for thermal stress that incorporate the various assumptions and parameterizations discussed above. As an instructive precursor to full stress models of oceanic lithosphere, we first briefly consider the thermoelastic stress resulting from the cooling of a simple elongated box within an otherwise uniform halfspace (Figure 4). The stress field can be regarded as an approximate Green's function, which when convolved with a distribution of

temperature changes such as those given by equation (9) will yield the full thermal stress field in the oceanic lithosphere problem. The magnitude of the thermal stress field scales with the initial temperature of the box (10°C). The box, 6 by 6 km in the x and z directions, is approximated by 9 line sources of half-length $b = 10,000$ km spaced 2 km apart in x and z. Within the cooling box, all principal values of the full stress tensor are extensional. Strain in the y direction is negligible within the box because the problem is nearly two-dimensional; as a result, σ_3 is in the y direction. Above and below the box are regions of horizontal compression; zones of horizontal extension are present on either side. In general, the σ_3 directions outside the cooling box tend to be oriented toward the box center.

Models for the thermal stress field in oceanic lithosphere can be thought of as a superposition of stress fields of the sort depicted in Figure 4, each weighted by the appropriate change in temperature. We consider a suite of models that variously incorporate the assumptions and parameters discussed earlier (Table 2), including half-space cooling, a cold layer over a cooling halfspace to simulate the effects of hydrothermal circulation, the incorporation of an elastic blocking temperature, a variable characteristic time for release of stress, and the superposition of ridge push. In all of the models below, the temperature and stress models are essentially two dimensional ($b = 10,000$ km). Thus ϵ_{yy} is small, and the models should be regarded as appropriate only to regions of

oceanic lithosphere at some distance from transform faults and fracture zones.

Half-Space Cooling

We first explore the thermal stress resulting from simple half-space cooling (model A). No elastic blocking temperature (i.e., $T_e = T_L$) and no hydrothermally-cooled layer (i.e., $h = 0$) are included. The deviatoric stress field for $\Delta t = 4$ m.y. is shown in Figure 5. The principal values of the full stress tensor are everywhere extensional, and the two nearly horizontal principal stresses are similar in magnitude and larger by one to two orders of magnitude than the nearly vertical principal stress. This stress state can be understood by considering a distribution of simple cooling boxes of the type depicted in Figure 4, weighted according to the distribution of incremental temperature given in Figure 3a. The state of generally horizontal deviatoric extension (Figure 5) is partly a consequence of the large temperature changes occurring near the ridge, which in the elastic solution act to pull material at greater ages toward the axis. This effect is magnified in the half-space model, because the continuity of normal stress across the ridge axis allows contraction on one side of the ridge to produce strains, the strongest of which is ϵ_{xx} , in material on the opposite side. These two effects are larger than the horizontal deviatoric compression predicted above any single cooling element (see Figure 4).

In Figure 5, normal faulting is predicted for all regions of the crust shown. Of course, ductile flow rather than brittle

failure is likely to be the dominant mechanism of strain release in those portions of the lithosphere at temperatures greater than 600-800°C, i.e., the lower left corner of the figure (see Figure 1). Because σ_2 and σ_3 are both extensional and are similar in magnitude (within 50%) throughout much of the lithosphere in model A, the predicted T-axes of normal faulting earthquakes produced by this stress field could vary in orientation from directions parallel to directions perpendicular to the spreading direction, with the result probably dependent on disturbances from local sources of stress. The maximum differential stress $\Delta\sigma$ at each lithospheric age occurs at the depth where the temperature change ΔT is greatest. The magnitude of differential stress depends on the assumed characteristic time for release of stress. For $\Delta t = 4$ m.y. (Figure 5), $\Delta\sigma$ exceeds 5 kbar in 8-m.y.-old lithosphere and exceeds 1 kbar at 30 m.y. Very near the ridge axis, the rapidly-changing temperature field predicts differential stresses in excess of 10 kbar for this model. It is likely, however, that accumulated stress will be released on a time scale much shorter than 4 m.y. in this part of the lithosphere, so such large stresses need never be sustained.

Model A thus matches the observed normal faulting mechanisms and the variable orientation of the T axes relative to the spreading direction for the deeper near-ridge earthquakes depicted in Figure 1. Further, any model with a constant Δt will account approximately for the decrease in the release of seismic moment with increasing lithosphere age because $\Delta\sigma$ also

decreases with age. (Equivalently, and more physically reasonable, would be a model in which $\Delta\sigma$ at failure depends on z but not on age; model A could then be interpreted as having Δt increase with age, in agreement with the observed distribution of near-ridge earthquakes.) However, model A fails to predict horizontal deviatoric compression and thus does not agree with the observation of thrust and strike-slip earthquakes within the uppermost 12 km of young oceanic lithosphere (Figure 1).

Effect of a Hydrothermally Cooled Layer

We next consider a thermal stress model (B) that differs from model A by its approximate incorporation of the effect of cooling a surficial layer by hydrothermal circulation. The direction of principal stresses and magnitude of deviatoric stress for model B are shown in Figure 6. The model incorporates a cold layer 10 km thick and a characteristic time Δt for stress release of 4 m.y. Because the geotherms are depressed relative to simple half-space cooling, model B predicts a near-ridge region where σ_1 is nearly horizontal. The effect is similar to that illustrated by the simple box model of Figure 4 in which contraction within the box leads to horizontal compression in the region above. A zone of predicted thrust faulting in young shallow lithosphere extends to ages just greater than Δt , or about 6 m.y. for model B. With increasing lithosphere age, the direction of maximum extension rotates increasingly toward the horizontal, a consequence of the large temperature changes and associated thermal strains near the ridge axis. For $\Delta t = 4$ m.y., $\Delta\sigma$ exceeds 5 kbar in 5-m.y.-old

lithosphere and is greater than 1 kbar at depth in lithosphere 25 m.y. in age. The pattern of stresses below 10 km depth is similar to that in model A shifted deeper by about 10 km.

In general, many of the characteristics of the stress field inferred from earthquake focal mechanisms (Figure 1) are reasonably well matched by model B. Within the region dominated by normal faulting, the horizontal deviatoric stresses σ_2 and σ_3 are similar in magnitude (to within 50%), in agreement with the variable orientation of T-axes observed for normal faulting earthquakes. Thrust and strike slip failure is predicted out to 10 m.y. age and to depths where these earthquake styles are observed in young lithosphere. Because the depth to which hydrothermal circulation cools the lithosphere determines the depth extent of horizontal deviatoric compression, a thinner hydrothermally cooled layer would not fit the earthquake observations as well as model B. The simplicity of the cooling and stress models, however, should discourage drawing significant conclusions on the depth extent of hydrothermal circulation from these results. A problem with model B is that the magnitude of $\Delta\sigma$ is small and only normal faulting is predicted in the upper 10-20 km of lithosphere at ages substantially greater than Δt . In particular, the thrust and strike slip earthquakes observed between 8 and 20 m.y. age (Figure 1) are not matched by model B. Also, even in the region of thrust faulting, model B predicts that the P-axes should be in the direction of spreading. The few thrust earthquakes to date observed in very young lithosphere do not show a consistent

alignment of the P axis and the spreading direction [Bergman and Solomon, 1984; Wiens and Stein, 1984], but the statistical sampling is too small to be confident of this result.

Effect of an Elastic Blocking Temperature

As discussed earlier, lithospheric material at temperatures above some "blocking" temperature T_e may not contribute significantly to the thermal stress field [Turcotte, 1974, 1983]. We therefore consider next a model (C) for thermal stress in oceanic lithosphere that is similar to model A except that an elastic blocking temperature T_e of 550°C is assumed. The directions of principal stresses and magnitude of differential stress for model C are shown in Figure 7.

Differences between models A and C can be attributed entirely to differences in ΔT (compare Figures 3a and 3b). In general, the differential stress in model C is small in magnitude at depths below the 550°C isotherm. (That the thermal stress is non-zero below this depth is a result of the elastic half-space solution.) This pattern is in qualitative agreement with the distribution with depth and age of near-ridge earthquakes (Figure 1), though the precise isotherm corresponding to the greatest depth of earthquake activity depends strongly on the cooling model. For simple halfspace cooling, this isotherm is between 700° and 800°C [Chen and Molnar, 1983; Wiens and Stein, 1983, 1984; Bergman and Solomon, 1984]. In contrast to model C, model A predicts differential stresses in excess of several kilobars in deep regions of young lithosphere where earthquakes are not observed.

In model C, the magnitudes of σ_2 and σ_3 are more nearly equal (within 20%) than in models A and B. The predicted T-axis of a normal faulting earthquake (at a depth above the 550°C isotherm) could lie at any azimuth between the x and y axes, in agreement with observations. As with Model A, however, no thrust or strike-slip failure is predicted in the upper part of the lithosphere.

We consider simultaneously the effects of a hydrothermally cooled layer and an elastic blocking temperature in Model D (Figure 8). Model D predicts a stress field not unlike that of Model B (Figure 6), except that the magnitude of $\Delta\sigma$ is everywhere less and stresses decrease rapidly at depths corresponding to temperatures greater than T_e . Because of the simple approximation used here to simulate the effects of hydrothermal circulation, isotherms are deeper by h at all ages relative to those shown in Figures 1 and 3b. It is interesting to note that the resulting location of the $T_e = 550^\circ\text{C}$ isotherm corresponds to the locus of the deepest observed near-ridge earthquakes (Figure 1) when $h = 5$ to 10 km. Within the normal faulting regime of model D, the magnitudes of σ_2 and σ_3 agree to within 20%, a match to the observed pattern of normal faulting. The region of predicted thrust failure extends to almost 9 m.y. age, though the magnitude of $\Delta\sigma$ drops rapidly at ages greater than Δt (4 m.y.). Again, the predicted direction of thrusting in the model is the x direction. A region of strike-slip failure is predicted at the base of the hydrothermal region. This may be compared with a concentration of strike-slip

earthquakes observed in lithosphere between 8 and 21 m.y. old and 5 and 12 km depth (Figure 1). The value of $\Delta\sigma$, however, is less than 100 bars over much of the strike-slip region in model D.

Increasing the Characteristic Time for Stress Release

Model E is identical to model D except that the characteristic time Δt for stress release is 10 m.y. rather than 4 m.y. The stress field and predicted faulting patterns for model E are shown in Figure 9. The predicted stress magnitudes increase with increasing Δt , while the sense of faulting is generally independent of Δt at ages greater than this quantity. The greater period of stress accumulation before release, however, leads to a field of thrust and strike-slip faulting that extends to greater age than in model D. Thrust faulting is predicted to about 14 m.y. age, and $\Delta\sigma$ remains greater than 500 bars until about 12 m.y. Most of the zone of strike-slip failure is, as in model D, characterized by small magnitudes of predicted differential stress. Normal faulting with a variable T-axis direction is predicted below 10 km depth. With $\Delta t = 10$ m.y., in excess of 2.5 kbar of differential stress is permitted to accumulate within the normal faulting field in lithosphere 25 m.y. old. In general, the orientations of principal stresses in model E are in approximate agreement with those inferred from the mechanisms of near-ridge earthquakes (Figure 1). Because of the large and uniform value of Δt , however, the observed age-dependence of moment release for young oceanic lithosphere is not as well matched by this model as by earlier ones.

Contribution of Ridge Push

As discussed above, thermoelastic stress is not the sole contributor to the state of stress in young oceanic lithosphere. In particular, ridge-push forces resulting from the elevation of ridge crests and the associated horizontal density variations in oceanic lithosphere represent a persistent source of stress that may play an important contributing role in the distribution of seismicity and earthquake focal mechanisms. The horizontal deviatoric compression and vertical deviatoric extension generated by ridge push should enhance, to some degree, the likelihood of thrust failure in the upper 10-15 km of oceanic lithosphere.

Dahlen [1981] derived an expression for the state of stress resulting from ridge-push forces under the assumptions that σ_{xx} and σ_{zz} are the only non-zero terms in the deviatoric stress tensor, that the lithosphere is in a state of isostatic equilibrium, and that deviatoric stresses within the lithosphere have relaxed to the minimum necessary to support the lateral variations in bathymetry and density. The stress field contributed by ridge-push is then

$$\sigma_{xx} = -\sigma_{zz} = g \alpha T_L \rho_m \left(\frac{k|x|}{v} \right)^{1/2} \text{ierfc} \left[\frac{z}{2} \left(\frac{v}{k|x|} \right)^{1/2} \right] \quad (10)$$

where g is gravitational acceleration, ρ_m is the density of uncooled lithosphere (i.e., at $T = T_L$), and the function ierfc is the first integral of the complementary error function [Gautschi, 1964].

Contours of equal $\Delta\sigma = \sigma_{xx} - \sigma_{zz}$ predicted from equation (10) are plotted in Figure 10. Adopted values of the necessary

physical parameters are given in Table 1. Compression in the direction of spreading increases with age and decreases with depth. The magnitude of differential stress in the crust accumulates to over 200 bars by 10 m.y. age and over 400 bars by 30 m.y. In the regions of young lithosphere where thrust and strike-slip events dominate failure (5-12 km depth, 3-15 m.y. age), $\Delta\sigma$ ranges from 100 to 250 bars.

If ridge-push forces strongly influenced seismic failure in young oceanic lithosphere, one would expect an increase in the predominance of thrust faulting with increasing plate age and a tendency for the P-axes of earthquake mechanisms to be parallel to the spreading direction. Because the mechanisms of near-ridge earthquakes display quite different characteristics, Bergman and Solomon [1984] concluded that the stress differences typical of oceanic lithosphere less than 35 m.y. in age probably exceed the stress differences produced by ridge push alone. By this reasoning, in stress models that combine thermal stress and ridge push, we should seek thermal stress contributions having $\Delta\sigma$ in excess of 200 to 400 bars for lithosphere ages between 15 and 35 m.y. In the context of the models of this chapter, the scaling of thermal stress to any renewable source of stress is through the characteristic time Δt for stress release. This condition on $\Delta\sigma$ is satisfied by values for Δt in the vicinity of several million years.

A stress model (F) consisting of the superposition of the ridge push stress field of Figure 10 with the thermal stress field of model D (Figure 8) is depicted in Figure 11. The most

significant changes in this model compared to model D are the expansion of the shallow thrust faulting field to encompass all lithosphere ages and an increase in the magnitude of $\Delta\sigma$ in the uppermost lithosphere. Within the region of predicted thrust faulting, $\Delta\sigma$ decreases with increasing age for ages less than about 10 m.y. and increases (more slowly) with age for ages greater than 10 m.y. as ridge push begins to dominate thermal stress in the uppermost lithosphere. The P-axes of thrust faulting earthquakes predicted by the model are in the direction of spreading, a result consistent with only a minority of the observed thrust events. The thin zone of predicted strike-slip faulting is somewhat deeper in model F than in model D; the zone of predicted normal faulting is quite similar in the two models.

DISCUSSION

While the comparison of stress predicted by these simple models to the mechanisms and distribution of near-ridge earthquakes supports the hypothesis that thermoelastic stress is an important contributor to the earthquake-generating stress field in young oceanic lithosphere, it is important to recall the limitations of the modeling scheme adopted here and to suggest alternative approaches that might be pursued in future models. The elastic half-space assumption, for instance, is probably not applicable near the ridge axis. Were the ridge axis modeled as a free surface or as a wedge of ductile material, σ_{xx} in the elastic part of very young lithosphere should be less extensional than in the elastic half-space

solution. This is because the ridge axis would not act to restrict strain in the x direction and thermal contraction on one side of the ridge would not be transmitted to the other side. Whether thrust faulting in the uppermost lithosphere would be predicted in such models without the presence of a cooled surface layer remains to be demonstrated.

The half-space model for thermal stress is also inappropriate near transform faults. If transforms act to localize relief of stress, then σ_{yy} in the uppermost part of the adjoining lithosphere may be less extensional than in the halfspace solution. Incorporation of such effects might lead to the prediction of thrust faults with P -axes more nearly perpendicular to the spreading direction, in contrast to the restriction of predicted P axes in the thrust faulting domain to the x - z plane in our models. A transform with significant offset would also be the site of significant heat conduction in the y direction. Three-dimensional models of temperature and thermal stress are necessary to assess the significance of this effect.

A major simplification imposed in all of the models presented here is that the characteristic time for release of accumulated thermal stress is a constant, independent of depth and age. It is more likely, however, that stress release is governed by criteria for failure and flow that vary with depth, age, and mode of deformation [e.g., Goetze and Evans, 1979; Brace and Kohlstedt, 1980]. It would be worthwhile in future models of thermal stress in oceanic lithosphere to consider a

stress release scheme based on strength envelopes and internally consistent strain rates and differential stresses.

Also neglected in the simple models of this chapter are the potentially important contributions of heterogeneous material properties and thermally-induced bending. As may be seen from equations (2-4) and (7), thermal stress is proportional to the product of the coefficient of thermal expansion α and Young's modulus E [Turcotte, 1974]. Though the variation of these properties in oceanic lithosphere is not well known, available data suggest that αE in the crust may be as little as half that of the mantle [e.g., Clark, 1966]. The lithosphere would then be analogous to a bimetallic strip [Turcotte, 1974] in which thermal stress and strain are larger in the uppermost mantle than in the crust. In Turcotte's [1974] models, in which σ_{yy} is the only non-zero component of the stress tensor, brittle failure of accumulated stress can induce a bending moment in the plate which places the upper part of the lithosphere under horizontal compression. Thus dissimilar crustal and upper mantle properties may provide an alternative to a hydrothermally cooled surface layer as an explanation of shallow thrust faulting. These alternatives deserve further testing in subsequent models of the thermal stress field in young oceanic lithosphere.

CONCLUSIONS

We have presented a set of simple exploratory models to test the hypothesis that thermoelastic stress is a significant

component of the stress field in young oceanic lithosphere. The models are based on the response of an elastic half-space to cooling, including the effects in some models of a surficial layer cooled rapidly at or very near the ridge axis by penetrative hydrothermal circulation [Lister, 1977]. The stress models have been compared with several key characteristics of near-ridge earthquakes [Bergman and Solomon, 1984], including a concentration of moment release in lithosphere younger than 15 m.y. in age, a paucity of seismicity in the crust, and a rough stratification of faulting type, with thrust and strike-slip mechanisms generally confined to depths less than 12 km and normal faulting dominating at greater depth.

The thermal stress models in best agreement with the characteristics of near-ridge earthquakes include a hydrothermally-cooled layer 5 to 10 km thick and a characteristic time for release of stress that is generally less than the age of the lithosphere. The effects of stress release by fracture and flow are only crudely modeled through the inclusion of two parameters, an elastic blocking temperature above which material is presumed not to contribute to the thermal field, and a characteristic time for stress release, assumed to be constant for each model. The models of this chapter support the hypothesis that thermoelastic stress is a significant contributor to the state of stress in young oceanic lithosphere, but the details of the models depend in untested ways on the many simplifying assumptions made. Further testing of this hypothesis with more sophisticated models capable of relaxing one or more of these assumptions is warranted.

REFERENCES

- Anderson, E.M., The Dynamics of Faulting, 2nd ed., Oliver and Boyd, Edinburgh, 206 pp., 1951.
- Artyushkov, E.V., Stresses in the lithosphere caused by crustal thickness inhomogeneities, J. Geophys. Res., 78, 7675-7708, 1973.
- Bergman, E.A., and S.C. Solomon, Source mechanisms of earthquakes near mid-ocean ridges from body waveform inversion: Implications for the early evolution of oceanic lithosphere, J. Geophys. Res., in press, 1984.
- Bergman, E.A., J.L. Nabelek, and S.C. Solomon, An extensive region of off-ridge normal-faulting earthquakes in the southern Indian Ocean, J. Geophys. Res., 89, 2425-2443, 1984.
- Bott, M.H.P., and N.J. Kusznir, The origin of tectonic stress in the lithosphere, Tectonophysics, 105, 1-13, 1984.
- Brace, W.F., and D.L. Kohlstedt, Limits on lithospheric stress imposed by laboratory experiments, J. Geophys. Res., 85, 6248-6252, 1980.
- Chen, W.-P., and P. Molnar, Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res., 88, 4183-4214, 1983.
- Clark, S.P., ed., Handbook of Physical Constants, Mem. 97, 587 pp., Geological Society of America, Boulder, CO, 1966.
- Collette, B.J., Thermal contraction joints in a spreading seafloor as origin of fracture zones, Nature, 251, 299-300, 1974.

- Dahlen, F.A., Isostasy and the ambient state of stress in the oceanic lithosphere, J. Geophys. Res., 86, 7801-7807, 1981.
- Davis, E.E., and C.R.B. Lister, Fundamentals of ridge crest topography, Earth Planet. Sci. Lett., 21, 405-413, 1974.
- Epp, D., and W. Suyenaga, Thermal contraction and alteration of the oceanic crust, Geology, 6, 726-728, 1978.
- Fleitout, L., and C. Froidevaux, Tectonic stresses in the lithosphere, Tectonics, 2, 315-324, 1983.
- Forsyth, D.J., and S. Uyeda, On the relative importance of driving forces in plate motion, Geophys. J. R. Astron. Soc., 43, 163-200, 1975.
- Fung, Y.C., Continuum Mechanics, 2nd ed., Prentice-Hall, Englewood Cliffs, NJ, 340 pp., 1977.
- Gautschi, W., Error function and Fresnel integrals, in Handbook of Mathematical Functions, edited by M. Abramowitz and I.A. Stegun, pp. 295-329, National Bureau of Standards, Washington, D.C., 1964.
- Goetze, C., and B. Evans, Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics, Geophys. J. R. Astron. Soc., 59, 463-478, 1979.
- Goodier, J.N., On the integration of the thermoelastic equations, Phil. Mag., 7, 1017-1032, 1937.
- Gregory, R.T., and H.P. Taylor, Jr., An oxygen isotope profile in a section of Cretaceous oceanic crust, Samail ophiolite, Oman: Evidence for $\delta^{18}\text{O}$ buffering of the oceans by deep (> 5 km) seawater-hydrothermal circulation at mid-ocean ridges, J. Geophys. Res., 86, 2737-2755, 1981.

- Hager, B.H., and R.J. O'Connell, Kinematic models of large-scale flow in the earth's mantle, J. Geophys. Res., 84, 1031-1048, 1979.
- Hales, A.L., Gravitational sliding and continental drift, Earth Planet. Sci. Lett., 16, 31-34, 1969.
- Lewis, B.T.R., The East Pacific rise and the thermal model, J. Geophys. Res., 88, 3348-3354, 1983.
- Lilwall, R.C., T.J.G. Francis and I.T. Porter, Ocean-bottom seismograph observations on the mid-Atlantic ridge near 45°N - further results, Geophys. J. R. Astron. Soc., 55, 255-262, 1978.
- Lister, C.R.B., On the thermal balance of a mid-ocean ridge, Geophys. J. R. Astron. Soc., 26, 515-555, 1972.
- Lister, C.R.B., Qualitative models of spreading center processes, including hydrothermal penetration, Tectonophysics, 37, 203-218, 1977.
- MacMillan, W.D., The Theory of the Potential, McGraw-Hill, New York, 469 pp., 1930.
- McKenzie, D.P., Speculations on the consequences and causes of plate motion, Geophys. J. R. Astron. Soc., 18, 1-32, 1969.
- McKenzie, D.P., Plate tectonics, in The Nature of the Solid Earth, edited by E.C. Robertson, pp. 323-360, McGraw-Hill, New York, 1972.
- McNutt, M., and H.W. Menard, Constraints on yield strength in the oceanic lithosphere derived from observations of flexure, Geophys. J. R. Astron. Soc., 71, 363-394, 1982.

- Mindlin, R.D., and D.H. Cheng, Thermoelastic stress in the semi-infinite solid, J. Appl. Phys., 21, 931-933, 1950.
- Minster, J.B., and T.H. Jordan, Present-day plate motions, J. Geophys. Res., 83, 5331-5354, 1978.
- Parsons, B., and J.G. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res., 82, 803-827, 1977.
- Richardson, R.M., S.C. Solomon, and N.H. Sleep, Tectonic stress in the plates, Rev. Geophys. Space Phys., 17, 981-1019, 1979.
- Richter, F.M., and B. Parsons, On the interaction of two scales of convection in the mantle, J. Geophys. Res., 80, 2529-2541, 1975.
- Sandwell, D., and G. Schubert, Lithospheric flexure at fracture zones, J. Geophys. Res., 87, 4657-4667, 1982.
- Smith, A.T., and M.N. Toksoz, Stress distribution beneath island arcs, Geophys. J. R. Astron. Soc., 29, 289-318, 1972.
- Solomon, S.C., N.H. Sleep, and R.M. Richardson, On the forces driving plate tectonics: Inferences from absolute plate velocities and intraplate stress, Geophys. J. R. Astron. Soc., 42, 769-801, 1975.
- Stein, S., and E.A. Okal, Seismicity and tectonics of the Ninety-east Ridge area: Evidence for internal deformation of the Indian plate, J. Geophys. Res., 83, 2233-2245, 1978.
- Sykes, L.R., and M.L. Sbar, Focal mechanism solutions of intraplate earthquakes and stresses in the lithosphere, in

- Geodynamics of Iceland and the North Atlantic Area, edited by L. Kristjansson, pp. 207-224, D. Reidel, Hingham, MA, 1974.
- Timoshenko, S.P., and J.N. Goodier, Theory of Elasticity, 3rd ed., McGraw-Hill, New York, 567 pp., 1970.
- Toomey, D.R., S.C. Solomon, G.M. Purdy, and M.H. Murray, Microearthquakes beneath the median valley of the Mid-Atlantic Ridge near 23°N: Hypocenters and focal mechanisms, J. Geophys. Res., in press, 1984.
- Turcotte, D.L., Are transform faults thermal contraction cracks?, J. Geophys. Res., 79, 2573-2577, 1974.
- Turcotte, D.L., Thermal stresses in planetary elastic lithospheres, Proc. 13th Lunar Planet. Sci. Conf., Part 2, J. Geophys. Res., 88, Suppl., A585-A587, 1983.
- Turcotte, D.L., and E.R. Oxburgh, Finite amplitude convective cells and continental drift, J. Fluid Mech., 23, 29-42, 1967.
- Turcotte, D.L., and E.R. Oxburgh, Mid-plate tectonics, Nature, 244, 337-339, 1973.
- Turcotte, D.L., and G.C. Schubert, Structure of the olivine-spinel phase boundary in the descending lithosphere, J. Geophys. Res., 76, 7980-7987, 1971.
- Walcott, R.I., Flexure of the lithosphere at Hawaii, Tectonophysics, 9, 435-446, 1970.
- Watts, A.B., and J.R. Cochran, Gravity anomalies and flexure of the lithosphere along the Hawaiian-Emperor seamount chain, Geophys. J. R. Astron. Soc., 38, 119-141, 1974.

Weissel, J.K., R.N. Anderson, and C.A. Geller, Deformation of the Indo-Australian plate, Nature, 287, 284-291, 1980.

Wiens, D.A., and S. Stein, Age dependence of oceanic intraplate seismicity and implications for lithospheric evolution, J. Geophys. Res., 88, 6455-6468, 1983.

Wiens, D.A., and S. Stein, Intraplate seismicity and stresses in young oceanic lithosphere, J. Geophys. Res., in press, 1984.

FIGURE CAPTIONS

- Figure 1. Centroid depth and focal mechanism type versus lithosphere age for 32 near-ridge earthquakes [Bergman and Solomon, 1984]. Depths are relative to the seafloor. Earthquakes are further distinguished on the basis of the ocean in which they occur (A = Atlantic, I = Indian, P = Pacific). Mechanism symbols are superposed for earthquakes showing a mixture of faulting types. Also shown are selected isotherms from the half-space cooling model of equation (1).
- Figure 2. Geometry of the thermoelastic stress problem for cooling oceanic lithosphere; see text.
- Figure 3. (a) Temperature change ($-\Delta T$ from equation 9) in a cooling halfspace over a time increment Δt of 4 m.y. (b) Same as (a) except that temperatures above an elastic "blocking temperature" of 550°C do not contribute to ΔT (see text).
- Figure 4. Thermal stress in an elastic halfspace due to the cooling of an elongated box 6 km by 6 km in cross section. The box is centered at $x=0$, at $z = 9$ km and has x , y and z half-dimensions $a=3$, $b=10,000$, and $c=3$ km, respectively. The initial temperature within the box is 10°C , while that of the rest of the halfspace and of the surface is 0°C ; the thermal stress shown is at large t after all cooling is complete. The orientations of the greatest (σ_1) and

least (σ_3) compressive deviatoric stresses are indicated by arrows. Also shown are contours, in bars, of the differential stress $\Delta\sigma = \sigma_1 - \sigma_3$. Zones of predicted thrust (TF) and normal faulting (NF) are outlined by dashed lines.

Figure 5. Predicted thermal stress for model A, including only simple half-space cooling. The model is symmetric about $x = 0$. The orientations of greatest (σ_1) and least (σ_3) compressive stress are indicated; three directions are shown when two of the principal stresses are comparable in magnitude. Also shown are contours of the differential stress $\Delta\sigma$ in kbar; a characteristic time scale Δt of 4 m.y. for release of accumulated thermal stress has been assumed. Zones of predicted thrust (TF), strike-slip (SS), and normal faulting (NF) are outlined by dashed lines.

Figure 6. Predicted thermal stress for model B. The model includes a 10 km thick cooled layer ($T=0^\circ\text{C}$) to simulate hydrothermal circulation. See Figure 5 for further explanation.

Figure 7. Predicted thermal stress for model C. An elastic blocking temperature $T_e = 550^\circ\text{C}$ has been assumed (see text). No hydrothermally cooled layer is present. See Figure 5 for further explanation.

Figure 8. Predicted thermal stress for model D. The model includes an elastic blocking temperature $T_e = 550^\circ\text{C}$

and a hydrothermally cooled layer 10 km in thickness. See Figure 5 for further explanation.

Figure 9. Predicted thermal stress for model E. The model is similar to D except that $\Delta t = 10$ m.y. See Figure 5 for further explanation.

Figure 10. Stress field predicted from the ridge-push model of Dahlen [1981]. Contours represent lines of constant stress difference $\Delta\sigma = \sigma_{xx} - \sigma_{zz}$. The principal deviatoric stresses σ_{xx} and σ_{zz} in this model have magnitudes equal to $\Delta\sigma/2$ and are opposite in sign.

Figure 11. Predicted stress field for model F. The model is a superposition of the thermal stress from model D and the ridge-push stress field from Figure 10. See Figures 5 and 10 for further explanation.

Table 1. Adopted values of parameters used in thermal stress models [Parsons and Sclater, 1977].

Variable	Description	Value
k	thermal diffusivity	$8 \times 10^{-3} \text{ cm}^2/\text{s}$
T_L	temperature at ridge crest and beneath the lithosphere	1350°C
v	spreading half-rate	1 cm/yr
α	volumetric coefficient of thermal expansion	$3.2 \times 10^{-5}/^\circ\text{C}$
E	Young's modulus	$1.7 \times 10^{12} \text{ dyne/cm}^2$
ν	Poisson's ratio	0.25
g	gravitational acceleration	980 cm/s^2
ρ_m	density of uncooled lithosphere	3.33 g/cm^3

Table 2. Models of thermal stress in young oceanic
lithosphere

Model Number	Hydrothermal Circulation?	Δt , m.y.	T_e , °C	Ridge Push?	Figure Number
A	No	4	T_L	No	5
B	Yes	4	T_L	No	6
C	No	4	550	No	7
D	Yes	4	550	No	8
E	Yes	10	550	No	9
F	Yes	4	550	Yes	11

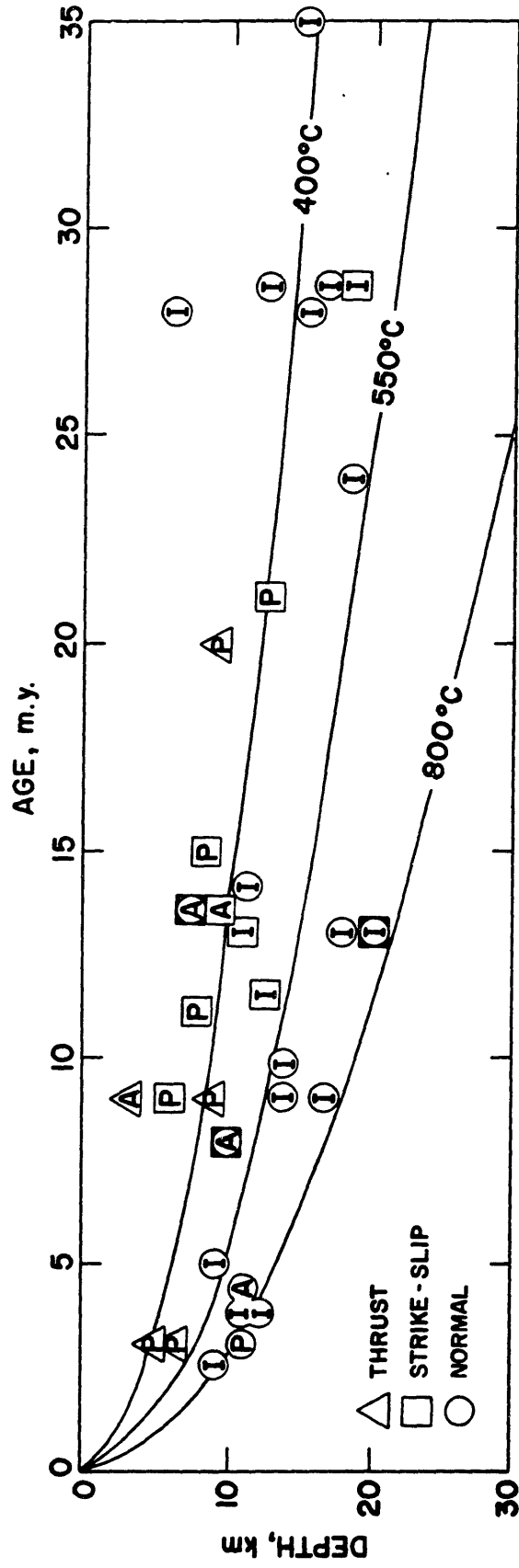


Figure 1

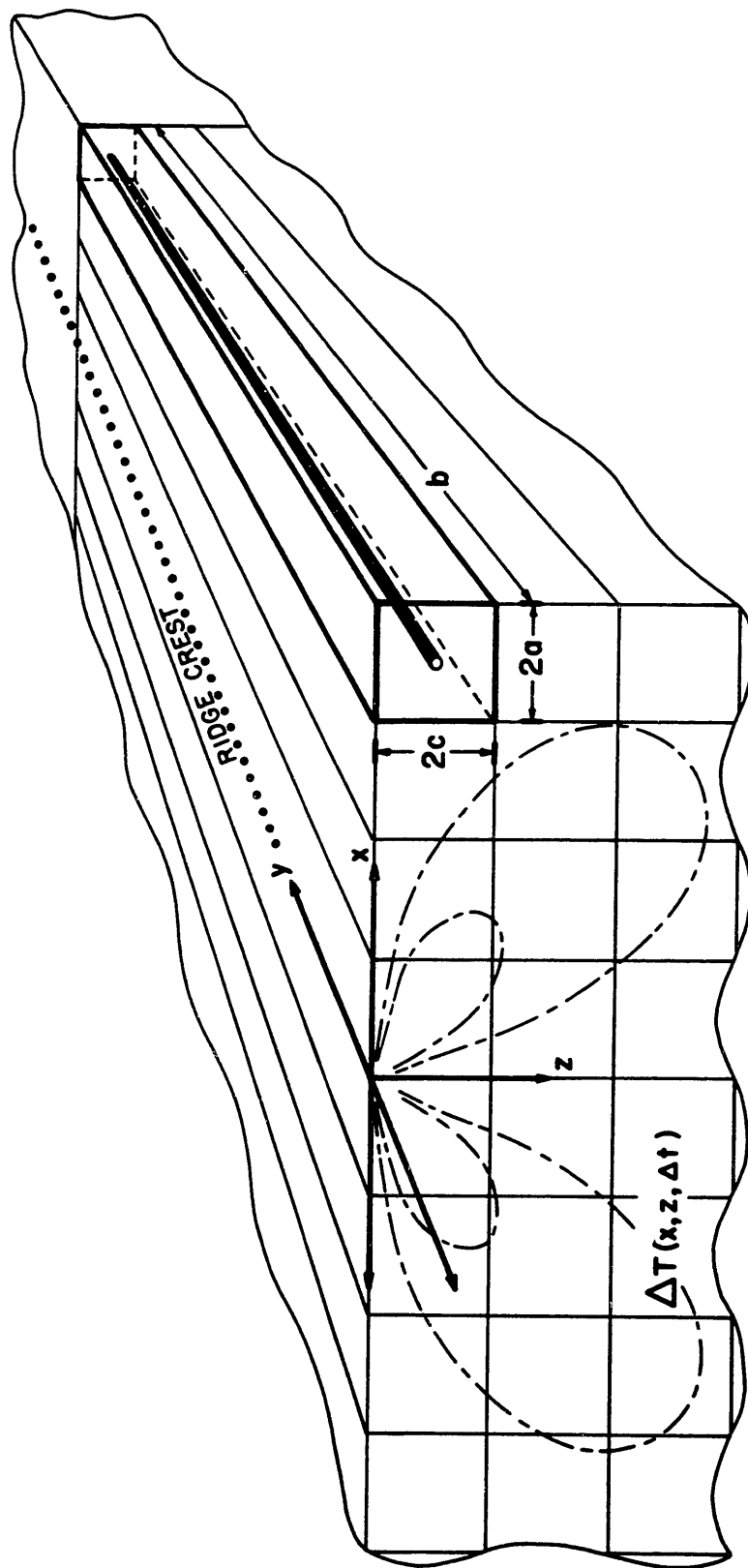


Figure 2

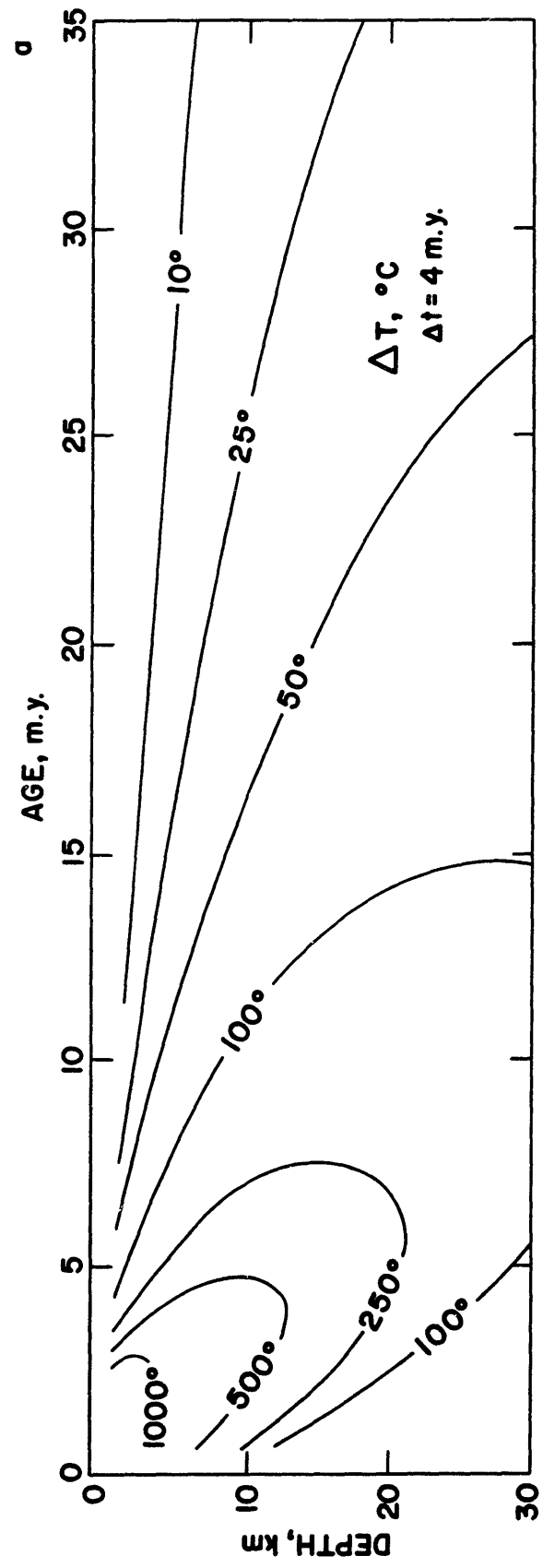


Figure 3a

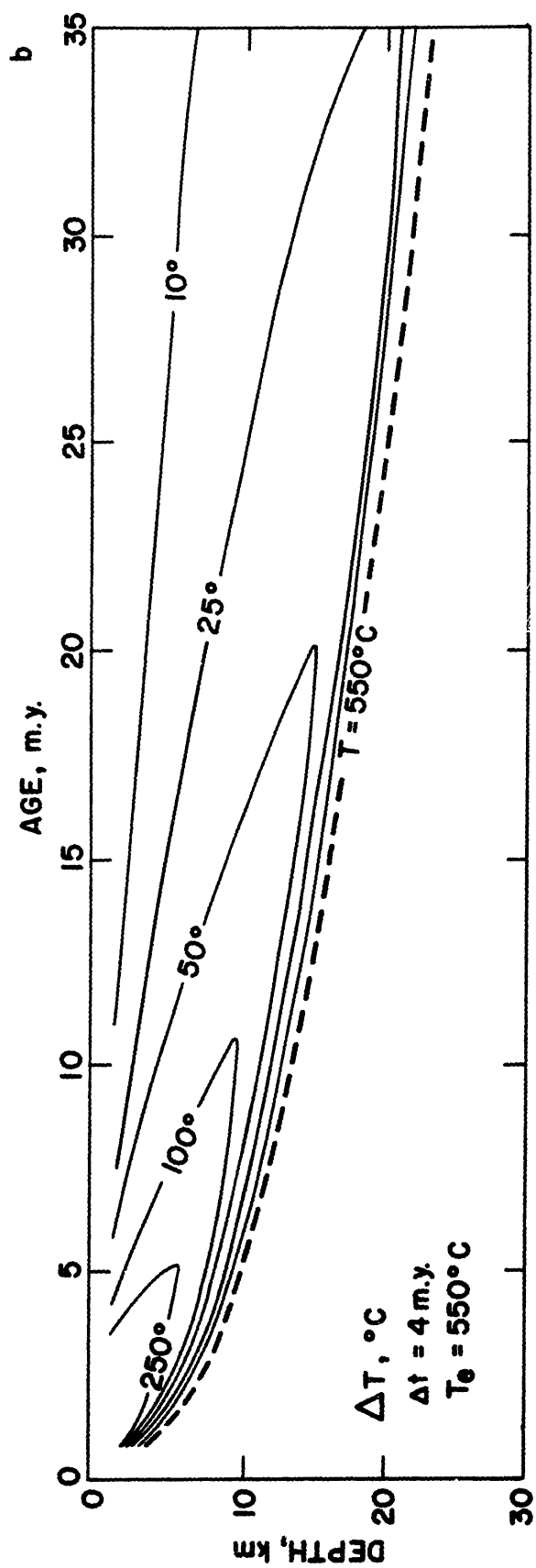


Figure 3b

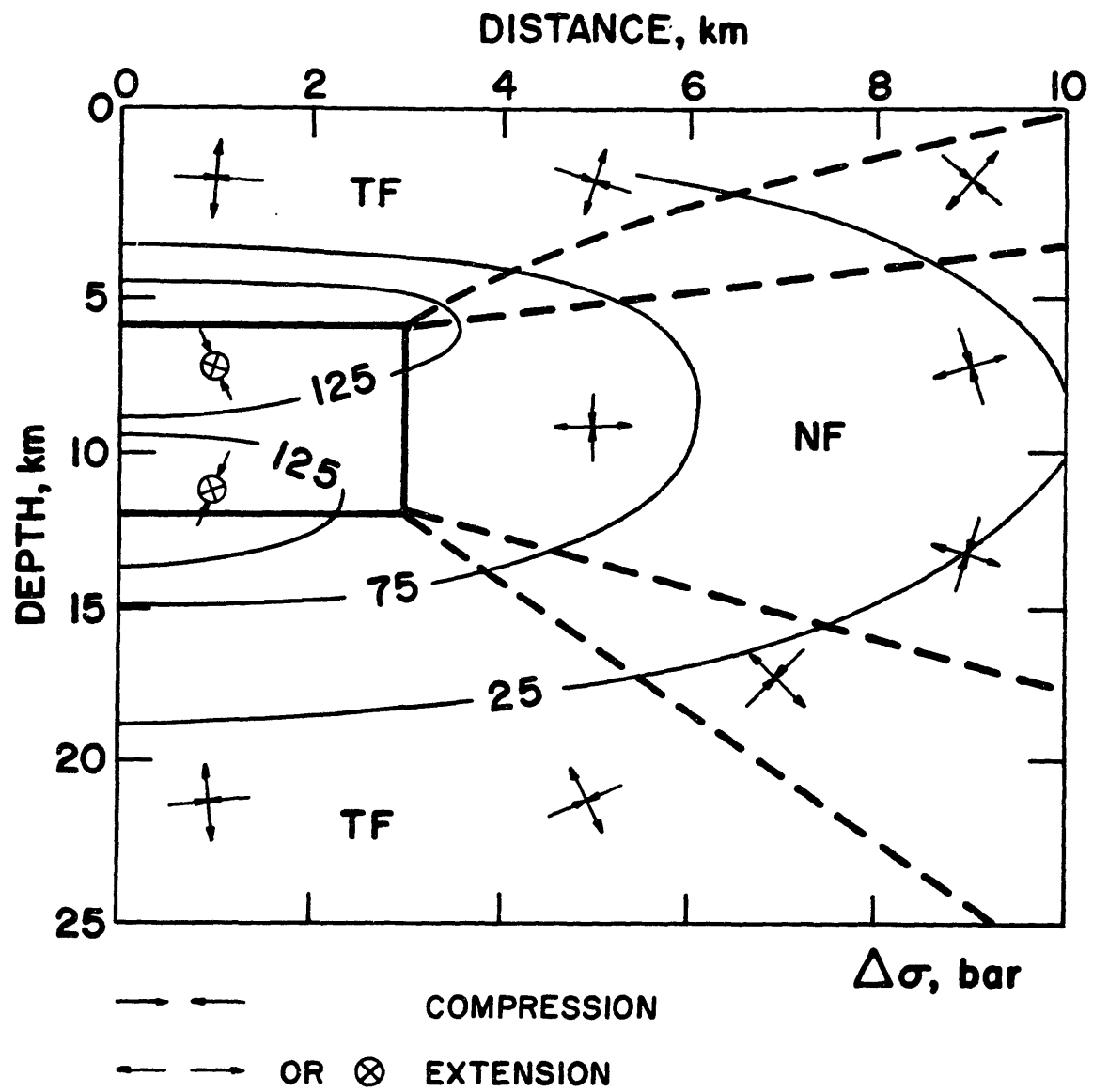


Figure 4

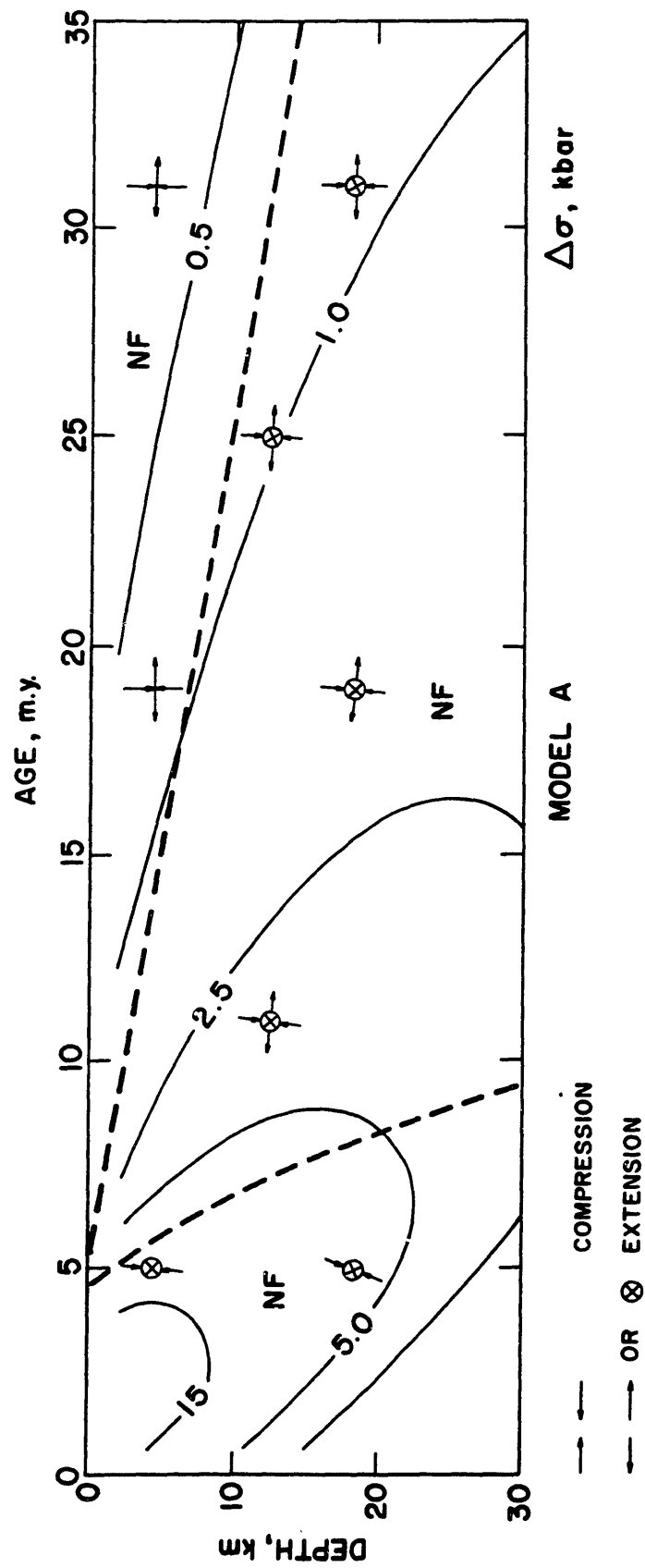


Figure 5

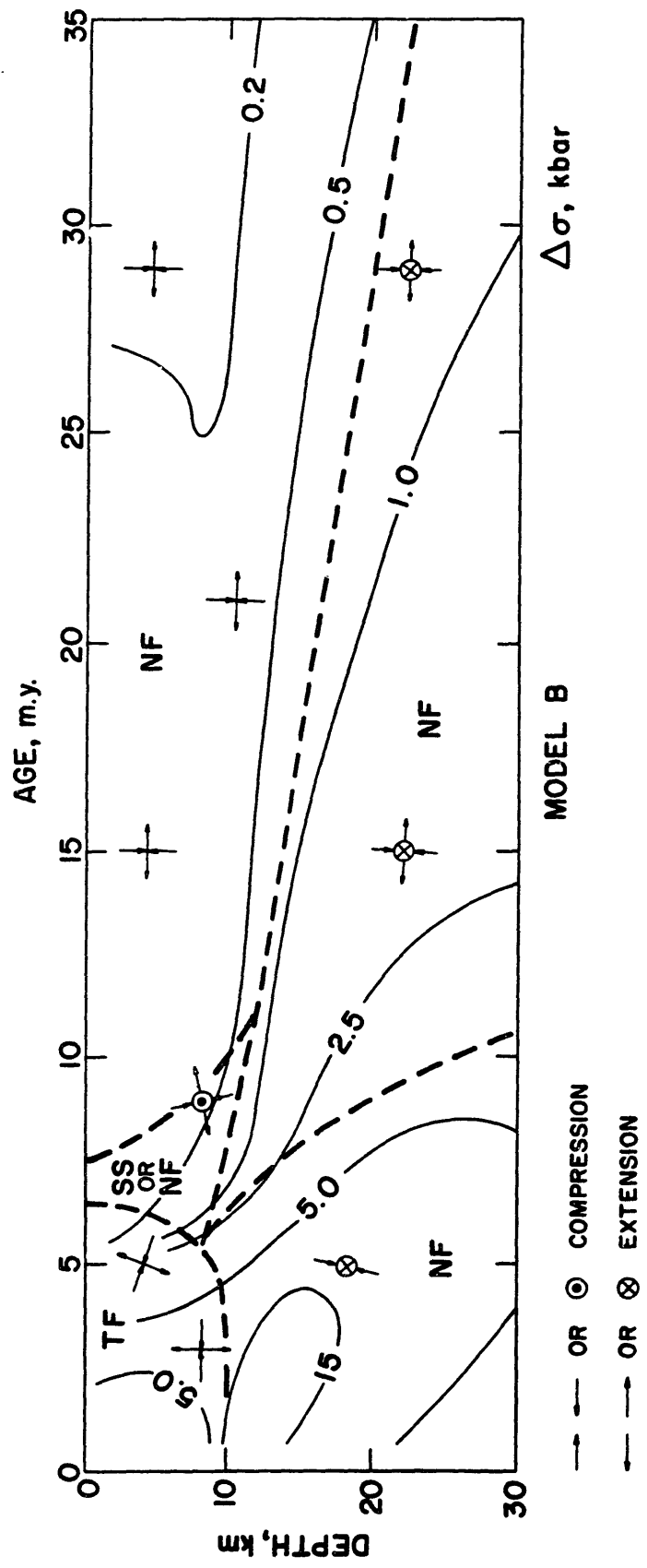


Figure 6

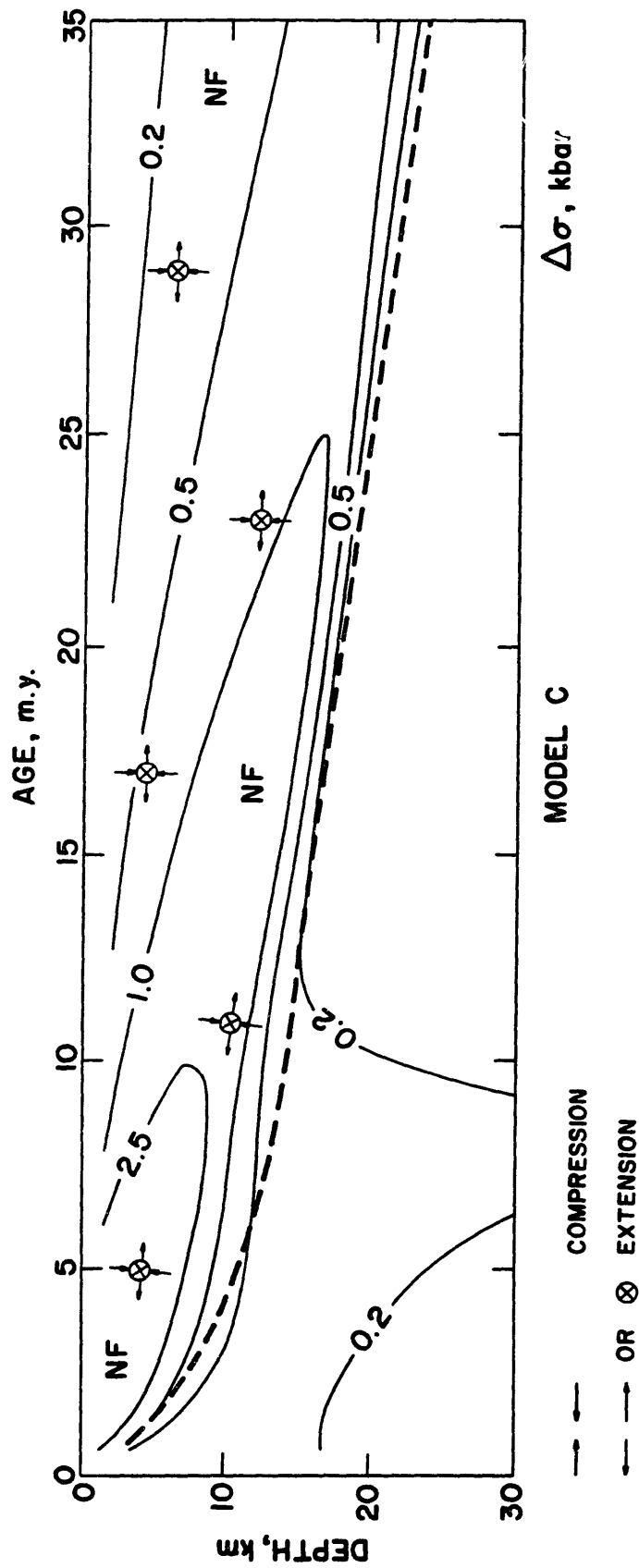


Figure 7

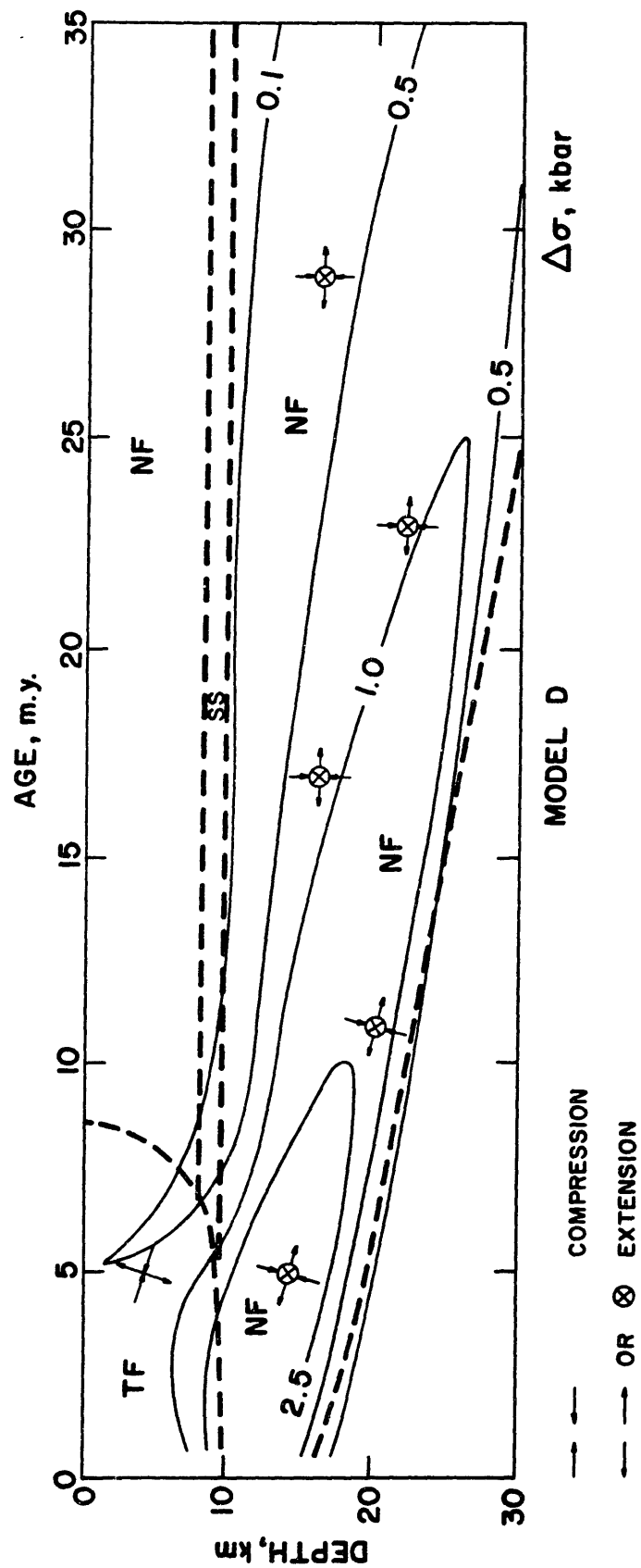


Figure 8

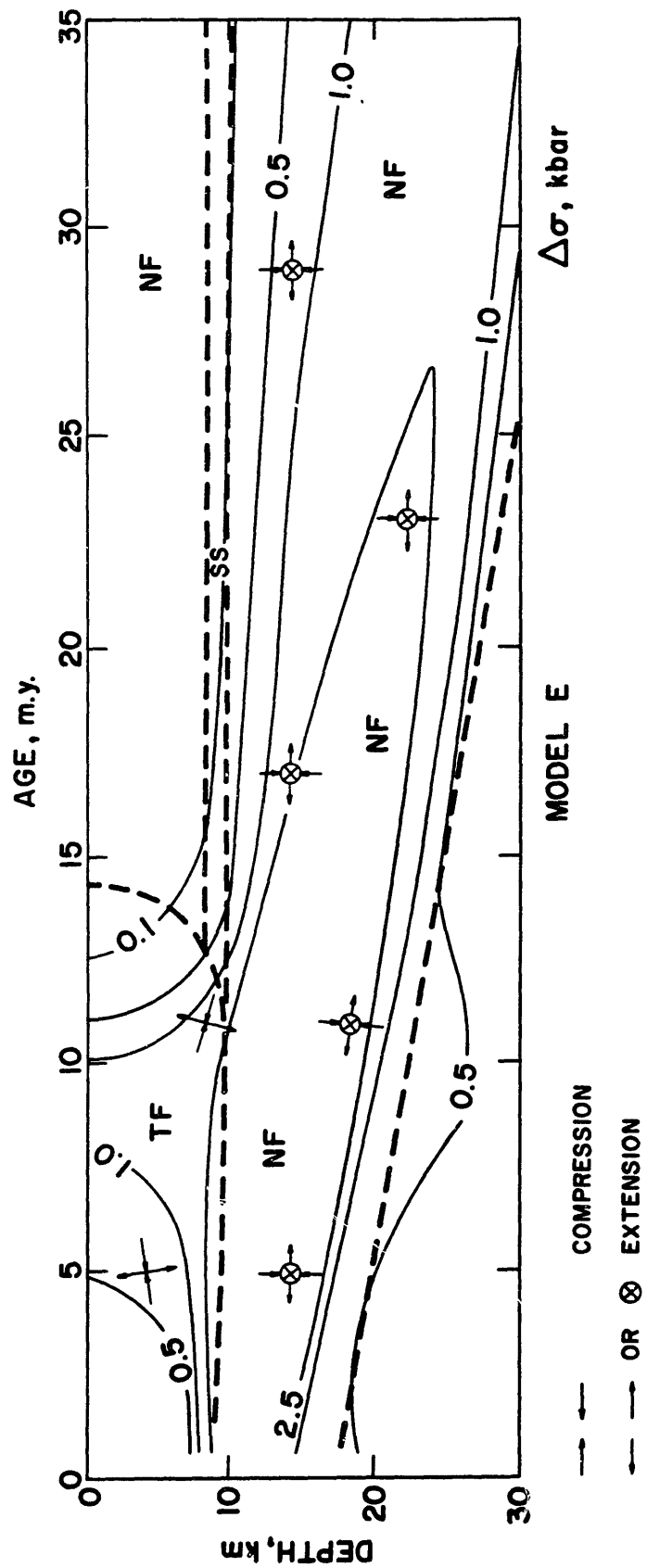


Figure 9

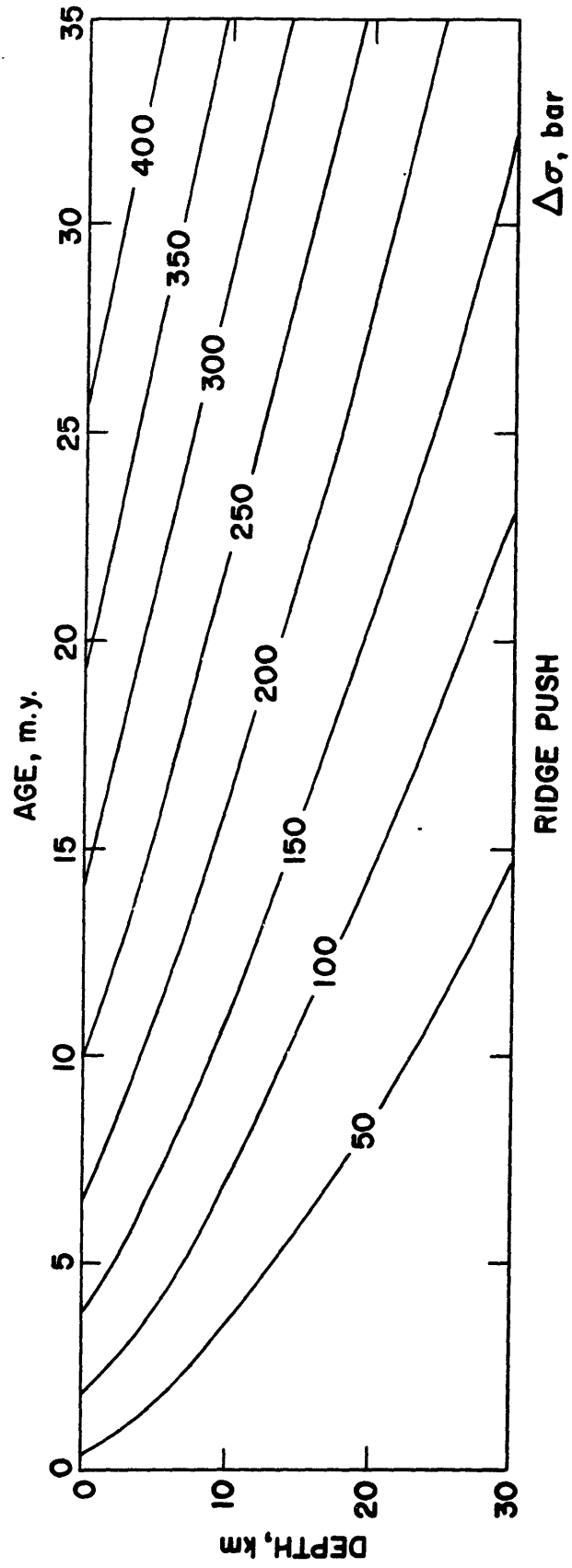


Figure 10

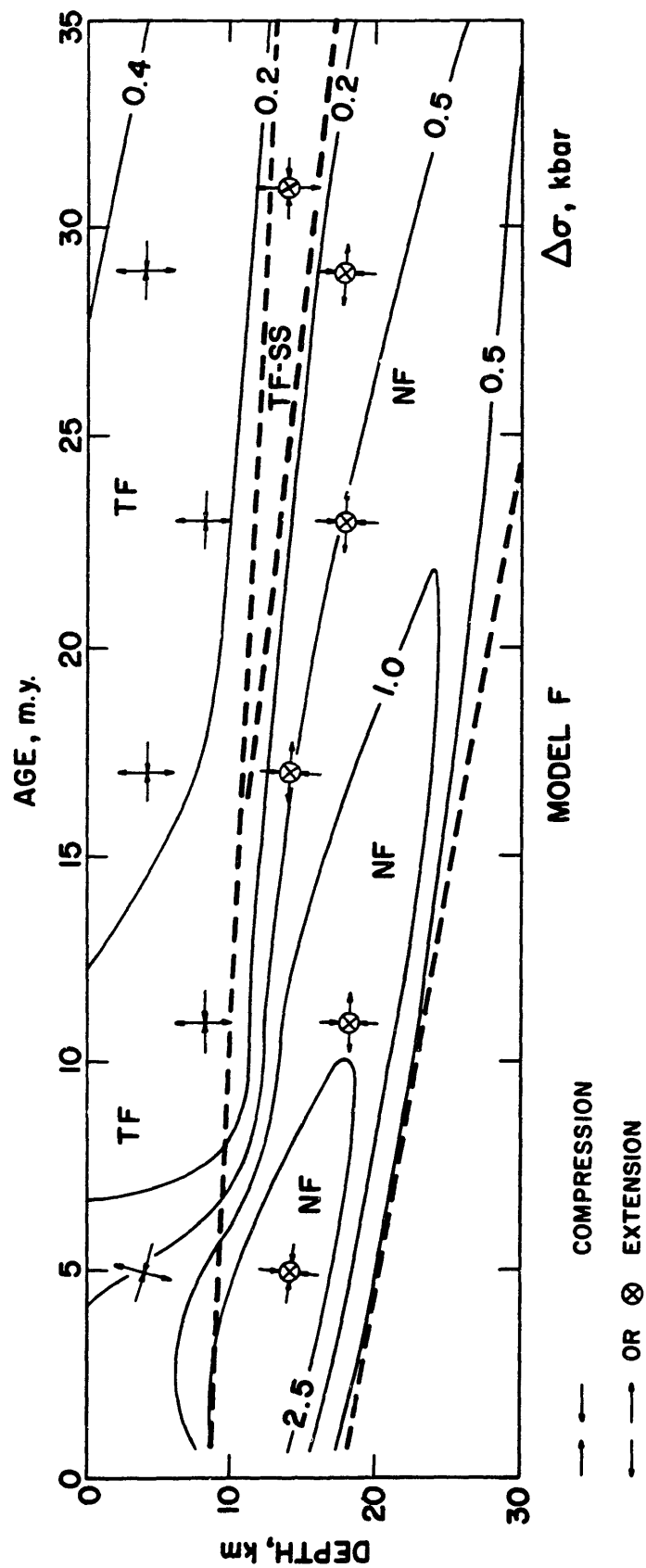


Figure 11

Chapter 5

THE DEEP STRUCTURE OF LUNAR BASINS: IMPLICATIONS FOR BASIN FORMATION AND MODIFICATION

INTRODUCTION

The formation of multi-ring impact basins has played a major role in the geological evolution of the Moon. During the first billion years of lunar history, the impact of large projectiles onto the lunar surface resulted in the excavation of basin cavities hundreds of kilometers in diameter [Wood and Head, 1976] and the implantation of large quantities of heat into the lunar interior [O'Keefe and Ahrens, 1977]. Impact basins also became the focus for volcanic and tectonic activity over a considerable time period following the basin formation events [e.g., Head, 1976; Solomon and Head, 1979; Solomon et al., 1982]. Important constraints on the processes of basin formation and modification are provided by the present volumes of the topographic basin, of material ejected during basin formation, and of mare basalt fill, as well as by the degree of involvement of the mantle in isostatic compensation of basin relief. The topographic volumes of the basins are reasonably well known. Estimates for the volumes of mare basalt and basin ejecta deposits have also been obtained from photogeological studies [e.g., Moore et al., 1974; Head et al., 1975; DeHon and Waskom, 1976; Head, 1982], but such techniques are generally limited in their ability to resolve the thickness of the deposits. The depth of excavation of several basin-forming events on the Moon has also been estimated from the chemistry and mineralogy of ejecta deposits inferred from remote sensing data [Spudis, 1982, 1983; Spudis et al., 1984], but such estimates depend critically on the accurate identification of

primary ejecta and on assumptions about chemical layering of the lunar crust. In this chapter, we apply gravity and topographic data to infer the three-dimensional structure of the crust and upper mantle beneath impact basins on the lunar nearside. With the derived structural models, we constrain several of the important geometrical and physical parameters related to the processes of basin formation and modification and we assess their variations with basin age and size.

Muller and Sjogren [1968] were the first to recognize that the youngest nearside mare basins are characterized by positive gravity anomalies, which they attributed to "mascons." Since that discovery, a number of efforts have been made to model the gravity anomalies over mascon basins with contributions to the anomalous mass placed at the surface [Conel and Holston, 1968; Baldwin, 1968; Booker et al., 1970], at the lunar Moho [e.g., Wise and Yates, 1970], or at both locations [e.g., Hulme, 1972; Wood, 1972; Bowin et al., 1975; Sjogren and Smith, 1976]. At least some portion of the anomalous mass contributing to mascon anomalies resides near the surface [Phillips et al., 1972], but models where mare basalt fill is the sole source of anomalous mass require an unreasonable thickness of mare basalt to fit the measured gravity field [Thurber and Solomon, 1978]. While both mare fill and an elevated Moho likely contribute to the observed gravity, the solution for the distribution of anomalous mass between the two locations given only gravity and topographic data requires additional assumptions.

Structural models consistent with gravity and topographic

data have been constructed for a number of individual basins [e.g., Bowin et al., 1975; Sjogren and Smith, 1976; Phillips and Dvorak, 1981; Janle, 1981a,b]. Most of these models, however, were developed under dissimilar sets of assumptions and constraints, thus hindering a comparison of the inferred structures beneath different basins. Also, the interpretation of gravity data over a single basin requires that the investigator make subjective judgements about regional trends in the gravity data arising from structures outside the area of interest or occurring over wavelengths greater than the scale of the basin. In contrast, global models for lunar crustal structure [Wood, 1973; Bills and Ferrari, 1977a; Thurber and Solomon, 1978] calculated under uniform sets of constraints and assumptions permit an internally consistent assessment of structural variability on a regional scale. These global models were developed to address crustal structure at scales greater than the dimensions of most lunar basins, however, and with the exception of the study of Thurber and Solomon [1978], none considered mare basalt as a significant contributor to the observed gravity field.

In this chapter we determine models for the crustal structure in the vicinity of 9 impact basins on the lunar nearside (Figure 1). The models are derived as part of a simultaneous inversion of nearside gravity and topographic data, using a procedure similar to that of Thurber and Solomon [1978]. The models include a low-density non-mare crustal layer and a mare basalt layer, both of variable thickness. The contributions

of Moho relief and mare basalt to the observed gravity anomalies are separated with the assumptions that basin topography was isostatically compensated prior to mare basalt fill and that crustal subsidence in response to loading by mare basalts may be neglected. These assumptions lead to minimum values for mare basalt thicknesses [Thurber and Solomon, 1978] but, because of a similarity in density of mare basalts and mantle material, the thickness of the low-density non-mare crust may be reliably estimated. An important constraint is that the thickness of the non-mare crust in the vicinity of the Apollo 12 and 14 landing sites is known from seismic refraction measurements [Toksoz et al., 1974]. On the basis of the nearside crustal model, we compare the structure beneath the largest lunar basins, we derive new bounds on the volume of material ejected from each basin, and we evaluate the implications of structural differences among basins for the processes of basin formation and modification as functions of time on the Moon.

PROCEDURE

To determine the crustal structure in the vicinity of lunar basins, we perform a simultaneous inversion of gravity and topography for the low-latitude portion of the lunar nearside, following a procedure similar to that of Thurber and Solomon [1978]. The Moon is divided into a grid of blocks, each 5° by 5° in horizontal extent (Figure 2). These block dimensions represent the approximate limits of resolution of the available gravity data, as discussed further below. The disturbing gravitational potential at any point over the nearside can

be approximated as the sum of the potentials due to the distribution of anomalous mass within each block. In this section we describe the adopted procedure for calculation of gravity anomalies, as a forward problem, given a distribution of topography or anomalous mass that is uniform within each block. We then discuss the constraining assumptions we have chosen to make the inverse problem well posed. Finally, we describe an iterative, linearized inversion procedure to determine the crustal structure within each block from gravity and topography, subject to the adopted constraints.

Computation of Gravity Anomalies

In previous analyses of the gravity anomaly fields of planets using spherical shell segments [Morrison, 1976; Thurber and Solomon, 1978], contributions to the anomalous mass within each segment have been approximated by uniform surface masses located at a constant radius from the center of the planet. While this method simplifies the computation of the gravitational potential, it may provide a poor approximation to the actual potential when the thickness of the block is similar in magnitude to the distance between the block and the observation point.

As an improved approximation, we represent contributions to the anomalous mass within a spherical shell segment as arising from a finite volume of thickness b_r and of latitudinal and longitudinal extent b_θ and b_ϕ , respectively. In general, the disturbing potential at a point $\underline{r}$ above the lunar surface due to a block of anomalous mass within a spherical shell segment is

$$u(\underline{r}) = \iiint_{\substack{\text{volume} \\ \text{of block}}} F \, d\theta' \, d\phi' \, dr' \quad (1)$$

where

$$F = G \rho(\underline{r}') \frac{(r')^2 \cos \theta'}{|\underline{r} - \underline{r}'|} \quad (2)$$

and G is the gravitational constant, ρ is the density (or density contrast), θ is the latitude, ϕ is the longitude, and primed variables denote coordinates of anomalous mass within the block. All radius vectors are with respect to the lunar center of mass. If F is expanded to second order in a Taylor series about the center of the block, equation (1) simplifies to

$$u(\underline{r}) = b_\theta \, b_\phi \, b_r \left[F + \frac{1}{24} (F_{\theta\theta} \, b_\theta^2 + F_{\phi\phi} \, b_\phi^2 + F_{rr} \, b_r^2) \right] \quad (3)$$

where the double subscripts designate second derivatives of F with respect to the primed coordinates evaluated at the center of each block.

The gravitational potential due to the anomalous mass within all M blocks in the grid is obtained by summation:

$$U(\underline{r}) = \sum_{j=1}^M u_j(\underline{r}) \quad (4)$$

and the free-air gravity anomaly field $g(\underline{r})$ is obtained by the simple finite-difference relation:

$$g(\underline{r}) = \frac{U(\underline{r} + \Delta r \, \underline{r}/r) - U(\underline{r})}{\Delta r} \quad (5)$$

Equation (3) is used to compute the contributions from blocks beneath and near the observation point. For a block more than

900 km distant from the observation point, the distribution of anomalous mass in the right-hand side of (1) for that block is represented by 4 point masses; for a block more than 1200 km distant, a single point mass is used. These simple approximations provide accuracy comparable to that of equation (3) with a considerable savings of computation time.

Constraints and Assumptions

Following Thurber and Solomon [1978] we assume that Airy compensation dominates on the Moon and that there are three principal contributions from each block to the anomalous gravity field (Figure 2): (1) surface topography, measured relative to a datum D_T and contributing to gravity in proportion to the density ρ_c of the upper crust; (2) Moho relief, measured relative to a datum D_M and contributing to gravity in proportion to the density contrast $\Delta\rho_m = \rho_m - \rho_c$ between the densities ρ_c and ρ_m of the crust and mantle, respectively; and (3) mare basalt fill with a density contrast $\Delta\rho_b = \rho_b - \rho_c$. We adopt the values $\rho_c = 2.9 \text{ g/cm}^3$, $\rho_m = 3.4 \text{ g/cm}^3$, and $\rho_b = 3.4 \text{ g/cm}^3$, estimates that are uncertain by about 0.1 g/cm^3 [Solomon, 1975]. Thus $\Delta\rho_b = \Delta\rho_m = 0.5 \text{ g/cm}^3 \equiv \Delta\rho$ in this problem. The contribution of topography is evaluated using equations (3-5) and is subtracted from the free-air anomaly field to yield the Bouguer anomaly field.

To invert formally the Bouguer anomaly field for crustal structure, it is necessary to impose two additional constraints [Thurber and Solomon, 1978]. The first constraint is that, by assumption, (i) topography in mare areas was isostatically

compensated by an Airy mechanism before the emplacement of mare basalts and (ii) no compensation of mare units has occurred since their emplacement. These two assumptions, for which we use the term "pre-mare isostasy constraint" below, permit a unique decomposition of the gravity anomaly into contributions from Moho relief and mare fill. The first assumption receives support from the argument that the crust surrounding a recently formed impact basin was likely to have been hot and incapable of supporting the large deviatoric stresses associated with a deep basin depression [e.g., Bratt et al., 1981; Chapter 6]. The second assumption is not strictly correct; mare ridges and linear rilles provide evidence that the lunar lithosphere has subsided in response to the mare basalt load in the mascon maria [Solomon and Head, 1979, 1980]. This second assumption, however, has little influence on the estimated thickness of non-mare crustal material. The effects of both assumptions on the derived structural models are discussed further below. Under the pre-mare isostasy constraint, the basalt thickness t_j and the depth to the Moho d_j within any mare block j are related by the equation:

$$(t_j + h_j)\rho_c = (D_I - d_j)(\rho_m - \rho_c) \quad (6)$$

where h_j is the topography. Both t_j and d_j are measured relative to the surface of a standard section which is assumed to be in isostatic equilibrium and which has a known depth to Moho D_I . This relationship is illustrated in Figure 3.

The second constraint on the derived crustal models follows

from the fact that the crust in the general vicinity of the Apollo 12 and 14 landing sites ($\sim 3^\circ\text{S}$, 20°W) is known from seismic refraction measurements to be 55 km thick [Toksoz et al., 1974]. Because the Apollo 12 and 14 sites are located in regions of ancient pre-mare structure and thin mare fill in the Oceanus Procellarum-Mare Cognitum area, it is reasonable to assume that the degree of isostasy in the region is nearly complete. We therefore adopt a 55-km thick layer of non-mare crustal material of density ρ_c as the standard section for equation (6). With this assumption, the model for the nearside crustal structure developed in this study automatically matches the crustal thickness in the vicinity of the Apollo 12 and 14 landing sites.

Inversion for Crustal Structure

To determine the Moho configuration and thickness of mare basalt on the lunar nearside, we invert the Bouguer anomaly data, subject to the adopted assumptions and constraints. The inversion scheme iteratively improves on an initial estimate of the crustal structure. The criterion used to determine a best-fit model is the minimization of the root mean square residual gravity anomaly:

$$\text{r.m.s. residual} = \sqrt{\frac{\sum_{k=1}^N (g_k^{\text{obs}} - g_k^{\text{calc}})^2}{N}} \quad (7)$$

where N is the number of gravity anomaly observations, g_k^{obs} is the k th observed gravity anomaly, and g_k^{calc} is the k th gravity anomaly calculated from the block model using the non-linear formulations given in equations (3-5).

At each iteration in the inversion scheme, equations (3-5) are approximated by a linear relation between the Bouguer gravity anomaly and the thickness of anomalous mass within a block:

$$g_k^{\text{calc}} = -G \Delta \rho \sum_{j=1}^M \frac{A_j b_j \cos \alpha_{kj}}{r_{kj}^2} \quad k = 1, 2, \dots, N \quad (8)$$

where b_j is the total thickness of the anomalous mass (Moho uplift plus mare fill) in the j^{th} block, $\Delta \rho$ is the density contrast between the block and the surrounding crust (0.5 g/cm^3), A_j is the surface area of the j^{th} block, r_{kj} is the vector between the k^{th} observation point and the center of mass of all anomalous mass in the j^{th} block, and α_{kj} is the angle at the observation point between r_{kj} and the downward vertical. From the linear relation (8) between g_k^{calc} and b_j , it follows that $\Delta g_k = g_k^{\text{calc}} - g_k^{\text{obs}}$ can be linked to perturbations in the thickness of anomalous mass by the relationship:

$$\Delta g_k = \sum_{j=1}^M \frac{\partial g_k^{\text{calc}}}{\partial b_j} \Delta b_j \quad (9)$$

or

$$\underline{\Delta g} = \underline{P} \underline{\Delta b} \quad (10)$$

where $\underline{\Delta g}$ is an $N \times 1$ column vector of residual gravity anomalies, $\underline{P}$ is an $N \times M$ matrix of partial derivatives ($P_{kj} = -G \Delta \rho A_j \cos \alpha_{kj} / r_{kj}^2$), and $\underline{\Delta b}$ is an $M \times 1$ vector of thickness corrections to the anomalous mass.

For $N > M$, equation (10) can be treated as a least squares problem and inverted by any number of methods [e.g., Lawson and Hanson, 1974] to obtain the corrections to the thickness of

anomalous mass in each block. These corrections are then added to the mass thicknesses in the crustal model from the prior iteration, subject to the constraint of pre-mare isostasy, imposed where applicable, to separate the contributions from Moho relief and mare fill. After each iteration, the global datum (D_M) for Moho relief is adjusted to meet the constraint on crustal thickness inferred from seismic measurements for the region of the Apollo 12 and 14 landing sites. Finally, a new Bouguer anomaly field g^{calc} is computed from the adjusted model using equations (3-5), the r.m.s. residual gravity anomaly is determined from (7), and the inversion process is repeated until the r.m.s. residual converges to a minimum.

We impose one additional constraint on blocks in mare regions. If the Bouguer anomaly over a mare block is sufficiently small, the pre-mare isostasy constraint may not be consistent with a finite thickness of basalt. Therefore, if the adjustment to the basalt thickness within a mare block forces the total mare thickness to fall below 250 m, the constraint of pre-mare isostasy for that block is relaxed and the basalt thickness is held constant at 250 m for the remaining iterations.

A substantial savings in computation time and core storage was achieved by filling the matrix $\underline{P}$ only with those partial derivatives associated with the block directly below the observation point and the 8 surrounding blocks. This approximation is equivalent to assuming that the residual anomaly above a given block is caused only by errors in the adopted thicknesses of the anomalous mass in the 9 nearest blocks. The resulting partial derivative matrix is then sparse, and can be easily

manipulated into a diagonally dominant form. We then solve for the vector of thickness corrections (Δb) by upper-triangularizing P using a series of Householder transformations applied to sequentially accumulated rows of P [Lawson and Hanson, 1974, pp. 207-311].

GRAVITY AND TOPOGRAPHIC DATA

The inversion procedure described above to determine crustal and upper mantle structure would be straightforward if the free-air anomaly and topography were known everywhere on the moon within a well-prescribed uncertainty. Such, unfortunately, is not the case. Information on the lunar gravity field is derived from measurements of Doppler shifts in the frequency of radio transmissions along a line-of-sight (LOS) direction between the Earth and a satellite in orbit about the Moon. High-frequency variations in these Doppler observations contain a signature of gravity anomalies arising from near-surface heterogeneities in density. To extract this signature, the effect of orbital parameters on the LOS Doppler data must be removed or modeled. The estimation of a representation of the free-air gravity field over a significant area of the lunar surface requires the simultaneous inversion of Doppler data from a large number of orbits for the parameters of both the spacecraft orbits and the gravity field. The first such representation for a large fraction of the lunar nearside was presented by Wong et al. [1971]. They presented both point-mass and disk-mass representations of lunar gravity from Lunar Orbiter and early Apollo tracking data, as well as a full discussion of the

procedures used in data inversion.

For this study, we employ an improved representation of the low-latitude nearside gravity field obtained by Wong et al. [1975] from an inversion of tracking data from low-altitude Apollo spacecraft. Their representation consists of 350 near-surface disk masses distributed between $\pm 30^\circ$ latitude and $\pm 100^\circ$ longitude and superimposed on the triaxial model for lunar gravity of Liu and Lang [1971]. Parameters for each of the mass disks and for the triaxial gravity model are given in Appendix A. The location and size of each disk were assigned a priori, either on the basis of geological information or in an otherwise regular spacing, generally 5° in latitude and longitude. This spacing is a measure of the horizontal resolution of the gravity field model. Disk radii range from 75 to 300 km; the largest disks represent major impact basins. The mass of each disk was derived from the simultaneous inversion of a large quantity of Doppler tracking data from Apollo 14, 15 and 16 spacecraft. Orbits used in the inversion ranged from 15 to 200 km in altitude and $\pm 30^\circ$ in latitude.

The free air gravity anomaly at an altitude of 100 km (or a radius of 1836 km from the lunar center of mass) may be readily calculated from the disk model of Wong et al. [1975] and is shown in Figure 4 (a similar figure is given by Sjogren [1974]). Unfortunately, the error in this free-air anomaly field is not as readily determined. Wong et al. [1975] did not determine formal errors in the mass of each disk. An estimate of the typical uncertainty in the free-air gravity field may be obtained, however, from the time derivatives of the differences between the LOS Doppler observations and the predictions of the disk-mass model of

Wong et al. [1975]. For 5 orbits over the central nearside with Doppler residual data displayed by Wong et al. [1975] and with spacecraft altitudes between 80 and 160 km, the residual LOS acceleration is typically within ± 7 mgal. For instance, the LOS acceleration residual for the Apollo 15 subsatellite at 100 to 120 km altitude over Serenitatis, Crisium, and Smythii for orbit 1516 are 6, 4, and 6 mgal, respectively. While these LOS acceleration residuals over the central nearside (e.g., Serenitatis) provide a measure of the error in the derived free-air gravity anomaly, the residuals at greater longitudes (e.g., Smythii) reflect more nearly horizontal spacecraft accelerations and are less simply related to errors in the gravity field. For the purpose of estimating the error in the derived crustal structure in the next section of this chapter, we estimate that the free-air anomalies at 100 km elevation shown in Figure 4 have an associated error of ± 10 mgals over the central nearside and ± 20 mgals at the corners of the area represented by the disk-mass solution.

Information on lunar topography comes from a variety of sources, including Apollo laser altimetry [Roberson and Kaula, 1972; Wollenhaupt and Sjogren, 1972; Wollenhaupt et al., 1974; Kaula et al., 1972, 1973, 1974], landmark tracking [Wollenhaupt et al., 1974], limb profiling [Watts, 1963], and photogrammetry [Hopmann, 1967; Mills and Sudbury, 1968; Arthur and Bates, 1968]. Bills and Ferrari [1977b] synthesized a large number of these topographic measurements from all sources and placed them in a consistent center-of-mass coordinate system.

In this chapter we use averages, in blocks of 5° latitude by 5° longitude, of 15,887 observations of nearside topography compiled by B.G. Bills [personal communication, 1982]. There are a total of 672 blocks spanning the latitude range $\pm 40^\circ$ and the longitude range $\pm 105^\circ$. In regions where the topography is poorly resolved, values from the harmonic representation of Bills and Ferrari [1977b] supplement the block averages. The topography of the eastern half of the Orientale basin, the youngest of the nearside basins, has recently been reevaluated by Head et al. [1981] using Earth-based telescopic measurements of limb heights [Watts, 1963]. Apollo laser altimetry over the northern edge of Orientale suggests that while these limb height measurements provide an accurate representation of the basin relief, they require a downward adjustment of about 2 km to fit smoothly with the other topographic data referenced to the lunar center of mass. The Orientale limb height measurements, so adjusted, complete the topographic data set. A contour map of the topography is shown in Figure 5. Averages of the 5° by 5° topographic measurements used here are given in Appendix B.

The accuracy of the block-averaged topography depends on both the technique employed in making individual measurements and the number of measurements in each block. For the individual determinations of topographic height from Apollo laser altimetry and orbital photogrammetry, Bills and Ferrari [1977b] have estimated an error of ± 0.3 km. The majority of these measurements were made between 15°S and 40°N latitude on the nearside of the Moon. Error estimates for topography measured over the remainder of the nearside using landmark

tracking, earth-based photogrammetry, and limb profiling range from 0.4 to 0.9 km.

The error in the average topography over a $5^\circ \times 5^\circ$ block can be estimated from the uncertainties in the individual determinations within that block and from the assumption that all individual measurements are independent. The estimated error in the average topography over Orientale, Humorum, and Nubium for some blocks approaches the estimates for individual observations because these basins are poorly covered by spacecraft tracking data. As a result, errors in topography over these basins represent a large source of uncertainty in the analysis of associated basin structure. For example, the contribution to the Bouguer correction from an 0.5 km error in topography within a 5° by 5° block of crust of density 2.9 g/cm^3 near the lunar equator is about 20 mgals at 100 km altitude, or up to twice the estimated uncertainty in the free-air anomaly derived from the mass disk gravity model. Because the total error in the Bouguer correction is a function of errors in topography everywhere, a baseline error in the topography over a larger region would produce a still greater error in the calculated gravity anomaly.

Fortunately, the large quantity of topographic data collected by orbiting spacecraft from the central nearside results in greatly improved estimates of the average topography with $5^\circ \times 5^\circ$ blocks in that region. Uncertainty in the mean topography of blocks covered by Apollo orbital tracking, some of which are sampled by as many as 135 observations, may be less than 0.1 km.

Thus the contribution of uncertainty in topography to errors in the Bouguer anomaly over the central nearside is small.

The "observed" Bouguer anomalies are calculated above the center of each block at two altitudes, 100 and 200 km, which bracket the altitudes of the majority of Apollo 14, 15 and 16 orbits used to derive the disk gravity model [Wong et al., 1975]. We found that using 1344, rather than 672, gravity "observations" greatly improves the stability of the inversion and acts to smooth the resulting structural model. Of course, the 1344 "observations" are not fully independent. The mass disk model is described by 1400 parameters, but many of these parameters are strongly correlated [Wong et al., 1975].

The Bouguer gravity anomaly field obtained from the application of equations (3-5) to the observed topography (Figure 5) and free-air gravity anomalies (Figure 4) is shown in Figure 6. The topographic datum D_T assumed in calculating the topographic correction is a sphere of radius of 1736 km, selected to be representative of the typical elevation of central nearside regions. The mascon basins Orientale, Humorum, Imbrium, Serenitatis, Nectaris, Crisium, and Smythii display the most prominent Bouguer anomalies on the figure. Orientale, Serenitatis, and Crisium are sites of the largest anomalies (> 200 mgal) relative to surrounding regions.

The estimated error $(\sigma_g)_k$ in the Bouguer anomaly field over the k th block due to uncertainty in both the gravity and topographic data sets may be estimated from:

$$(\sigma_g)_k^2 = (\sigma_f)_k^2 + \sum_{j=1}^M (\sigma_t)_{jk}^2 \quad k=1,2,\dots,N \quad (11)$$

where $(\sigma_f)_k$ is the error in the free-air gravity anomaly over the k th block, and $(\sigma_t)_{jk}$ is the error in the Bouguer correction for the k th block due to the uncertainty in mean topography for the j th block. The area of most reliable Bouguer gravity lies beneath the orbital tracks of the Apollo spacecraft. This area includes the Serenitatis, Tranquillitatis, Crisium, Fecunditatis, Nectaris, and Smythii basins (Figure 1). The Bouguer gravity anomalies over Orientale, Humorum, and Nubium are less accurately represented, but these basins have nonetheless been included in our study for completeness. Signatures of the Imbrium and Australe basins are visible in the gravity and topographic data, but the structures determined in this study for these regions, located at the edge of the area of good coverage, should be regarded as highly uncertain.

BASIN STRUCTURAL MODELS

We have inverted the Bouguer gravity anomaly field of Figure 6, following the inversion procedure described above, to obtain values for the thicknesses of the non-mare and mare basalt components of the crust in each of the 672 blocks on the lunar nearside. Maps of these quantities are shown in Figures 7 and 8, respectively. The crustal thicknesses in blocks containing the Apollo 12 and 14 landing sites are within 1.5 km of the seismically measured value of 55 km. After the final iteration, the r.m.s. residual Bouguer anomaly was 6.5 mgal.

The uncertainty $(\sigma_c)_k$ in the thickness of t_c non-mare crustal material in the k th block can be estimated approximately with the

assumption that it is primarily the result of the uncertainty in the Bouguer anomaly for that block. Under this assumption

$$(\sigma_c)_k = \frac{s_k^2 (\sigma_g)_k}{G A_k \Delta \rho} \quad (12)$$

where s_k is the distance from the observation point (100 km elevation) to a point located at the centroid of the block of non-mare crustal material. We use $s_k = 127.5$ km for all k . The estimated uncertainties in the values of t_c shown in Figure 7 are displayed in map form in Figure 9. The estimated error in t_c over most of the nearside ranges from about 2 to 3 km in areas beneath the tracks of Apollo spacecraft (between 15°S and 40°N latitude) to about 5 km at high latitudes. Errors in t_c derived for the western limb, including the Orientale basin, are generally larger (up to 12 km) because of the uncertainty in the correct baseline for the local topography. Near the edges of the grid area, additional errors may arise from neglect of structural variations beyond the grid and from the assumption of a uniform topographic datum. We repeated the inversion using Bouguer anomaly data obtained with D_T larger by up to 3 km than the value assumed above and found the resulting structure similar to that of Figures 7 and 8; even in blocks at the edge of the grid the crustal thicknesses differ by only a few percent for different values of D_T . Another source of error in the computed structure is the uncertainty (± 0.1 g/cm³) in the assigned densities. An underestimate of the density contrast between crustal and mare or mantle material of 20%, for instance, results in an overestimate of both the Moho relief and mare basalt thickness by about the same percentage.

It is important to recognize that relaxation of the pre-mare isostasy constraint would not alter significantly the values of crustal thickness depicted in Figure 7. If isostatic compensation of the pre-mare basin topography was less than complete or if some percentage of the mare fill has been compensated, the actual thickness of mare basalt would be greater than in our model by some amount. Because mantle material and mare basalt are similar in density, however, the Moho relief would be less by an approximately equal amount. This approximately even trade-off of anomalous masses occurs because the Bouguer anomaly, when measured 100 to 200 km above the sources of anomalous mass, is much more sensitive to the total anomalous mass than to the distribution of that mass between the surface and Moho. If high-density basaltic to gabbroic intrusions (dikes, sills) permeate the crust beneath the basalt fill in large basins [e.g., Janle, 1981a,b], then lesser amounts of Moho relief and mare fill than shown in Figures 7 and 8 would be necessary to match the observed gravity. The equivalent thickness of low-density crustal material, however, would not differ greatly from the values shown in Figure 7. Though some system of magma conduits clearly must have existed at depth to feed the extensive mare units within most basins, the volume of such structures is not known.

The crustal thickness model of Figure 7 indicates that the crust is thinner beneath each of the major nearside basins than in immediately surrounding areas. Such a result, of course, would follow at least qualitatively from the assumption of pre-mare isostasy. We interpret this thinner crust to be

principally the result of the basin formation process, including formation of the transient cavity and collapse of the cavity by some combination of inward and upward flow of crustal and mantle material [e.g., Melosh and McKinnon, 1978]. By this interpretation, the Moho relief beneath a basin region is a measure of the amount of uplift of mantle material during cavity collapse, an important parameter in describing the thermal evolution of a basin region [Bratt et al., 1981; Chapter 6]. Of course, the crustal model of Figure 7 reflects only the present Moho relief beneath basins; the relief immediately following basin formation may have been substantially modified by the effects of ductile flow in the crust.

Cross sections through the basin structural models are shown in Figure 10. The dashed lines show the Moho depth according to the block model along the profile lines shown in Figure 1. The solid lines show smoothed cross sections of both the Moho and the base of mare basalt deposits. The smoothed profiles were obtained from the contour maps in Figures 7 and 8 and represent azimuthally averaged cross sections. From the smoothed Moho profiles and the assumption of cylindrical symmetry, we have estimated the magnitude of Moho relief and the apparent volume V_u of uplifted mantle beneath each basin. These quantities are listed in Table 1, in order of increasing basin age [Wilhelms, 1981].

Table 1 and Figure 10 suggest a general decrease in the apparent volume of uplifted mantle with increasing basin age. The Moho beneath the central mare units of Orientale (Figure 10a) is raised by 66 km relative to the ambient depth of about 80 km. Serenitatis (Figure 10b), a basin about the same diameter as

Oriente and of somewhat greater age, is surrounded by crust with an average thickness of 55 km; mantle relief beneath Serenitatis is 30 km, but the uplifted volume is broader and thus somewhat larger ($7.5 \times 10^6 \text{ km}^3$ versus $6.0 \times 10^6 \text{ km}^3$) than that beneath Oriente. The volume of apparent mantle uplift beneath Oriente is also more uncertain, however, because of poor data coverage over the western portion of the basin. The difference in the magnitude of Moho relief between the two basins nonetheless exceeds the sum of the errors in these two quantities (Figure 9). The oldest basin examined is Tranquillitatis (Figure 10i). Though slightly larger in diameter than the nearby Serenitatis basin, Moho relief beneath Tranquillitatis is only 12 km, also a significant difference in relation to the errors shown in Figure 9. We interpret the decrease in the apparent mantle uplift with increasing basin age (Table 1, Figure 10) to be the result of an age dependence of the basin formation or modification processes. We consider the implications of this inference in a later section.

Figure 7 shows a generally thinner crust over the central nearside of the Moon than near the limb regions, a result also noted in several earlier studies of lunar crustal structure [e.g., Kaula et al., 1974]. This large region of thinner crust coincides approximately with the location of the ancient Procellarum basin [Whitaker, 1981], a 3200-km diameter structure centered near 26°N and 15°W . This coincidence suggests that some record of mantle uplift during basin formation is still apparent even for the largest and one of the oldest of the preserved nearside basins [Wilhelms, 1981].

The thicknesses of mare basalt, shown in map view in Figure 8, are deserving of several comments. The inversion procedure preserved the constraint of pre-mare isostasy within all major basins for the complete set of iterations. Equivalently, the Bouguer gravity anomaly over these basins is sufficiently large so as to require both an elevated Moho and more than 250 m of high density mare basalt. Basalt thicknesses beneath the centers of the irregular mare basins (Nubium, Fecunditatis, and Tranquillitatis) range from 0.5 to 1.3 km, while the younger circular mare basins (Orientale, Serenitatis, Crisium, Humorum, Nectaris and Smythii) contain as much as 3.5 to 4.5 km of basalt fill within the central depression region. That these thicknesses in the central portions of the circular basins are greater by up to 2 km than the values derived by Thurber and Solomon [1978] under a similar set of assumptions can be attributed to the improved accuracy of the topographic data used here and to the method adopted to evaluate the gravitational potential due to anomalous mass (equation 3). It should be noted, however, that the region of thickest mare fill in each of the Orientale and Smythii basins constitutes only a single block in the model depicted in Figure 8. In some irregular mare regions (e.g., Oceanus Procellarum), the constraint of pre-mare isostasy was relaxed during the inversion because the basalt thickness within one or more blocks dropped below the 250 m limit after an iteration. For such regions, either the surficial mare basalt unit is very thin or the crust is somewhat thicker than would be predicted from complete isostatic compensation of pre-mare relief.

As noted earlier, the mare basalt thicknesses depicted in Figure 8 should probably be regarded only as lower bounds on the actual thicknesses within the central regions of most basins [Thurber and Solomon, 1978]. If the lunar lithosphere maintained finite strength during the formation of at least the younger basins, then the pre-mare basin would have had incompletely compensated relief and a greater thickness of mare basalt would be required to match the Bouguer gravity field. Further, some portion of the mare basalt load in the mascon maria has likely been compensated by lithospheric flexure [e.g., Solomon and Head, 1979, 1980], which also would lead to mare units somewhat thicker and Moho relief somewhat less than indicated by the structural model presented here. It is interesting to note that mare thicknesses estimated by DeHon and Waskom [1976] and DeHon [1978] from rim heights of craters buried or partially buried by mare deposits are generally comparable to the values in Figure 8 for many mare regions. This similarity between the two determinations is probably coincidental, however, because the rim-height technique tends to underestimate mare thickness by as much as a factor of 2 [Head, 1982].

It is also possible that the mare basalt thicknesses shown in Figure 8 for some regions may be overestimates because of a significant contribution to the gravity field from subsurface intrusions of high-density mare basaltic material. The thickness of mare basalt within the central block of the Orientale basin, for instance, is 3 to 4 km in Figure 8. On photogeological grounds, however, the basalt thickness has been estimated to be

no greater than 1 km [Head, 1974]. A significant excess mass within the sub-mare crust is a possible explanation for this discrepancy. The large uncertainty in the topography and gravity in the Orientale region may be at least an equal contributor, however.

GEOLOGICAL IMPLICATIONS

The structural models derived in this chapter have a number of implications for the processes that accompanied the formation and evolution of large impact basins on the Moon. We divide the discussion into basin formation processes, including transient cavity collapse and emplacement of ejecta deposits, and basin modification processes, including volcanism and lithospheric deformation on time scales longer than those normally associated with the basin formation event. We recognize that this division is somewhat arbitrary and that the present characteristics of each basin may be the product of a combination of both types of processes.

Early Cavity Geometry

The volume of material ejected from a basin during impact is an important quantity, for several reasons. Its relation to basin diameter provides a strong constraint on the cratering process for large impacts, particularly on the early cavity geometry. The ejecta volume also provides information on the sampling depths of large impacts and thus on the interpretation of returned lunar samples for crustal composition. Head et al. [1975] have defined the volume V_p of "primary ejecta" as the volume of material excavated from a crater and thrown beyond the

crater rim. The quantity V_p is thus approximately equal to the volume of the early transient cavity minus the volume of material which falls back in to the collapsing cavity. This equality ignores the effects of density changes due to compression of target material, brecciation of fallback material, or other inelastic effects. The volume of primary ejecta residing outside a major basin, however, is a difficult quantity to determine from photogeologic observations of ejecta deposits, among other reasons because of the difficulty in separating primary ejecta from the locally-derived, secondary ejecta surrounding the basin [Head et al., 1975] . From a variety of observations, models, and scaling relations, Head et al. [1975] estimated that V_p for a basin the size of Orientale lies between $0.4 \times 10^6 \text{ km}^3$ and $12 \times 10^6 \text{ km}^3$. This wide range of possible ejecta volumes is equivalent to a large uncertainty in the thickness of ejecta deposits expected from an Orientale-sized impact. For instance, if the ejecta volume were spread evenly over the lunar surface, the thickness could be as little as 10 m or as large as 300 m.

More recently, Spudis [1982, 1983] and Spudis et al. [1984] have estimated the depth of excavation for lunar basins from remote geochemical observations and scaling relations under the assumption that the lunar crust is layered, with an anorthosite layer overlying a more noritic layer [Ryder and Wood, 1977]. Spudis [1982, 1983] and Spudis et al. [1984] concluded that the chemistry and mineralogy of the ejecta deposits of 5 nearside basins provide support for an excavation model in which the ratio of the depth of excavation to basin diameter is about 0.1 for

basin-size impacts. On the basis of such a model, the excavation depth for the younger nearside basins ranged from 40 to 80 km [Spudis, 1983] and the volume V_p of primary ejecta excavated during the Orientale basin impact is estimated to be 5 to $8 \times 10^6 \text{ km}^3$ [Spudis et al., 1984]. These estimates have considerable uncertainty, however, associated with the reality of the assumed simple model of crustal layering and with the ability to separate primary from secondary ejecta.

The structural models for lunar nearside basins depicted in Figure 10 provide a straightforward and internally consistent framework with which to address cavity and ejecta volumes. Let V_b be the volume of the present basin, including the volumes of the topographic depression and of mare fill. We make the assumption that the crustal thickness of the pre-impact basin region was similar to the present thickness at 2 to 3 basin radii from the basin center (excluding other nearby basins). Then the quantity $V_e = V_b + V_u$, where V_u is the volume of apparently uplifted mantle, provides a measure of pre-impact crustal material now external to the basin rim.

The quantity V_e should be somewhat less than V_p . This is because inward transport of crustal material during basin modification as well as deposition of ejecta from younger basins or of non-mare volcanic material will act to reduce one or both of V_b and V_u . Therefore, V_e can be regarded as the difference between the volume of crustal material ejected beyond the basin rim and the volume of crust returned to the basin region by these later processes. For the youngest nearside basins, which

may have formed at a time when near-surface temperatures were comparatively low and the elastic lithosphere relatively thick (e.g., Orientale, Serenitatis, and Crisium), V_e may approach V_p .

Values for V_e derived from the smoothed cross sections of Figure 10 are given in Table 1. The values range from 2 to 9×10^6 km³. It should be repeated that these values are insensitive to the assumption of pre-mare isostasy, because to first order the Bouguer anomaly data yield the total anomalous mass (mantle uplift plus mare basalts), and it is the sum of the two components that contributes to V_e . By analogous reasoning, these values of V_e are also insensitive to the presence of high-density intrusions within the low density crust underlying the basin. The uncertainties in gravity and topography and in the assumed densities and model simplifications result in an uncertainty in V_e of about 30% for Orientale and 20% for the other basins studied. The values of V_e in Table 1 fall within the range of estimates for V_p for basins the size of Orientale or Serenitatis [Head et al., 1975; Spudis et al., 1984]. The average value for V_e for the three youngest basins studied (Orientale, Serenitatis, and Crisium) is 8×10^6 km³ and falls near the upper end of this range. An ejecta volume of 8×10^6 km³, if spread evenly over the lunar surface, would constitute a layer 200 m thick. Considering the 30 or more impact basins preserved on the Moon [e.g., Wood and Head, 1976], such an average thickness for ejected material from a single basin suggests that the accumulated thickness of basin ejecta deposits on the lunar surface may, on average, be several kilometers.

An interesting relationship between apparent mantle uplift and crustal thickness is suggested by the data from the 6 youngest basins in Table 1. The thickness t_c of the non-mare crust beneath the central region of each of these 6 basins is remarkably uniform, 20 to 30 km, despite differences in basin size, age, and pre-impact crustal thickness. The Moho relief for 5 basins in crust 55 to 65 thick is nearly constant at about 30 km, despite a variation of 200 km in basin diameter. While the Moho relief beneath Orientale is greater than that of the other 5 basins by 30 to 40 km, the surrounding crust is also thicker by 20 to 30 km.

We suggest that this near-uniformity of non-mare crustal thickness beneath the younger basins can be understood only if excavation of the transient cavity for basins 400 to 600 km in diameter extended to the base of the crust or to the uppermost mantle. If the excavated volumes did not extend at least to near the base of the crust for these basins (Figure 11a), then the value of t_c for a basin of a given diameter should be greater for a thicker pre-impact crust. The quantity t_c should also decrease with increasing diameter for basins formed in crust of similar thickness. Neither relation is indicated by the values in Table 1. Of course, the youngest 6 basins in the table may have been affected to varying degrees by long-term viscous relaxation and other modification processes, but it is hard to imagine how such processes would have led coincidentally to the similar present values of t_c . As noted above, the conclusion that excavation of the youngest nearside basins extended to the lower

crust has been suggested on independent geochemical and photogeological grounds by Spudis [1982, 1983] and Spudis et al. [1984].

If we accept that the excavated volumes for the basins in Table 1 extended in depth at least to the lowermost crust, we still must seek a mechanism for the near constancy of t_c . One explanation is motivated by the observation that mantle material is significantly stronger than crustal material at laboratory strain rates [e.g., Brace and Kohlstedt, 1980]. If this difference in strength can be extrapolated to the high strain rates thought to accompany cavity formation [e.g., O'Keefe and Ahrens, 1977], the Moho may have acted more or less as a barrier to the deepening of cavity excavation (Figure 11b). The 20 to 30 km thickness of non-mare crust beneath the central regions of large young impact basins, by this hypothesis, would be a combination of ejecta fallback and crustal material transported laterally during cavity collapse. These two components would be difficult to distinguish.

An alternative explanation for the uniformity of t_c is that the transient cavity for the largest basins excavated through the crust and into the mantle (Figure 11c). If the equivalent thickness of pre-impact crustal material that fell back into the central basin region was only weakly dependent on basin diameter, then the indicated value of t_c would be similar for basins that had excavated to different mantle depths. Of course, the actual ejecta would have consisted of a mixture of crustal and mantle material, but the inversion procedure we have followed cannot

distinguish a thin non-mare crust from a thicker layer with the same volume of low-density crustal rock admixed with mantle material. The paucity of candidates in the Apollo sample collection for material from the lunar mantle [see Head et al., 1975], however, suggests that excavation depths rarely exceeded the Moho depth by any significant amount. We therefore favor the first explanation.

Basin Modification

One of the principal results of this chapter is that the extent of apparent mantle uplift or crustal thinning preserved beneath large lunar basins generally decreases with increasing basin age. The cross sections shown in Figure 10 illustrate this trend. We have attributed this variation to more extensive modification of older basins than younger ones. The time scale for this modification is highly uncertain and may range from the time scale of cavity collapse to the time interval between basin formation and one or more major episodes of mare basalt volcanism. In this section, we consider the modification processes that can act to reduce basin topographic and Moho relief, and we evaluate the implications of our structural models for the variation of these processes with time on the Moon.

A significant element of all endogenic basin modification processes is the additional heating at depth that accompanies basin formation [Bratt et al., 1981; Chapter 6]. Both ductile flow and magma genesis can be expected to be enhanced by this type of deep heating. Because the outer portions of the Moon have generally cooled with time, the effects of such temperature-dependent

processes might be enhanced for basins formed at an early stage in lunar history. Older basins are also more likely to sense the lingering effects of the formation of large, pre-existing basins in the same region.

We consider four principal basin-modification processes for young and old lunar basins: (1) ejecta from younger basins; (2) ductile flow of the crust, (3) non-mare volcanic activity, and (4) mare volcanism.

Basin ejecta. The topographic relief of an older basin is reduced by the infill of ejecta deposits from younger basins. This effect probably constitutes only a minor contribution, however, to the topographic modification of most lunar basins. The difference in the amount of crustal thinning between the youngest nearside basins (e.g., Orientale, Serenitatis, Crisium) and older basins of comparable diameter (Nubium, Fecunditatis, Tranquillitatis) is several tens of kilometers. The accumulated thickness of younger basin ejecta deposits in these older basins is not likely to exceed a few kilometers [McGetchin et al., 1973].

Ductile flow. Viscous relaxation of an impact basin by ductile flow of crustal material can reduce both topographic and Moho relief [Solomon et al., 1982]. Because of the strong dependence of effective viscosity on temperature, viscous relaxation was likely to have been a more important modification process for older basins formed when near-surface temperatures on the Moon were relatively high and the elastic lithosphere was thin. In particular, the pre-mare topography

and the crustal thickness in the Tranquillitatis region are both consistent with viscous relaxation from an initial structure similar to the present structure of Orientale [Solomon et al., 1982]. It is, of course, uncertain whether the oldest basins actually had initial structures similar to that of Orientale or, perhaps because of a different thermal state at the time of basin formation, never displayed the topographic relief and extent of crustal thinning now observed for young lunar basins; the structural models derived from present gravity and topography cannot distinguish between these alternatives (Figure 12).

The formation of the Nubium, Fecunditatis, and Tranquillitatis basins--not resolvably different in age [Wilhelms, 1981]--may have occurred over a sufficiently short interval of time following the formation of the 3200-km diameter Procellarum basin [Whitaker, 1981] so that the thermal effects of the Procellarum impact on the creep rates in the crust were still important [Wilhelms, 1983]. Nubium and Tranquillitatis lie just within, and Fecunditatis just outside, the third ring of the Procellarum basin [Whitaker, 1981] and are thus within a radial range where significant thermal effects can be expected [Bratt et al., 1981; Chapter 6]. As noted earlier, the Procellarum basin may have contributed to the generally lower elevations and thinner crust on the central nearside than elsewhere on the Moon. That the Procellarum basin did not itself experience complete viscous relaxation of surface and Moho topography may be ascribed to the insensitivity of the controlling time constant to spatial scale

[Solomon et al., 1982] and perhaps as well to the constraints on lateral flow imposed by spherical geometry when the basin diameter approaches the diameter of the Moon.

Non-mare volcanism. An additional modification process that can act to reduce basin relief and that would be enhanced by higher temperatures is non-mare volcanism; i.e., deposition of non-mare basalts with a density similar to the typical density of the crust and therefore not resolvable from gravity anomalies. The volumetric significance of non-mare volcanism for lunar crustal evolution in general and for old lunar basins in particular is not well known. To the extent that the crust, as opposed to the upper mantle, is the source of magmas for this volcanic activity, this process involves lateral transport of crustal material to fill the basin depression and is not readily distinguishable from ductile flow (Figure 12). If temperatures exceed the solidus in crustal material adjacent to a newly formed basin, in fact, the effective viscosity might be sufficiently low so that bulk flow would dominate magma transport as a mechanism for reducing relief.

Mare volcanism. The source regions of mare basalt magmas are thought to lie at depths in excess of 100 km [e.g., Wyllie et al., 1981]. At such depths, the deposition of kinetic energy as heat, the release of hydrostatic pressure, and the uplift of mantle isotherms during basin formation [Bratt et al., 1981; Chapter 6] are likely to have had only a small effect on magma genesis, and then only if the impact was very large [Hubbard and Andre, 1983] and the ambient temperature was already near the solidus. That

the thickness of mare basalts appears to be less within the older basins (Nubium, Fecunditatis, Tranquillitatis) than nearby younger basins (Figure 8) suggests that mare volcanism probably has not contributed substantially to the age dependence of preserved basin structure.

Of the basin modification processes considered above, lateral flow of lunar crustal material, perhaps augmented by non-mare volcanism, is the most likely explanation of the general decrease of preserved mantle uplift with increasing basin age. Because both of these processes involve the radially inward movement of crustal material, the two are difficult to separate on the basis of gravity and topographic data alone. The poor degree of preservation of relief of the older basins on the central nearside may be a consequence of both the generally high temperatures in the early lunar crust and the extensive additional crustal and sub-crustal heating associated with the formation of the ancient Procellarum basin.

SUMMARY AND CONCLUSIONS

We have presented models for the crustal and upper mantle structure beneath major impact basins on the nearside of the Moon. The models are derived from an inversion of nearside gravity and topography, subject to the assumptions that pre-mare topography was isostatically compensated by an Airy mechanism and that compensation of mare basalt units may be neglected. An additional constraint is the crustal thickness inferred beneath the Apollo 12 and 14 landing sites from seismic observations.

The basin structural models are characterized by thinned

crust and an elevated Moho beneath the central basin regions, compared to surrounding areas, presumably the preserved effects of basin excavation and mantle uplift during collapse of the transient cavity. Orientale exhibits over 60 km of apparent mantle uplift, while the Moho relief beneath Serenitatis, Crisium, Humorum, Nectaris and Smythii is about 30 km. The three oldest basins included in this study, Nubium, Fecunditatis, and Tranquillitatis, display only a modest (~ 10 km) amount of preserved Moho relief.

The general decrease in the amount of preserved mantle uplift with increasing basin age is attributed primarily to enhanced rates of lateral flow of crustal material early in lunar history when crustal temperatures were relatively high and the elastic lithosphere comparatively thin. Non-mare volcanic activity may also have contributed to the greater extent of modification of older basins. The relaxed topographic and Moho relief for older basins on the central nearside, in particular, may be at least partly a consequence of the extensive crustal and subcrustal heating associated with the formation of the large Procellarum basin.

If we make the assumption that the pre-impact crustal thickness beneath the basin region was similar to the present thickness beneath surrounding regions, then the sum of the volume of uplifted mantle, the volume of mare basalt fill, and the volume of the topographic basin provides a lower bound to the volume of crustal material ejected beyond the basin rim during basin formation. This bound may be similar to the true ejected volume for the youngest basins formed when the elastic lithosphere may have been

significant in thickness and when any modification subsequent to cavity collapse may have been relatively minor. On this basis, we estimate the volume of crustal material ejected from basins the size of Orientale to be on the order of 10^7 km³.

The thickness of non-mare crustal material beneath the central regions of the 6 youngest basins considered in this study is remarkably uniform (20 to 30 km), despite large differences in basin size and in the pre-impact thickness of the crust. These results suggest the hypothesis that basin excavation extended in depth at least to the lowermost crust for these impacts, and that significant deepening of cavity excavation for the largest basins was most likely impeded by an abrupt increase in strength at the crust-mantle boundary. The preserved layer of non-mare crust beneath the basin centers, by this view, may be some combination of fallback and crustal material transported laterally during cavity collapse.

The structural models described in this chapter provide important constraints on the thermal budget of the basin-forming process and on the subsequent thermal and tectonic evolution of the basin region [Bratt et al., 1981; Chapter 6]. While these models are of necessity simple, they provide a basis for testing the ideas presented here with data of higher spatial resolution in individual basin regions on the Moon and the other terrestrial planets.

REFERENCES

- Arthur, D.W.G., and P. Bates, The Tucson selenodetic triangulation, Commun. Lunar Planet. Lab., 7, 313-361, 1968.
- Baldwin, R.B., Lunar mascons: Another interpretation, Science, 162, 1407-1408, 1968.
- Bills, B.G., and A.J. Ferrari, A lunar density model consistent with topography, gravitational, librational, and seismic data, J. Geophys. Res., 82, 1306-1314, 1977a.
- Bills, B.G., and A.J. Ferrari, A harmonic analysis of lunar topography, Icarus, 31, 244-259, 1977b.
- Booker, J.R., R.L. Kovach, and L. Lu, Mascons and lunar gravity, J. Geophys. Res., 75, 6558-6564, 1970.
- Bowin, C., B. Simon, and W.R. Wollenhaupt, Mascons: A two body solution, J. Geophys. Res., 80, 4947-4955, 1975.
- Brace, W.F., and D.L. Kohlstedt, Limits on lithospheric stress imposed by laboratory experiments, J. Geophys. Res., 85, 6248-6252, 1980.
- Bratt, S.R., S.C. Solomon, and J.W. Head, The evolution of multi-ringed basins: Cooling, subsidence, and thermal stress (abstract), Lunar Planet. Sci., 12, 109-111, 1981.
- Conel, J.E., and G.B. Holstrom, Lunar mascons: A near-surface interpretation, Science, 162, 1403-1405, 1968.
- DeHon, R.A., Maximum thickness of materials in the western mare basins (abstract), Lunar Planet. Sci., 9, 229-231, 1978.
- DeHon, R.A., and J.D. Waskom, Geologic structure of the eastern mare basins, Proc. Lunar Sci. Conf. 7th, 2729-2746, 1976.

- Head, J.W., Orientale multi-ringed basin interior and implications for the petrogenesis of lunar highland samples, Moon, 11, 327-356, 1974.
- Head, J.W., Lunar volcanism in space and time, Rev. Geophys. Space Phys., 12, 265-300, 1976.
- Head, J.W., Origin of outer rings in lunar multi-ringed basins: Evidence from morphology and ring spacing, in Impact and Explosion Cratering, edited by D.J. Roddy, R.O. Pepin, and R.B. Merrill, pp. 563-573, Pergamon Press, New York, NY, 1977.
- Head, J.W., Lava flooding of ancient planetary crust: Geometry, thickness, and volumes of flooded lunar impact basins, Moon and Planets, 26, 61-88, 1982.
- Head, J.W., M. Settle, and R.S. Stein, Volume of material ejected from major lunar basins and implications for the depth of excavation of lunar samples, Proc. Lunar Sci. Conf. 6th, 2805-2829, 1975.
- Head, J.W., E. Robinson, and R.J. Phillips, Topography of the Orientale basin (abstract), Lunar Planet. Sci., 12, 421-423, 1981.
- Hopmann, J., General-Katalog absoluter Höhen auf dem Mond, Ann. Univ. Sternw. Wien, 26, 7 Heft, 1967.
- Hubbard, N., and C.G. Andre, Magma genesis in a battered Moon: Effects of basin-forming impacts, Moon and Planets, 29, 15-37, 1983.
- Hulme, G., Mascons and isostasy, Nature, 238, 448-450, 1972.

- Janle, P., A crustal gravity model of the Mare Serenitatis - Mare Crisium area of the Moon, J. Geophys., 49, 57-65, 1981a.
- Janle, P., A lunar crustal gravity model across the Triesnecker - Hyginus area, Mare Tranquillitatis, and Mare Fecunditatis, Moon and Planets, 24, 441-451, 1981b.
- Kaula, W.M., G. Schubert, R.E. Lingenfelter, W.L. Sjogren, and W.R. Wollenhaupt, Analysis and interpretation of lunar laser altimetry, Proc. Lunar Sci. Conf. 3rd, 2189-2204, 1972.
- Kaula, W.M., G. Schubert, R.E. Lingenfelter, W.L. Sjogren, and W.R. Wollenhaupt, Lunar topography from Apollo 15 and 16 laser altimetry, Proc. Lunar Sci. Conf. 4th, 2811-2819, 1973.
- Kaula, W.M., G. Schubert, R.E. Lingenfelter, W.L. Sjogren, and W.R. Wollenhaupt, Apollo laser altimetry and inferences as to lunar structure, Proc. Lunar Sci. Conf. 5th, 3049-3058, 1974.
- Lawson, C.L., and R.J. Hanson, Solving Least Squares Problems, 340pp., Prentice-Hall, Engelwood Cliffs, NJ, 1974.
- Liu, A.S., and P.A. Laing, Lunar gravity analysis from long-term effects, Science, 173, 1017-1020, 1971.
- McGetchin, T., M. Settle, and J. Head, Radial thickness variations in impact crater ejecta: Implications for lunar basin deposits, Earth Planet. Sci. Lett., 20, 226-236, 1973.
- Melosh, H.J., and W.B. McKinnon, The mechanics of basin formation, Geophys. Res. Lett., 5, 985-988, 1978.

- Mills, G.A., and P.V. Sudbury, Absolute coordinates of lunar features, Icarus, 9, 538-561, 1968.
- Moore, H.J., C.A. Hodges, and D.H. Scott, Multi-ringed basins - illustrated by Orientale and associated features, Proc. Lunar Sci. Conf. 5th, 71-100, 1974.
- Morrison, F., Algorithms for computing the geopotential using a simple density layer, J. Geophys. Res., 81, 4933-4936, 1976.
- Muller, P.M., and W.L. Sjogren, Mascons: Lunar mass concentrations, Science, 161, 680-684, 1968.
- O'Keefe, J.D., and T.J. Ahrens, Impact-induced energy partitioning, melting, and vaporization on terrestrial planets, Proc. Lunar Sci. Conf. 8th, 3357-3374, 1977.
- Phillips, R.J., and J. Dvorak, The origin of lunar mascons: Analysis of the Bouguer gravity associated with Grimaldi, in Multi-ring Basins, Proc. Lunar Planet. Sci., 12A, 91-104, 1981.
- Phillips, R.J., J.E. Conel, E.A. Abbott, W.L. Sjogren, and J.B. Morton, Mascons: Progress toward a unique solution for mass distribution, J. Geophys. Res., 77, 7106-7114, 1972.
- Roberson, F.I., and W.M. Kaula, Apollo 15 laser altimeter, in Apollo 15 Preliminary Science Report, 25-48 to 25-50, NASA SP-289, 1972.
- Ryder, G., and J.A. Wood, Serenitatis and Imbrium impact melts: Implications for large-scale layering in the lunar crust, Proc. Lunar Sci. Conf. 8th, 655-668, 1977.

- Sjogren, W.L., Lunar gravity at 100 km altitude, Proc. Lunar Sci. Conf. 5th, frontispiece, 1974.
- Sjogren, W.L., and J.C. Smith, Quantitative mass distribution models for Mare Orientale, Proc. Lunar Sci. Conf. 7th, 2639-2648, 1976.
- Solomon, S.C., Mare volcanism and lunar crustal structure, Proc. Lunar Sci. Conf. 6th, 1021-1042, 1975.
- Solomon, S.C., and J.W. Head, Vertical movement in mare basins: Relation to mare emplacement, basin tectonics, and lunar thermal history, J. Geophys. Res., 84, 1667-1682, 1979.
- Solomon, S.C., and J.W. Head, Lunar mascon basins: Lava filling, tectonics, and evolution of the lithosphere, Rev. Geophys. Space Phys., 18, 107-141, 1980.
- Solomon, S.C., R.P. Comer, and J.W. Head, The evolution of impact basins: Viscous relaxation of topographic relief, J. Geophys. Res., 87, 3975-3992, 1982.
- Spudis, P.D., Orientale basin ejecta: Depths of derivation and implications for the basin-forming process (abstract), Lunar Planet Sci., 13, 760-761, 1982.
- Spudis, P.D., The excavation of lunar multi-ringed basins: Additional results for four nearside basins (abstract), Lunar Planet Sci., 14, 735-736, 1983.
- Spudis, P.D., B.R. Hawke, and P. Lucey, Composition of Orientale basin deposits and implications for the lunar basin-forming process, J. Geophys. Res., Proc. 15th Lunar Planet. Sci. Conf., suppl., in press, 1984.

- Thurber, C.H., and S.C. Solomon, An assessment of crustal thickness variations on the lunar near side: Models, uncertainties, and implications for crustal differentiation, Proc. Lunar Planet. Sci. Conf. 9th, 3481-3497, 1978.
- Toksöz, M.N., A.M. Dainty, S.C. Solomon, and K.R. Anderson Structure of the moon, Rev. Geophys. Space Phys., 12, 539-567, 1974.
- Watts, C.B., The marginal zone of the Moon, Astron. Papers Amer. Ephemer. Naut. Almanac, 17, USGPO, Washington, D.C., 1963.
- Whitaker, E.A., The lunar Procellarum basin, in Multi-ring Basins, Proc. Lunar Planet. Sci., 12A, 105-111, 1981.
- Wilhelms, D.E., Relative ages of lunar basins (II); Serenitatis (abstract), in Reports of Planetary Geology Program -1981, NASA TM-84211, 405-407, 1981.
- Wilhelms, D.E., Effects of the Procellarum basin on lunar geology, petrology, and tectonism (abstract), Lunar Planet. Sci., 14, 845-846, 1983.
- Wise, D.U., and M.Y. Yates, Mascons as structural relief on a lunar "Moho", J. Geophys. Res., 75, 261-268, 1970.
- Wollenhaupt, W.R., and W.L. Sjogren, Apollo 16 laser altimeter, in Apollo 16 Preliminary Science Report, 30-1 to 30-5, NASA SP-315, 1972.
- Wollenhaupt, W.R., and W.L. Sjogren, R.E. Lingenfelter, G. Schubert, and W.M. Kaula, Apollo 17 laser altimeter, in

- Apollo 17 Preliminary Science Report, 33-41 to 33-42, NASA SP-330, 1974.
- Wong, L., G. Buechler, W. Downs, W. Sjogren, P. Muller, and P. Gottlieb, A surface-layer representation of the lunar gravitational field, J. Geophys. Res., 76, 6220-6236, 1971.
- Wong, L., W. Downs, W. Sjogren, P. Muller, A lunar mass distribution from Apollo 14, 15, and 16 CSM-LM and subsatellite radio Doppler data, Jet Propulsion Laboratory, 1975.
- Wood, C.A., and J.W. Head, Comparison of impact basins on Mercury, Mars, and the moon, Proc. Lunar Sci. Conf. 7th, 3629-3651, 1976.
- Wood, J.A., Fragments of terra rock in the Apollo 12 soil samples and a structural model of the moon, Icarus, 16, 462-501, 1972.
- Wood, J.A., Bombardment as a cause of lunar asymmetry, Moon, 8, 73-103, 1973.
- Wyllie, P.J., C.H. Donaldson, A.J. Irving, S.E. Kesson, R.B. Merrill, D.C. Presnall, E.M. Stolper, T.M. Usselman, and D. Walker, Experimental petrology of basalts and their source rocks, in Basaltic Volcanism on the Terrestrial Planets, pp. 493-630, Pergamon, Elmsford, NY, 1981.

FIGURE CAPTIONS

- Figure 1. Outlines of the lunar basins considered in this study. The labeled lines indicate the locations of cross sections discussed below.
- Figure 2. Schematic representation of the block model used for calculating contributions to the free-air gravity anomaly from topography (relative to the datum D_T), Moho relief (relative to the datum D_M), and mare basalt fill.
- Figure 3. Schematic cross-section of the crust and upper mantle, illustrating how the assumption of pre-mare isostasy is used to constrain the distribution of anomalous mass in lunar mare regions. We use equation (6) and the observed topography h to constrain the thickness t of mare basalt and the depth d to the Moho in mare blocks. The seismically determined crustal thickness in the vicinity of the Apollo 12 and 14 landing sites serves as a reference section.
- Figure 4. Free-air gravity anomalies over the lunar nearside at 100 km elevation (1836 km from the lunar center of mass). Gravity anomalies were computed from the disk mass model of Wong et al. [1975] superimposed on the triaxial gravity model of Liu and Laing [1971]. The contour interval is 25 mgal. The peak anomalies over several basins are also indicated.

Figure 5. Topography of the lunar nearside, relative to a datum at 1730 km radius. The map is obtained from 5° by 5° averages of topography compiled by Bills and Ferrari [1977b] and also incorporates limb-height observations of the Orientale basin summarized by Head et al. [1981]. The contour interval is 1 km. The elevation minima within several basins are also indicated.

Figure 6. Bouguer gravity anomaly for the lunar nearside at 100 km elevation. A datum D_T at a radius of 1736 km was used to make the topographic correction. The contour interval is 50 mgal. Local maxima over several basins are also indicated.

Figure 7. Crustal thickness (not including mare basalt fill) on the lunar nearside. This model is derived from the Bouguer anomaly data of Figure 6, the constraint that the crust be 55 km thick beneath the Apollo 12 and 14 sites, and the assumptions that, in mare regions, the pre-mare topography was completely compensated by an Airy mechanism and no further compensation followed the emplacement of mare units (see text). Contour interval is 10 km.

Figure 8. Mare basalt thickness on the lunar nearside. See Figure 7 (and text) for model assumptions. The dashed line indicates the boundary of mare regions. Contour interval is 1 km.

Figure 9. Estimated error in crustal thickness (exclusive of mare basalt fill) resulting from uncertainties in gravity and topographic data as calculated using equation (12). The contour interval is 2 km.

Figure 10. Cross-sections of crustal structure for 9 basins on the lunar nearside. Dashed lines denote the Moho as determined from the block model along the profile lines shown in Figure 1. The solid curves depict smoothed profiles of the lunar Moho and the basin topography at the base of the mare basalt (see text). Depths are relative to a datum at 1730 km radius. Basins are illustrated in order of increasing age [Wilhelms, 1981]: (a) Orientale, (b) Serenitatis, (c) Crisium, (d) Humorum, (e) Nectaris, (f) Smythii, (g) Nubium, (h) Fecunditatis, and (i) Tranquillitatis.

Figure 11. Three scenarios for the formation of young impact basins formed when the elastic lithosphere is comparable or greater in thickness than the crust. In all cases, uplift of the sublithospheric mantle dominates the late-stage collapse of the transient cavity. In case (a), the excavated cavity is confined to the crust. In case (b), with a larger ratio of crater diameter to crustal thickness, the excavated cavity is prevented from extending significantly deeper than the Moho by the relatively greater strength of mantle material. In case (c),

with a still larger ratio of crater diameter to crustal thickness, the transient cavity may extend into the uppermost mantle. If the preserved non-mare crust beneath the basin center is dominated by fallback of ejected crustal material in cases (b) and (c), it may be similar in thickness for the two cases. Discovery of mantle material among the returned lunar samples would strengthen the possibility of case (c).

Figure 12. Two possible scenarios to account for the structure of older lunar basins (e.g., Figures 10g through 10i). In case (a), because of the high crustal temperatures and thin elastic lithosphere at the time of impact, flow of both the crust and the mantle accompany transient cavity collapse, leading to a structure similar to that presently observed. In case (b), the collapse of the transient cavity at left is accomplished primarily by uplift of mantle material contemporaneously with the return of impact ejecta and melt to the basin, yielding an initial structure similar to that of younger basins. Because of high near-surface temperatures, this initial structure undergoes long-term viscous relaxation of both topographic and Moho relief [Solomon et al., 1982].

Table 1. Estimates of Moho relief, apparent volume of mantle uplift, basin volume, and ejecta volume for major basins on the lunar nearside.

Basin	Radius, km	Moho relief, km	Ambient crustal thickness, km	V_u , 10^6 km^3	V_b , 10^6 km^3	V_e , 10^6 km^3
Orientale	310	66	85	6	0.7	7
Serenitatis	305	30	60	8	1	9
Crisium	225	31	55	6	1	7
Humorum	205	27	60	3	0.5	4
Nectaris	300	28	60	4	0.7	5
Smythii	300	33	65	4	0.6	5
Nubium	340	10	60	2	0.3	2
Fecunditatis	345	14	60	2	0.2	2
Tranquillitatis	340	12	60	2	0.2	2

Basins are listed in order of increasing age [Wilhelms, 1981]. The indicated radius corresponds to the second ring [Head, 1977; Wilhelms, 1981].

NEARSIDE BASINS

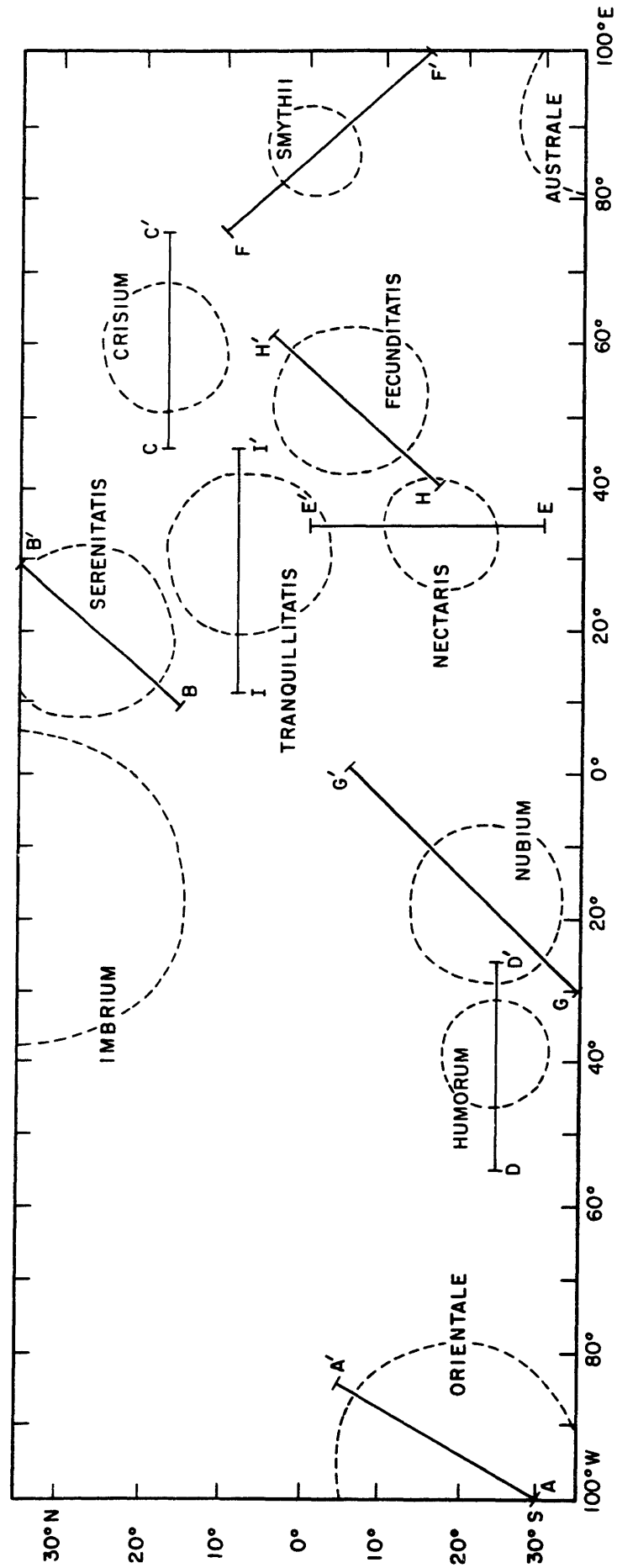


Figure 1

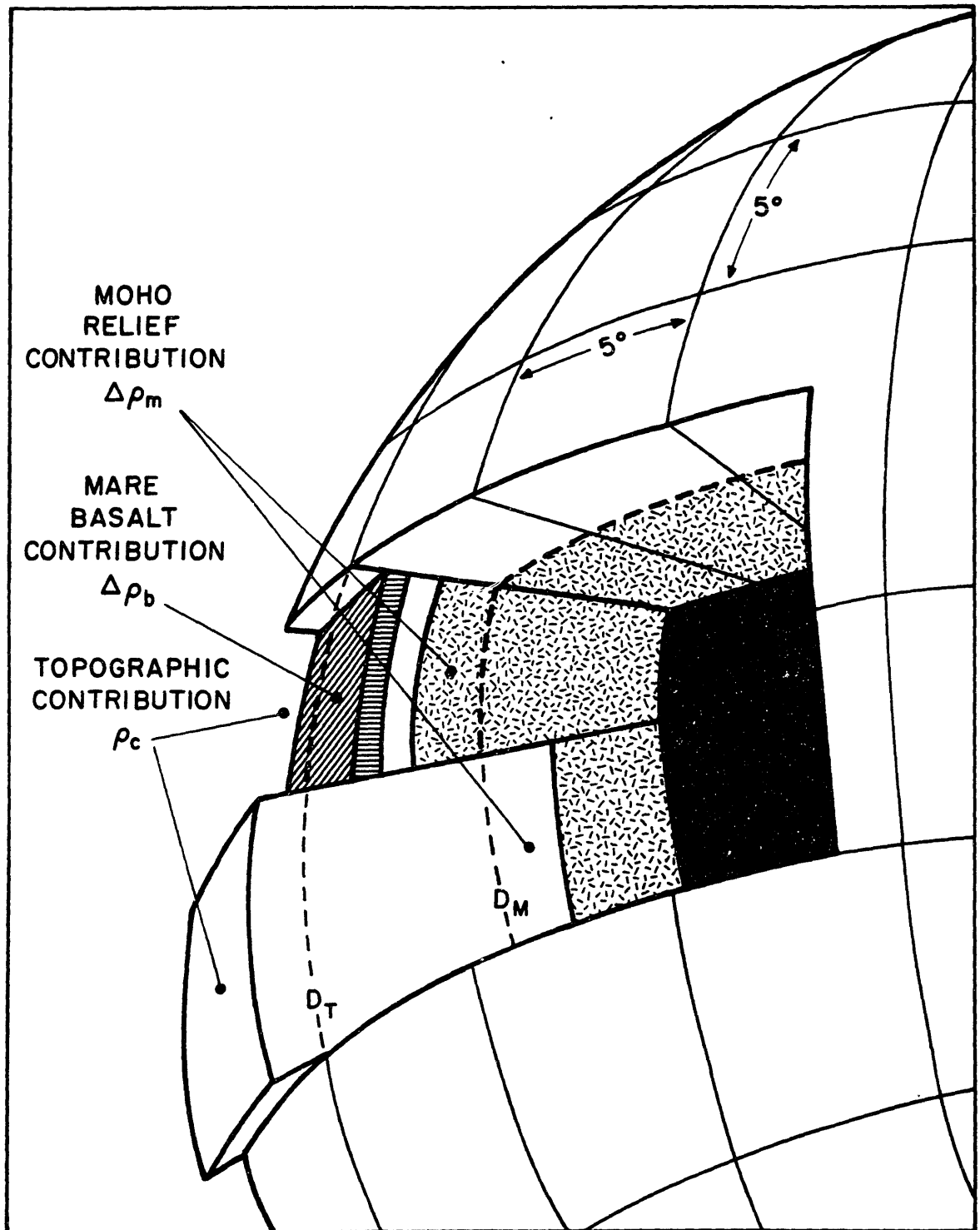


Figure 2

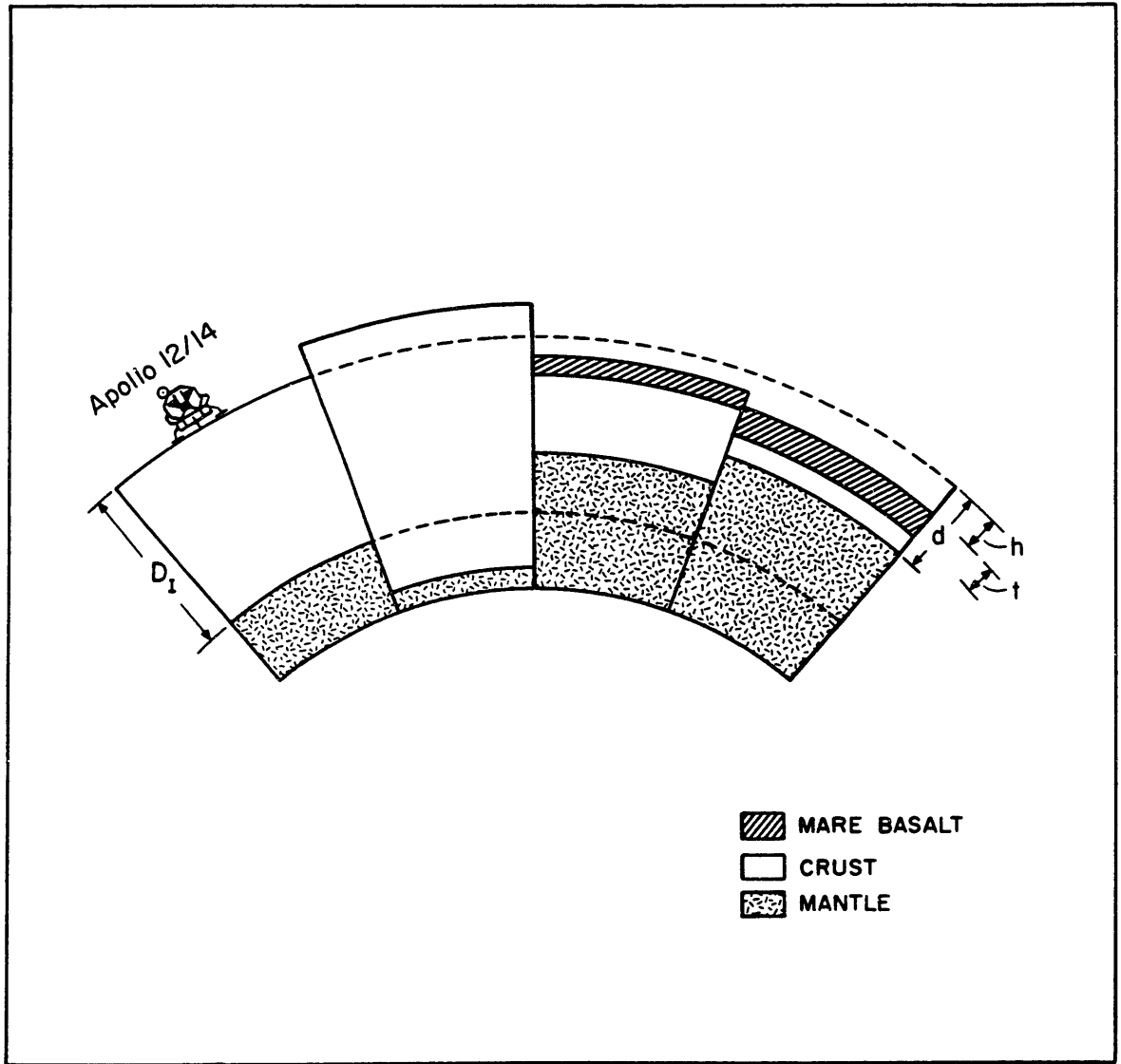


Figure 3

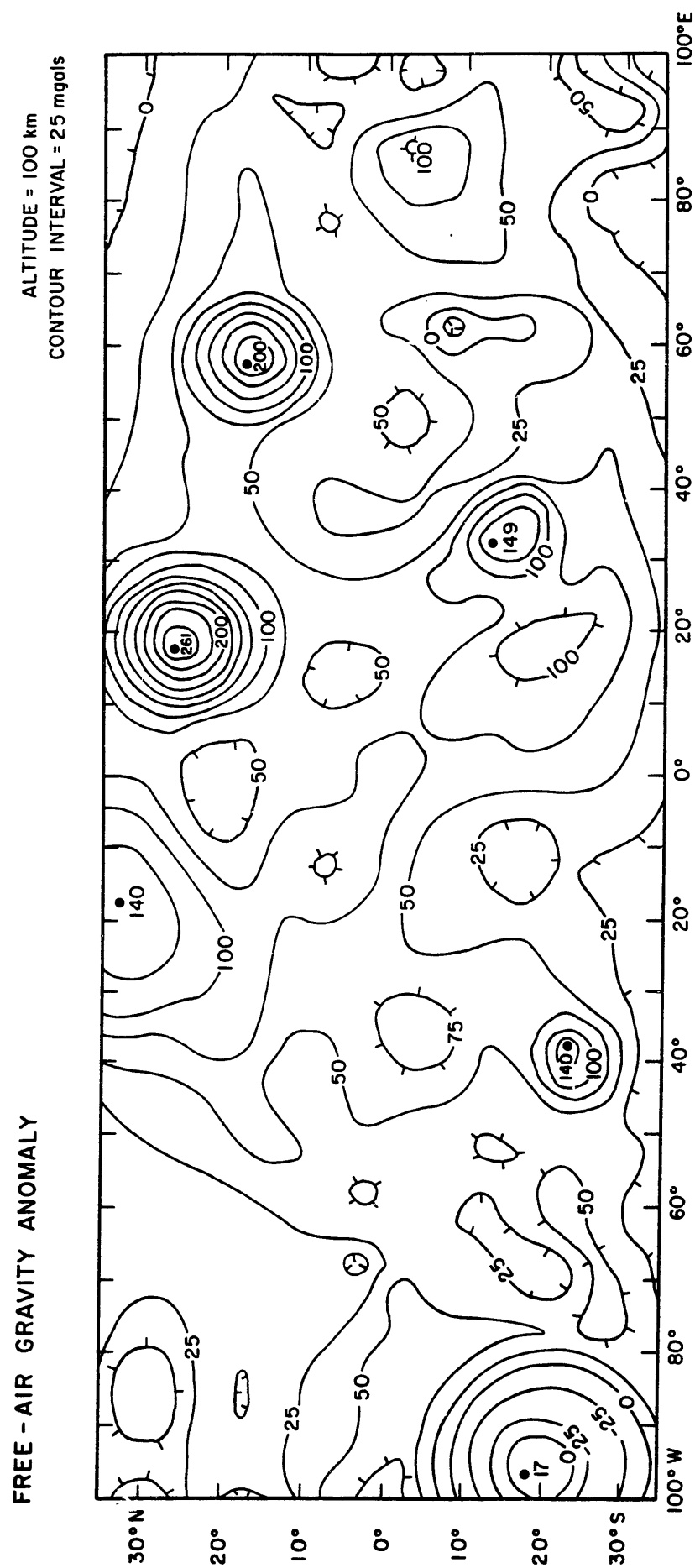


Figure 4

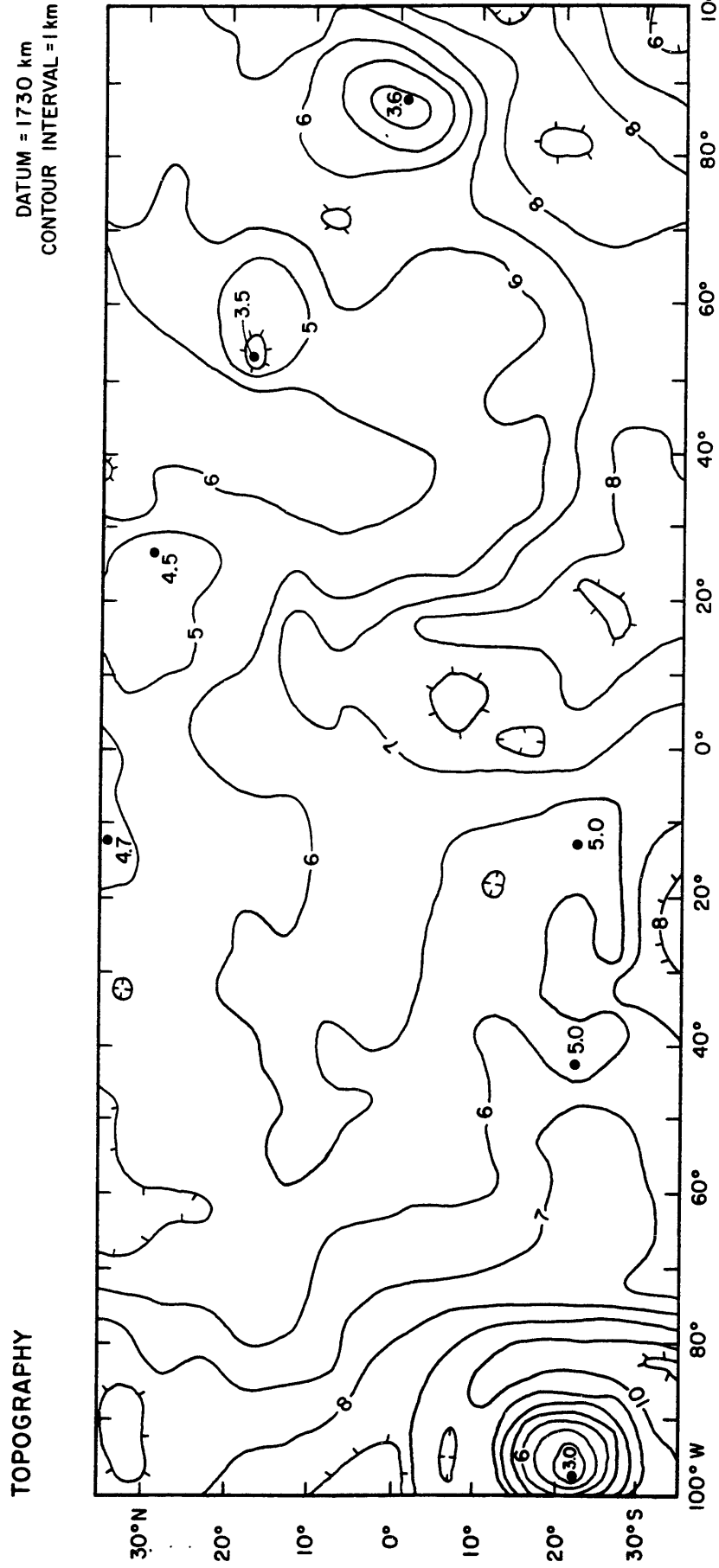


Figure 5

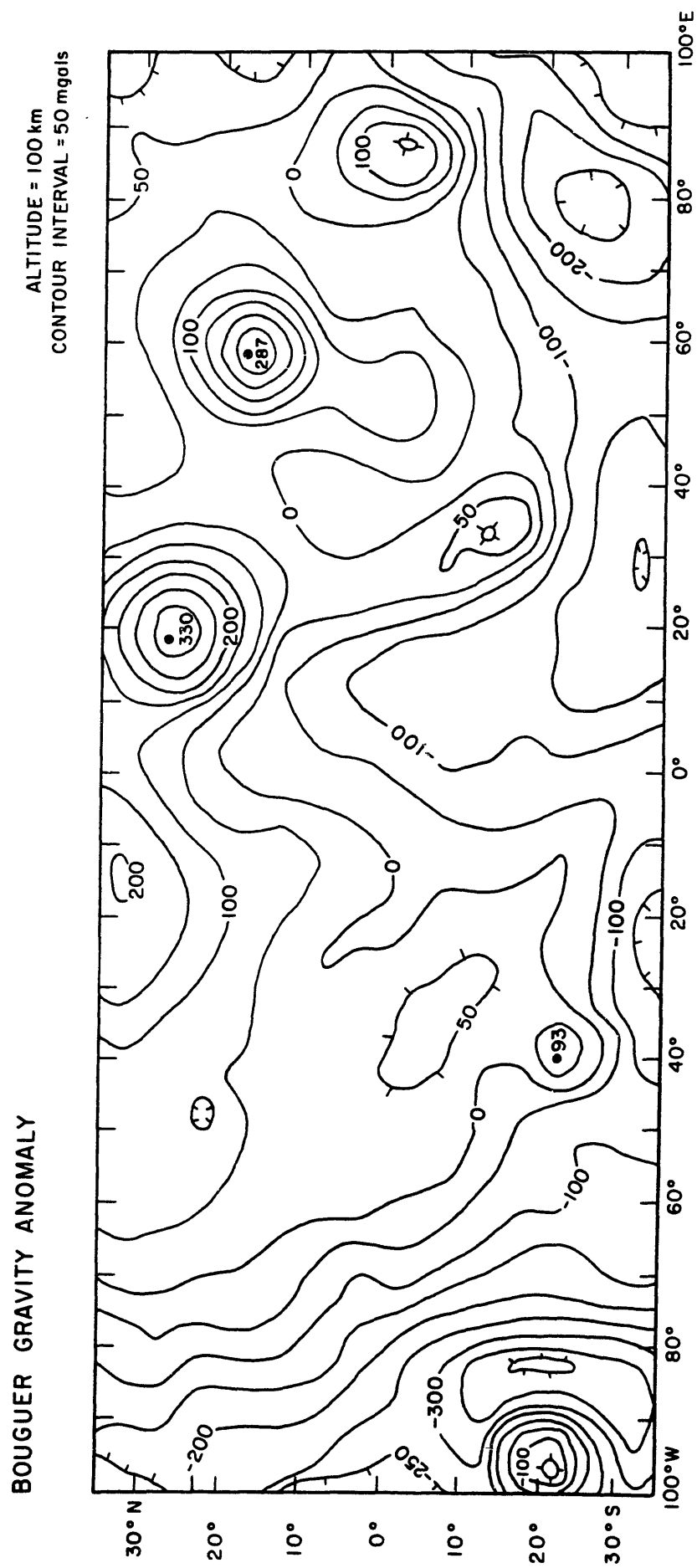


Figure 6

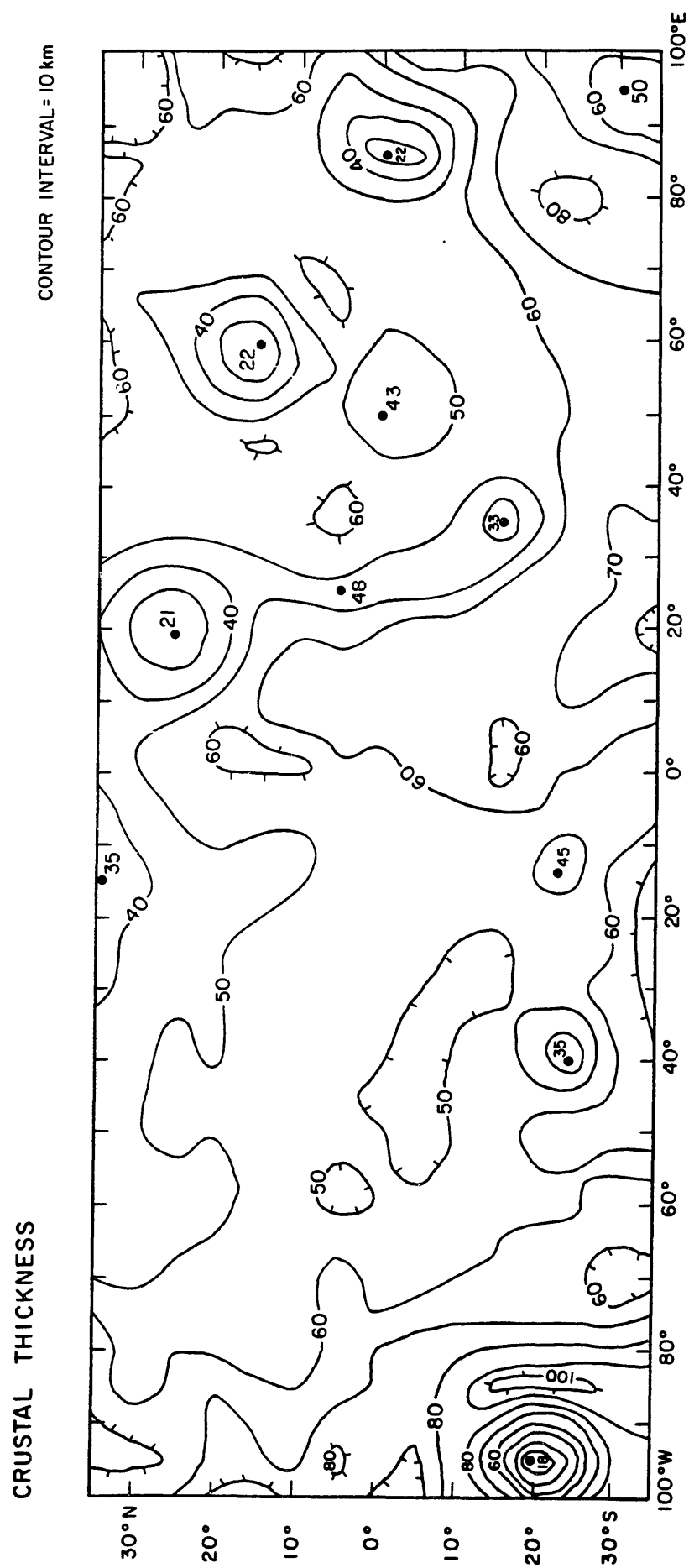


Figure 7

MARE BASALT THICKNESS

CONTOUR INTERVAL = 1 km

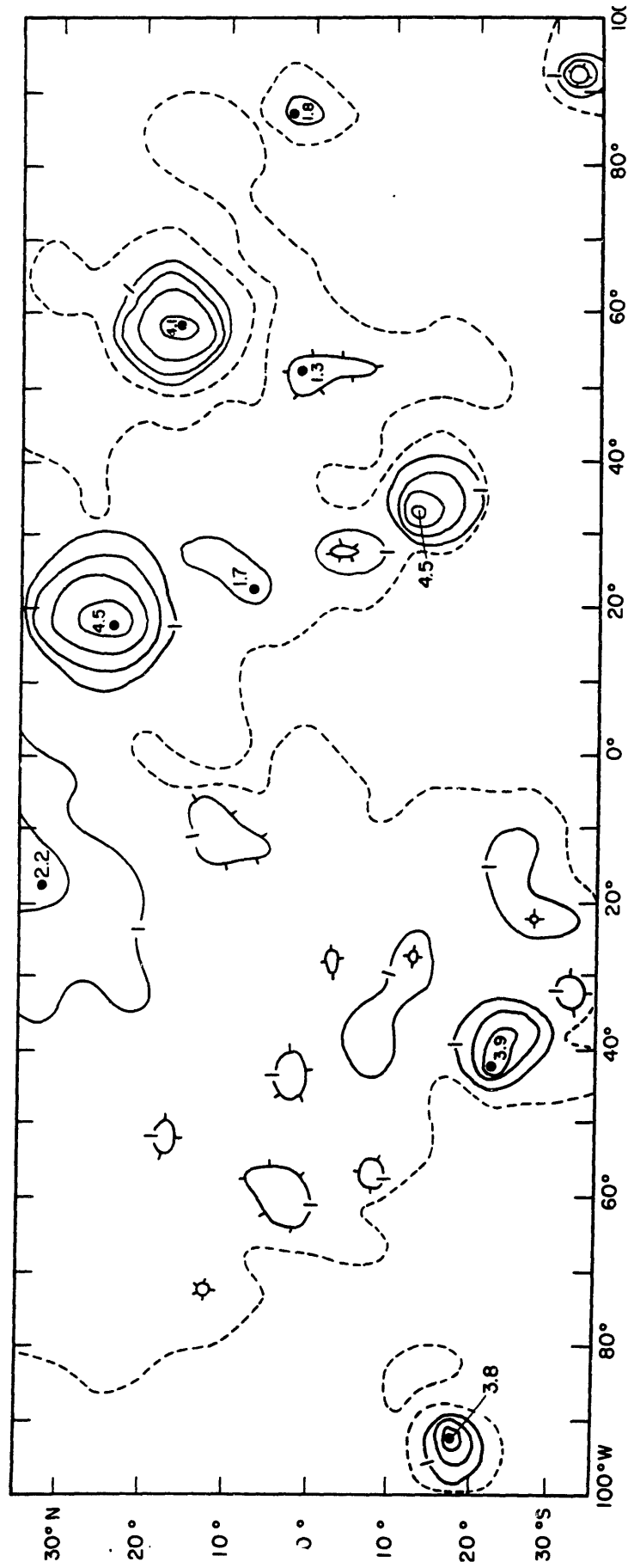


Figure 8

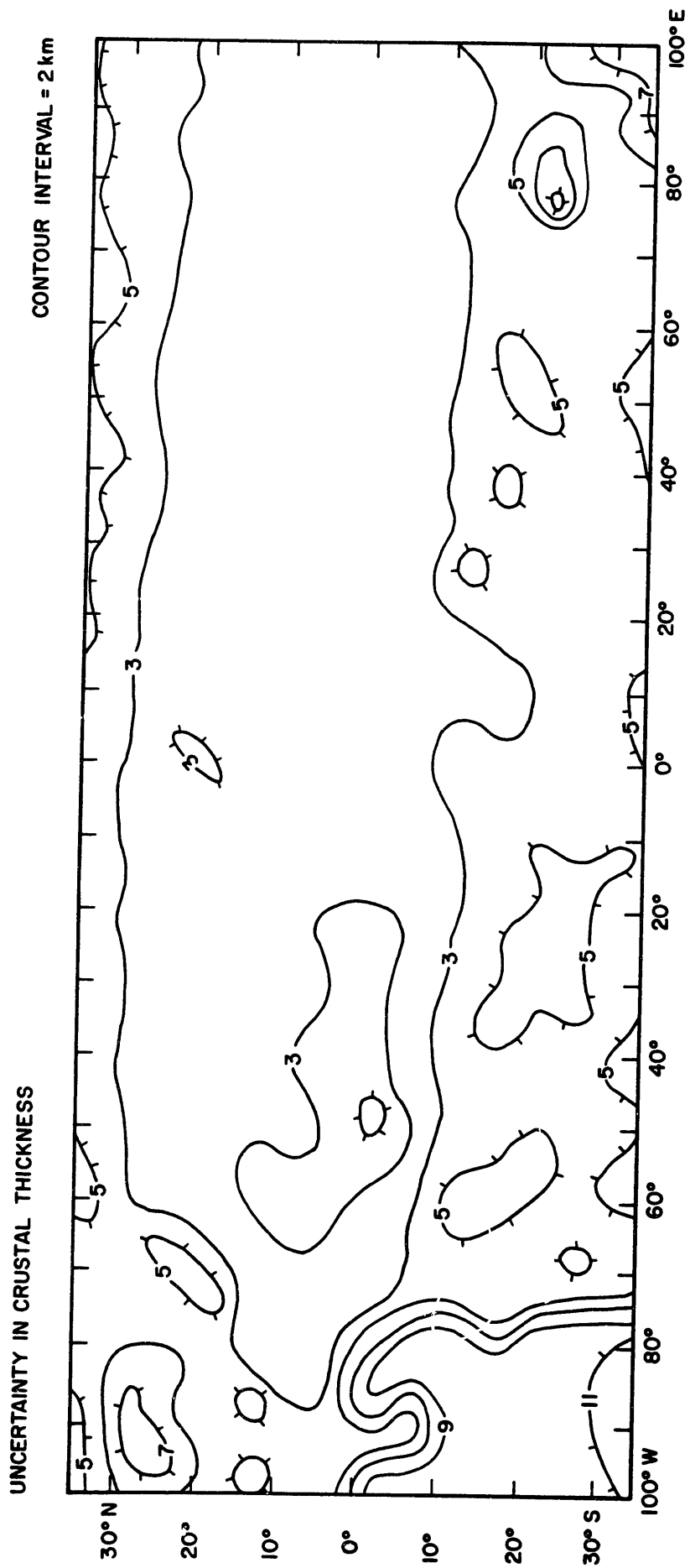


Figure 9

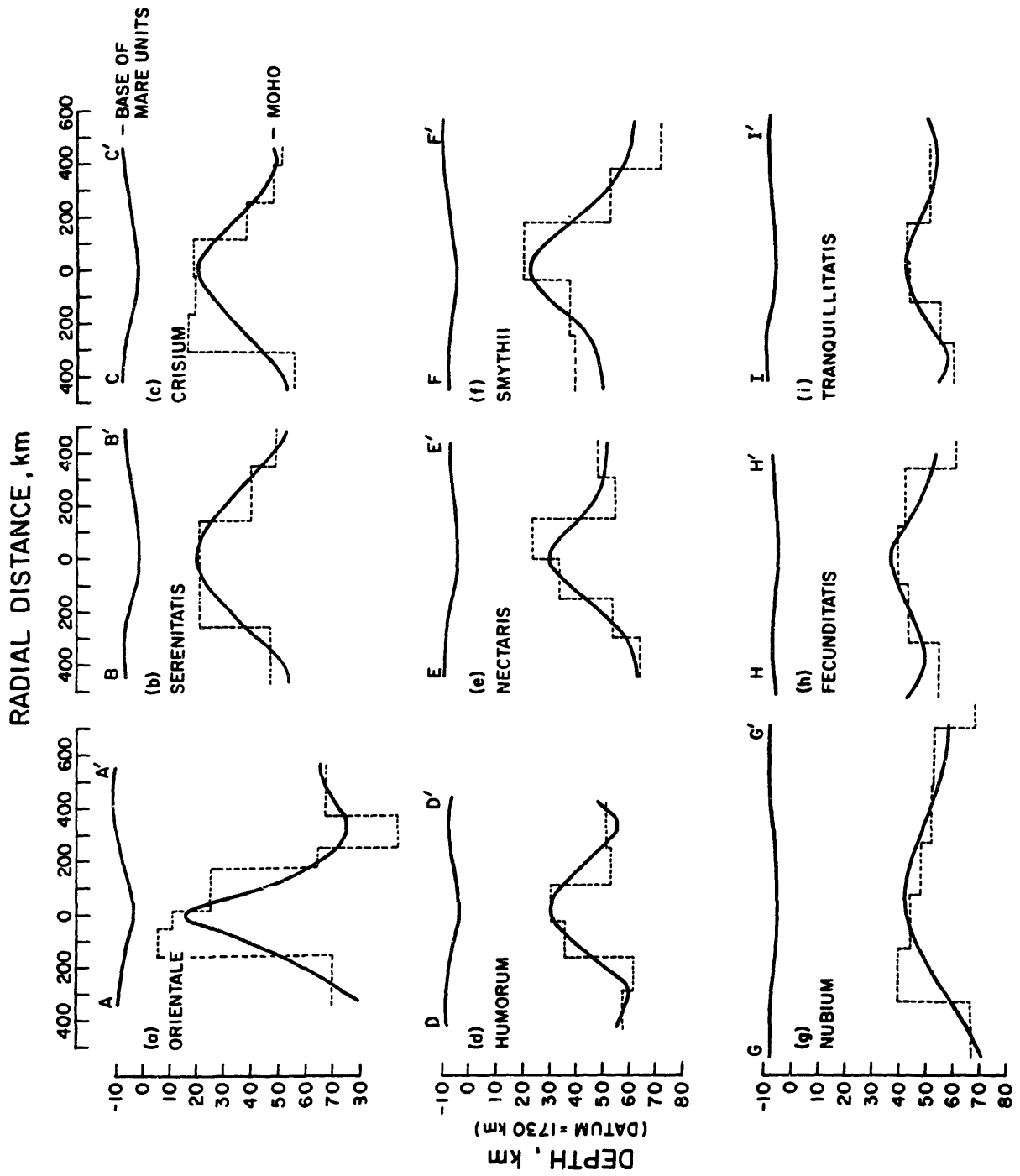


Figure 10

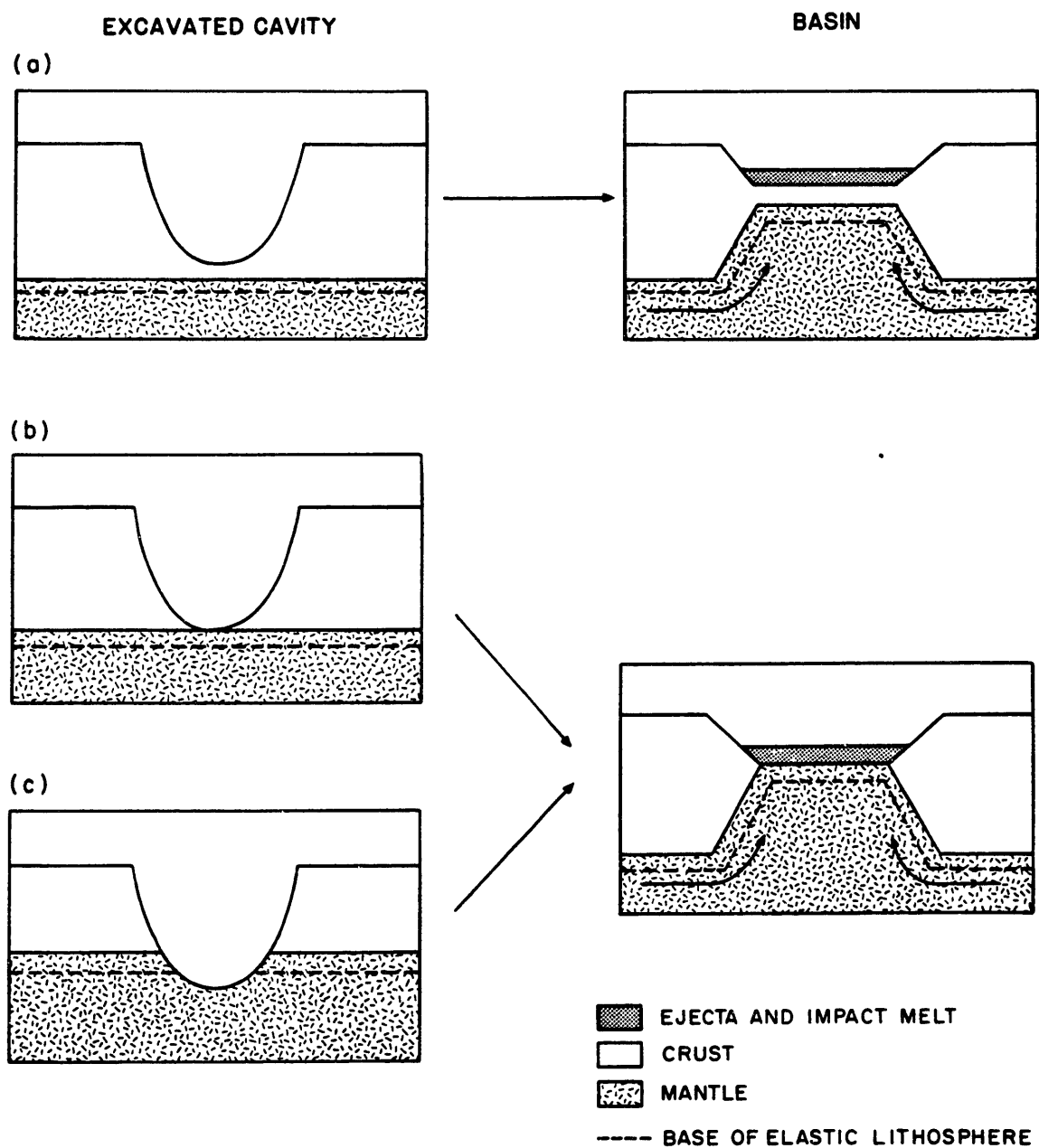


Figure 11

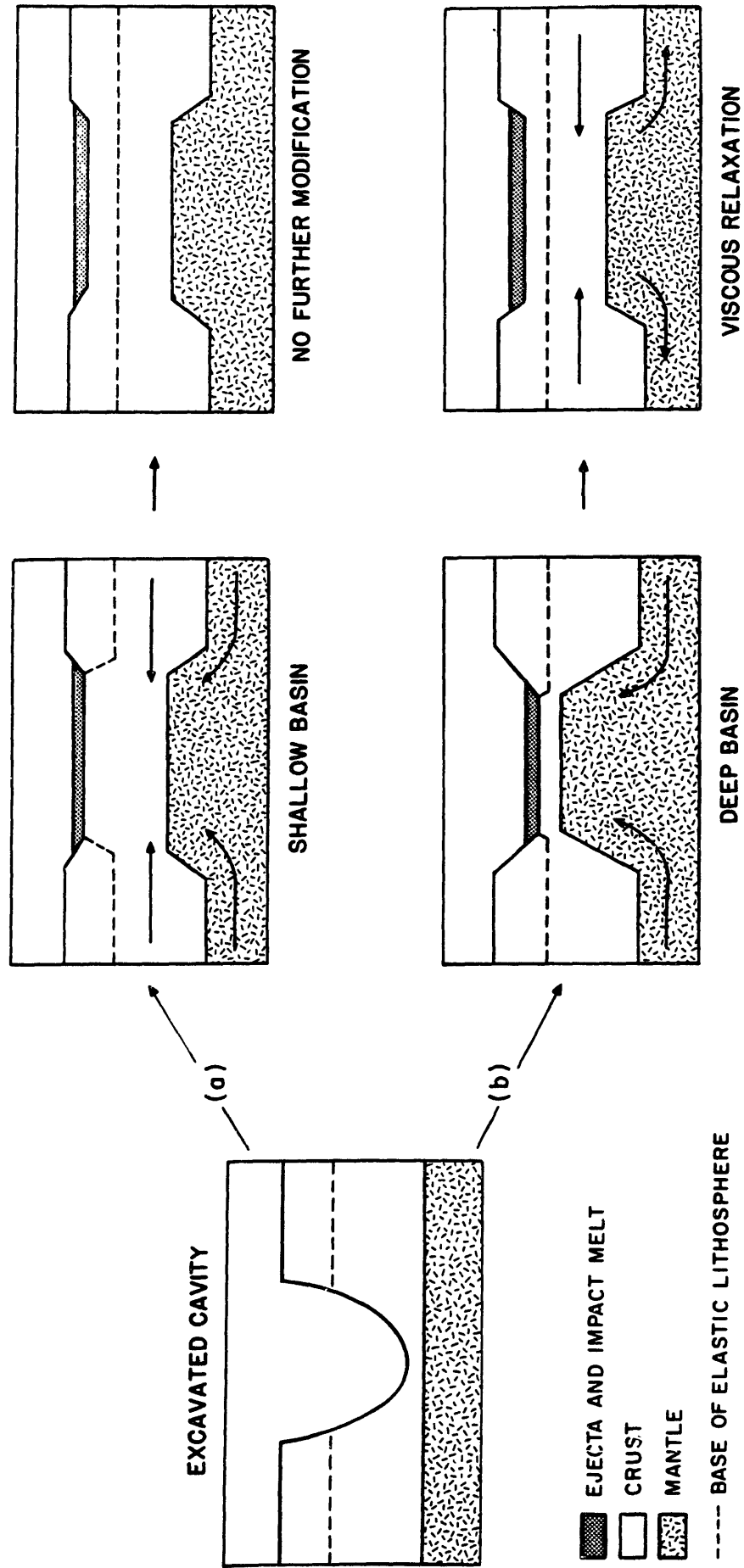


Figure 12

Chapter 6

THERMAL STRESS AND THE TECTONIC EVOLUTION OF LUNAR IMPACT BASINS

INTRODUCTION

Multi-ringed impact basins on the Moon exhibit wide variations in their present geometry and structure [Hartman and Wood, 1971; Wilhelms, 1973; Wood and Head, 1976]. Some of the variations may be related to differences in the properties of the lithosphere or impacting projectile at the time of basin formation [e.g. Melosh and McKinnon, 1978; Holsapple and Schmidt, 1982]. It is likely, however, that many of the presently observed variations reflect different degrees of modification of initial basin geometry and structure on time scales long compared to those for excavation and ring formation. For instance, the subdued surface and subsurface relief of basins formed early in lunar history when the lithosphere was relatively warm is probably a consequence of lateral ductile transport of crustal material over times scales of up to millions of years [Solomon et al., 1982; Chapter 5]. Ridges and graben associated with mare basins were probably formed in response to the load imposed by the emplacement of volcanic units perhaps 100 to 200 m.y. after basin excavation [Solomon and Head, 1979, 1980; Comer et al., 1979].

Another potentially important contributor to the long-term modification of a basin is thermal contraction and stress produced by the loss of heat from basin formation [Bratt et al., 1981]. An impacting projectile transfers some fraction of its original kinetic energy to the basin subsurface as deeply buried heat [O'Keefe and Ahrens, 1976, 1977]. In addition, the uplifted mantle preserved beneath the youngest near-side basins

[Chapter 5] suggests that uplift of the crustal and mantle isotherms also contributed to early heating beneath basins.

To better understand the morphology of impact basins, we assess here the significance of thermal stress as a tectonic process in the vicinity of lunar basins. Exploratory models are developed for the temperature structure created by the above mentioned sources of heat, the subsequent cooling of the basin subsurface, and the resulting thermal displacements and stresses as a function of time. The rate of basin cooling and the occurrence of greatest displacement and stress are compared to the values inferred from observations of the Orientale basin [Head, 1974; Church et al., 1982]. We then vary the quantity and distribution of heat in these models in an attempt to match observed topography and faulting patterns within the young, multi-ringed basin. A high degree of sensitivity of our models to input parameters permits us to use the observed topography and tectonics to derive rough constraints on the quantity and distribution of heat implanted during the impact process.

GEOLOGIC OBSERVATIONS: ORIENTALE

The Orientale basin (Figure 1), the youngest and best preserved of all lunar basins [Head, 1974; Moore et al., 1974], provides important information about the formation and modification of impact basins on all the terrestrial planets. Only the centralmost 220 km of the 900 km diameter topographic depression is extensively covered by mare basalts [Head, 1974], leaving exposed many geologic units and tectonic features now hidden beneath maria within other nearside basins. Because

Orientele was likely formed after other major basins on the Moon [Wilhelms, 1974], it has been left relatively unmasked by material ejected from other basins [Chapter 5]. The basin is, then, the optimum location to look for tectonic and topographic expressions of basin cooling and contraction.

Figure 2 shows a generalized topographic profile [Head et al., 1981] of Orientele along with the locations of major basin deposits and of features suggested by Church et al. [1982] to be of tectonic origin. The outer Rook Mountains, believed to represent the rim of the original impact cavity, form a ring about 620 km in diameter [Head, 1974, 1977]. Mare ridges and concentric rilles (graben) comprise the inner and outermost tectonic features, respectively, shown in Figure 2. Both features are thought to postdate the emplacement of most of the volcanic units and have been shown to be consistent with the pattern of stresses produced by mare loading and global thermal evolution [Solomon and Head, 1979, 1980; Comer et al., 1979]. No graben such as those found around Orientele and the other mare basins have been identified in the region of farside basins not filled by mare units. Thus, it is unlikely that the rille systems were formed in response to basin cooling, a process which would be expected to affect basins of a given age and size to a similar extent.

The plains and corrugated facies which surround and probably underlie the central maria are thought to represent cooled impact melt [Head, 1974]. Head [1974] and Church et al. [1982] suggest that the pitted and cracked texture of the

corrugated facies resulted from cooling and internal thermal contractions acting from the time of melt emplacement until its complete cooling [$\sim 10^3$ to 10^4 yr, Onorato et al., 1978].

Two features of Orientale not obviously related to basin formation, volcanic loading, or melt-sheet cooling but which may have been created or enhanced by local thermal contraction and stress are V-shaped fissures and the deep central depression [Church et al., 1982]. The band of intersecting fissures (Figures 3 and 4) at distances 150 to 230 km from the basin center differ in both plan and shape from the graben residing at 240 to 320 km radial distances. Whereas graben are roughly concentric with the center of the basin and have flat floors up to 4 km across, fissures trend both concentrically and obliquely with respect to the basin center and have V-shaped cross sections.

It is difficult, from the resolution of existing photographs, to infer the stress state responsible for fissuring within Orientale. Fissures cannot be classified as normal faults, because significant vertical displacement is not resolvable. One hypothesis is that fissures are analogous to the extension fractures formed normal to the direction of deviatoric extension in laboratory samples under low confining pressure [e.g., Brace, 1974]. That the majority of the fissures show opening toward the center of the basin (Figure 3) suggests that horizontal extension is primarily in the radial direction. The conjugate nature of a subset of the fissures located in the southwest (Figure 4) and northeast quadrants of

the basin is not unlike that of terrestrial conjugate shear zones such as those observed in continental extensional basins [e.g., Royden et al., 1983]. For strike-slip displacement to occur at the surface, the two horizontal stresses must be of opposite sign. Though we cannot resolve with certainty any horizontal displacement along fissures, we cannot discount the possibility that fissuring may be at least partially the product of strike-slip processes.

Because the youngest mare units flood portions of fissures, the fissures must predate at least the most recent mare volcanism. In addition, Church et al. [1982] suggest that fissures postdate cooling of the corrugated and plains facies (i.e., impact melt) in which they were formed. One reason for this view is that portions of the exposed melt sheet are not cut by fissures, suggesting that fissuring is the result of processes other than cooling and contraction of the melt sheet. These two temporal relations imply that if basin cooling was an integral cause of fissuring within Orientale, thermal stresses produced by basin cooling must have reached significant levels during a time interval after basin formation and before emplacement of the central maria 100 to 200 m.y. later [Greeley, 1976; Head and Solomon, 1980].

Limb-height measurements compiled by Head et al. [1981] provide information on the topography of the Orientale basin (Figure 2). Most impressive in these data is the great relief of the central depression of the basin relative to the encircling Rook and Cordillera Mountains. The total topographic relief is more than 9 km over a radial distance of

450 km from the center of the depression to the Cordillera ring. Relief from the center of the depression to the foot of the inner Rook Mountains about 180 km distant is almost 4 km. Photogeological evidence suggests that no more than 1 km of mare basalts fill the central depression [Head, 1974] as compared to the more than 4 km thickness of mare units which fill other young, nearside basins [Chapter 5]. Therefore, the pre-mare relief was probably not much greater than that observed today. It is likely that some fraction of the central depression is the result of thermal subsidence due to basin cooling. However, the size of this fraction is unconstrained. In this chapter, we model basin cooling and thermal tectonics in an attempt to explain a portion of the relief of the central depression, the fissuring at radial distances from 150 to 230 km, and the timing of observed tectonic deformation within Orientale.

THERMAL EVOLUTION

Models for the thermal evolution of an initially hot impact basin are based on the assumption of cylindrical symmetry. Thus, the temperature T at any time t is a function of the horizontal radius r and depth z and independent of azimuth θ . It is sufficient for our problem to solve for the case where the initial temperature is uniformly T_0 within a buried vertical cylinder extending from $r = 0$ to $r = a$ and from $z = h_1$ to $z = h_2$. Outside the cylinder, the temperature is zero. The solution for any circularly symmetric initial temperature distribution may then be approximated by a sum of

suitably weighted solutions for uniform cylinders. This geometry is illustrated in Figure 5. The solution for the temperature structure of a cooling cylinder in a half-space begins with the solution for the temperature in an infinite medium due to an instantaneous temperature change T_0 in a unit volume of material located at the origin [Carslaw and Jaeger, 1959]:

$$T(r,z;t) = \frac{T_0}{\pi^{3/2}c^3} e^{-(r^2+z^2)/c^2} \quad (1)$$

where $c = 2\sqrt{kt}$, and k is thermal diffusivity. We next apply the method of images to satisfy the boundary condition $T = 0^\circ\text{C}$ at the free surface and then integrate over a cylindrical volume of temperature T_0 to derive the following solution for a buried cylinder:

$$T(r,z;t) = \frac{T_0}{c^2} e^{-r^2/c^2} \int_0^a r' e^{-r'^2/c^2} I_0\left(\frac{2rr'}{c^2}\right) dr' \\ \left[\operatorname{erf}\left(\frac{h_2-z}{c}\right) - \operatorname{erf}\left(\frac{h_1-z}{c}\right) - \operatorname{erf}\left(\frac{h_2+z}{c}\right) + \operatorname{erf}\left(\frac{h_1+z}{c}\right) \right] \quad (2)$$

where erf is the error function [Gautschi, 1964] and I_0 is a modified Bessel function [Olver, 1964]. The integral in equation (2) is identically the P function described by Masters [1955].

THERMAL DISPLACEMENT AND STRESS

The cooling of a series of cylindrical bodies in a semi-infinite, elastic solid will produce radial and vertical displacements within that medium. Also, because cooling in the half-space is non-uniform, thermal stresses will accumulate in the elastic portion of the medium. To assemble models with

which we can assess the importance of thermal tectonics to basin modification, we use the method of thermoelastic displacement potentials [Goodier, 1937]. The response of an infinite, elastic solid to a field of temperature changes $\Delta T = T(r, z; t) - T(r, z; t - \Delta t)$ can be represented by a distribution of centers of contraction (or dilatation) of magnitude:

$$\beta = \frac{\alpha \Delta T (1+\nu)}{12 \pi (1-\nu)} \quad (3)$$

where α is the volumetric coefficient of thermal expansion, and ν is Poisson's ratio. The solution of the thermal stress problem in a semi-infinite solid requires that the surface be free of shear and normal stresses. To satisfy these boundary conditions, Mindlin and Cheng [1950] modified the formulation of Goodier [1937] by imposing a center of contraction, a double force, and a doublet, each of appropriate strength at the image point $(r, -z)$ corresponding to each center of contraction (r, z) in the half-space. As a result, the thermal displacement field $\underline{u} = (u, v, w)$ is given by

$$\underline{u} = - \nabla \phi_1 - \nabla^2 \phi_2 \quad (4)$$

where the thermoelastic displacement potentials in the axisymmetric problem are:

$$\phi_1 = \iiint_V \frac{\beta}{R_1} r' dr' d\theta' dz' \quad (5)$$

$$\phi_2 = \iiint_V \frac{\beta}{R_2} r' dr' d\theta' dz' \quad (6)$$

are integrals over volume V in the half-space weighted by the distance scalars:

$$R_1' = [(r-r')^2 - 2rr' \cos(\theta-\theta') + (z-z')^2]^{1/2} \quad (7)$$

$$R_2' = [(r-r')^2 - 2rr' \cos(\theta-\theta') + (z+z')^2]^{1/2} \quad (8)$$

The operator applied to ϕ_2 is:

$$\nabla^2 \phi_2 = (3-4\nu)\nabla^2 \phi_2 + 2\nabla[z (\frac{\partial \phi_2}{\partial z})] - 4(1-\nu)\nabla^2(z\phi_2)\underline{e_k} \quad (9)$$

where ∇^2 is the Laplacian and $\underline{e_k}$ is the unit vector in the z direction.

Because of the symmetry of the time-dependent temperature solution (equation 2), we can solve for the total displacement field by approximating the temperature change field $\Delta T(r,z;t)$ by a set of cylinders of uniform temperature change, computing the displacements due to cooling of individual cylinders, and summing the resulting displacement solutions (see Figure 5). The calculation of displacements from equation (4) reduces to differentiation if the potential of the distribution of temperature changes is known. Unfortunately, a solution for the potential surrounding a cylinder does not condense to a simple equation. We follow the work of Thurber (1980, personal communication) and Moon [1981] to develop a suitable equation for the potential of an infinitely thin disk. The details of this derivation are given in Appendix C. We then integrate the disk potential vertically over z' to obtain the thermoelastic displacement potential ϕ_1 for a cylinder of finite thickness and constant temperature. The displacement potential ϕ_2 is computed by substituting the quantity $(z+z')$ for $(z-z')$ in the solution for the potential of a disk prior to integration over z' . The displacement field follows directly from centered difference numerical differentiation of ϕ_1 and ϕ_2 as prescribed by equations (4) and (9).

The accuracy of the above described method for computing

the potential of a cylinder was tested by constructing a sphere using a stack of buried cylinders. We calculated the resulting potential and compared the solution to the well known potential of a sphere. The procedure was repeated with spheres of various radii. Because we are primarily concerned in the models discussed below with the displacements and stresses near the free surface, it is essential that the approximation for the potential of a cylinder be accurate when the cylinders are close to the observation points at the surface. Even when the sphere is placed within 0.02 sphere radii of the surface, the potential of the stacked cylinders approximating the sphere can be brought as close as desired to the analytic solution for the potential of a sphere by decreasing the thickness and increasing the number of cylinders in the representation. We also calculated the displacement field produced by the same sets of stacked cylinders and compared the results to the displacements at the free surface due to a cooling sphere given by Mindlin and Cheng [1950]. Again, the difference between our approximation and their exact solution can be made to vanish by using thinner cylinders. We are, thus, confident that the summation of solutions for the displacement fields computed using the approximation for a cylindrical potential (Appendix C) and equation (4) can adequately represent displacement field due to cooling and contraction of any axisymmetric temperature structure.

In the models that follow, we examine the displacements and stresses only at the surface of the basin. Thus, the stress normal to the surface (σ_{zz}), all shear stresses, and the

local temperature are all zero. The field of thermoelastic stresses produced at the free surface is calculated from the displacement field (equation 4) using the formulae [Timoshenko and Goodier, 1970]:

$$\sigma_{rr} = - \frac{E}{(1+\nu)(1-2\nu)} [(1-\nu)\epsilon_{rr} + \nu(\epsilon_{\theta\theta} + \epsilon_{zz})] \quad (10)$$

$$\sigma_{\theta\theta} = - \frac{E}{(1+\nu)(1-2\nu)} [(1-\nu)\epsilon_{\theta\theta} + \nu(\epsilon_{rr} + \epsilon_{zz})] \quad (11)$$

where E is Young's modulus, compressive stress is positive, and strains are given by:

$$\epsilon_{rr} = \frac{\partial u}{\partial r} \quad (12)$$

$$\epsilon_{\theta\theta} = \frac{u}{r} \quad (13)$$

$$\epsilon_{zz} = \frac{\partial w}{\partial z} \quad (14)$$

where u and w are the radial and vertical components of displacement and differentiation is accomplished numerically using the centered difference method. The stresses examined in this chapter are measured relative to a zero-stress state at $t = 0$. Though the state of stress soon after the impact may contain contributions from basin formation and cavity collapse, we concentrate here on stress produced by basin cooling.

ASSESSMENT OF PROCEDURE

The equations derived above provide a simple procedure with which the elastic effects of an axisymmetric temperature change in a half-space can be computed. Most uncertain in our models is effect of the assumption that the lunar lithosphere behaves as an elastic half-space during the thermal evolution

of a basin. The stress field calculated by the aforementioned method will be formally correct only if the material undergoing temperature changes ΔT behaves elastically. It is likely, however, that at high temperatures stress will be rapidly relieved by ductile flow. From the depth distribution of earthquakes away from plate boundaries, Chen and Molnar [1983] estimate that temperature limits to which the Earth's crust and upper mantle will deform seismically are 250°-450°C and 600°-800°C, respectively. McNutt and Menard [1982] conclude that the base of the elastic lithosphere beneath the oceans is near the 550°C isotherm. The lunar thermal profile (selenotherm) is highly uncertain. The lunar thermal history model of Solomon and Head [1979] indicates that only the upper 25 km of crust was below 250°C, and that little or none of the mantle was cooler than 500°C near the time of formation of Orientale (~ 3.9 b.y. ago). In addition, the anomalous heat related to basin formation further reduces the volume of subsurface material capable of sustaining stress. However, caution should be exhibited in transferring to the Moon information on the depth extent of elastic behavior in the Earth. The selenotherm and the rheology of the deep lithosphere at any given time are not well understood. One certainty in our problem is that the material near the lunar surface will cool rapidly after basin formation. Therefore, the upper crust will accumulate elastic stress (and episodically undergo brittle deformation) produced by cooling within the elastic lithosphere and transmitted from material

cooling and contracting at depth.

In the half-space model presented here, it is impossible to examine completely the effects of depth- and temperature-dependent rheology on the strains and stresses at the surface. We can simulate part of the anelastic effects of high temperature regions on the stress field by stipulating that material not be permitted to contribute to thermal stress until it has cooled below some elastic blocking temperature T_e [Turcotte, 1974, 1983]. The details and limitations of incorporating a blocking temperature in models of cooling oceanic lithosphere are described in Chapter 5. In the models that follow, we present cases with and without a blocking temperature.

INITIAL THERMAL STRUCTURE

The impact of a basin-forming projectile with a planet (Figure 6) involves a transfer of projectile kinetic energy primarily to heating of the target area and secondarily to ejection of some portion of heated target material [e.g. O'Keefe and Ahrens, 1976, 1977]. The deep structure of the youngest nearside basins [Chapter 5] suggests that the excavated impact cavities of basins the size of Serenitatis or Orientale penetrated to near or beneath the lunar Moho. The excavated volumes of these basins were probably on the order of 10^7 km^3 . Collapse of the cavity walls and uplift of the crust, upper mantle, and associated isotherms followed shortly after basin formation [Head, 1974]. Thus soon after the impact event, the basin subsurface was subjected to two primary

sources of anomalous heating above the ambient lunar temperature structure: (1) conversion of impact kinetic energy to deeply buried heat (impact heating), and (2) uplift of crustal and mantle isotherms during the short-term modification of the excavated cavity (isotherm uplift). We consider both sources of heat, and we derive simple relationships for the quantity and distribution of anomalous temperature implanted beneath a large basin. We then use these temperature structures to explore the contribution of cooling tectonics to the modification of a basin the size of Orientale.

Impact Heating

The quantity and distribution of heat emplaced during a hypervelocity impact, as well as the size of the basin remaining after excavation and short-term modification, are complex functions of the mass, volume, and velocity of the projectile and of the gravity, density and strength near the planet's surface [e.g., Holsapple and Schmidt, 1982]. Given only the observed basin morphology, it is difficult to estimate the original kinetic energy (E_K) of the impacting body. Many of our present conceptions about the energetics and mechanics of basin formation follow from scaling of smaller craters formed by ancient impacts on the Earth [Shoemaker, 1960; Grieve et al., 1977], nuclear and chemical explosions in the field [e.g., Vaile, 1961; Nordyke, 1977] and impact experiments in the laboratory [e.g., Gault and Wedekind, 1977; Schmidt, 1977]. Valuable information on cratering has also been derived from analytical and numerical simulations of crater formation [e.g., Gault and Heitowit, 1963; Maxwell, 1977; O'Keefe and Ahrens,

1976, 1977].

The crater produced by the Teapot-Ess nuclear explosion is considered a good terrestrial analog of larger impact craters because the overturned strata at the crater rim resemble rim formations observed at the Meteor impact crater [Shoemaker, 1960]. The crater formed by the explosion was 89 m in diameter (D_T) and 27 m deep [Nordyke, 1961]. The yield (E_T) of the event was 5×10^{19} erg. To obtain a rough estimate of the energy (E_K) required to form an Orientale-sized basin, we assume the crater radius is a function of gravity and kinetic energy only and scale the Teapot-Ess data using the relation:

$$E_K = E_T \left[\frac{D_C}{D_T} \left(\frac{g_E}{g_M} \right)^n \right]^m \quad (15)$$

where g_E and g_M are the gravitational acceleration of the Earth (980 cm/s^2) and Moon (162 cm/s^2), respectively. Values for the exponent m range from about 3 (volume proportional to energy) to 4 [Holsapple and Schmidt, 1982] and, for n , from 0.14 [Ivanov, 1976] to 0.25 [Holsapple and Schmidt, 1982]. Using the extreme values of m and n produces estimates of E_K varying by over four orders of magnitude. We adopt $m = 3.4$ in accord with energy-diameter relationships from craters formed by nuclear and chemical explosives [Vaile, 1961]. We set n equal to 0.165 after the work of Gault and Wedekind [1977], who examined the effect of varying downward acceleration ('gravity') in laboratory experiments using aluminum projectiles impacting quartz sand. Adopting an impact crater

diameter (D_c) of 610 km for Orientale [Head, 1977], equation (15) suggests that E_K for that basin is about 2×10^{32} erg. We use this number only as a starting point with which the thermal evolution of Orientale can be modeled.

Numerical models of small impacts ($E_K \sim 10^{16}$ erg) performed by O'Keefe and Ahrens [1976] suggest that more than 90% of E_K (call this quantity E_H) goes into heating the projectile and target, that perhaps 25% of E_K never leaves the basin, and that some unknown fraction of heated ejecta returns to rest within the basin. It is possible that less heat is ejected from the target region following a basin-size impact because excavation of heated target material may have been impeded by an increase in strength across the lunar Moho [Chapter 5].

The spatial distribution of anomalous temperatures produced by shock heating under a large basin is largely unconstrained. Though the distribution of impact-related energy density within the target region can be derived from cratering models [e.g., Gault and Heitowit, 1963; O'Keefe and Ahrens, 1975], little consideration has been given to recovering the temperature field from subsurface energy density. For simplicity, we assume that the implanted heat per unit volume decreases exponentially with distance within a hemi-spherical volume centered at the point of impact on the pre-impact surface [Kaula, 1979]. This geometry is shown in Figure 6. An exponential decay of impact heat density is partially supported by the geologic observation that shock

metamorphism and melting is concentrated at shallow depths below terrestrial craters [e.g. Dence, 1971; Grieve and Cintala, 1981] and by the roughly exponential relationship between the internal energy density and depth in finite-difference models of the formation of the Imbrium basin [Figure 5, O'Keefe and Ahrens, 1975].

We compute the spatial distribution of impact-generated temperature as a function of slant-range distance $q = (r^2 + z^2)^{1/2}$ from the center of symmetry on the pre-impact surface using the relation:

$$T(q) = \frac{\epsilon e^{-q/s}}{\rho C_p} \quad (16)$$

where s is a decay constant (the distance from the center of symmetry at which heating falls to $1/e$ of its peak value), ρ is density (either crustal or mantle), and C_p is specific heat. The latent heat of phase changes is ignored here. The quantity ϵ is an energy density derived by solving the integral equation

$$E_H = \int_0^L (2\pi q^2) \cdot (\epsilon e^{-q/s}) dq \quad (17)$$

where L is the value of q at which temperatures drop below some arbitrarily small level of significance. The energy density at radius q is given by the terms between the second set of parentheses.

Some fraction of this implanted heat is ejected during basin excavation. To estimate the amount of heat thrown beyond the basin rim, we utilize the model for nearside crustal

structure derived in Chapter 5. The difference between the assumed pre-impact crustal thickness and the thickness of non-mare crust beneath the youngest basins such as Orientale provides a lower bound on the depth from which material was permanently excavated from the basin. We remove from the top of the target region a plug of heated material of thickness equal to the apparent depth of excavation (Figure 6). Left behind is a thickness of crust equivalent to that inferred from the analysis of Chapter 5 and a spherical cap of heated crust and upper mantle. Let E_B be the quantity of impact heat residing beneath the basin. The quantity E_B and the decay constant s will be the important parameters in the models that follow. We assume that uplift of the basin subsurface in response to the mass deficiency created by cavity excavation raised the hot material along vertical trajectories. Though the return of heated ejecta probably contributed to the post-impact thermal structure of basins, we assume here that the contribution of returned ejecta is small in comparison to that of shock heating of non-excavated target material.

Isotherm Uplift

Uplift of crustal and mantle isotherms accompanied the uplift of shock heated material during the short-term modification of a basin. The extent of uplift of sub-basin isotherms may be estimated from the preserved relief of the lunar Moho beneath young basins (Figure 10, Chapter 5), a basin age, and a global thermal history model. The effects of isotherm uplift are illustrated in Figure 7. Isotherm uplift

after basin formation is assumed to follow vertical trajectories, with the maximum uplift occurring in the central region of the basin. The shape of the uplifted volume (Figure 6) is taken to be that of a truncated cone with upper and lower radii as inferred from the crustal structure models of Chapter 5. The pre-impact temperature distribution is taken to be constant below 100 km depth. The anomalous temperature distribution is calculated as the difference between the uplifted temperature profile and the pre-impact selenotherm. The temperature changes contributed by isotherm uplift are maximum at the surface and in the basin center and go to zero at 100 km depth.

ADOPTED PHYSICAL PARAMETERS

The choice of physical parameters affects the rate of cooling and the magnitude of thermal displacements and stresses in the thermally and tectonically evolving basin. Measurements of the physical properties of lunar samples are, however, few in number. Also the pressure and, especially, temperature conditions during the time of basin cooling are uncertain.

The densities of the crust (ρ_c) and mantle (ρ_m) used here are 2.9 and 3.4 g/cm³, respectively, and are probably accurate to ± 0.1 g/cm³ [Solomon, 1975]. We adopt a specific heat of 1.2×10^7 erg/g-°C [Solomon and Longhi, 1977]. Density and specific heat affect the conversion from shock heat density to temperature (equation 16) and probably pose the least important source of error in this analysis. For diffusivity (k) we use the value of .01 cm²/s measured by Schatz and Simmons [1972] at

180°C in olivine. At about 1200°C, k in olivine drops to half this value. Because the cooling time of a body is approximately inversely proportional to k , our results may overestimate by as much as a factor of two the rate of cooling at early times when sub-basin temperatures are greatest. We set the volumetric coefficient of thermal expansion α to $2.7 \times 10^{-5} \text{ } ^\circ\text{C}^{-1}$ throughout the cooling region. This value lies between the average value of $1.4 \times 10^{-5} \text{ } ^\circ\text{C}^{-1}$ measured by Baldrige and Simmons [1971] in porous lunar basalts and breccias between 25 and 200°C at atmospheric pressure and an average of $3.7 \times 10^{-6} \text{ } ^\circ\text{C}^{-1}$ from measurements at olivines made at 1 atm and 800°C and given by Skinner [1966].

As may be seen from equations (10) and (11), thermoelastic stress calculated is proportional to Young's modulus (E). Both static (strain measurements of rocks subject to pressure changes) and dynamic (ultrasonic measurements of compressional and shear velocity) experiments provide information on E . At higher pressures, dynamic measurements generally result in values of E higher than statically derived values [Simmons and Brace, 1965]. The calculation of thermal stress in this chapter is made at the lunar surface; thus a consolidated breccia would be a suitable lunar sample to use as an analog for the battered upper crust. We use Mizutani and Osako's [1974] ultrasonic measurement of compressional velocity (5.5 km/s) in Apollo 17 metabreccia (sample 73235, $\rho = 2.93 \text{ g/cm}^3$) to compute E . Assuming a Poisson's ratio of 0.25, $E = 7.4 \times 10^{11} \text{ dyne/cm}^2$. At 0.5 kbar confining pressure ($\sim 10 \text{ km depth}$)

$E = 8.1 \times 10^{11}$ dyne/cm². Though no static measurements of compressibility of this metabreccia have been made, the work of Simmons and Brace [1965] suggests that a static determination of E would be somewhat lower than Mizutani and Osako's [1974] values because of the different effects of unclosed cracks on velocity and strain measurements at low (< 2 kbar) confining pressure. In the models presented below, we select $E = 7 \times 10^{11}$ dyne/cm² as characteristic of the uppermost crust. A summary of physical parameters adopted in this chapter is given in Table 1.

THERMO-TECTONIC MODELS: ORIENTALE

Following the above guidelines, we present models for the contributions of isotherm uplift and impact heating to the heat emplaced beneath Orientale (Figure 1) during basin formation and short-term modification. A summary of the models presented below and the important parameters used in their computation is given in Table 2.

Isotherm uplift

The results of Chapter 5 provide constraints on the quantity of heat contributed by uplift of the crust, upper mantle, and associated isotherms following cavity formation. The shape of the Moho beneath Orientale as inferred from an inversion of lunar gravity and topographic data (Figure 10a, Chapter 5) is that of a truncated cone with an upper radius of 50 km, a lower radius of 310 km, and a height of 55 km. Using the estimated age of the basin (~ 3.9 b.y.), a global thermal history model [Solomon and Head, 1979], and this Moho relief,

the quantity and distribution of anomalous temperatures produced by isotherm uplift can be estimated. Figure 7 shows the pre-impact and post-uplift temperature profiles under the center of Orientale as well as the profile of anomalous temperature ΔT taken to be the difference between these two curves. The quantity ΔT at the lunar surface is nearly 600°C . The temperature change at 100 km depth is zero by assumption. Figure 8a shows the field of anomalous temperatures produced by isotherm uplift (model A) as a function of r and z . At radii greater than 50 km from the basin center, structural and isotherm uplift is diminished relative to that beneath the central basin region. The total anomalous heat contributed by isotherm uplift beneath Orientale is 1.4×10^{32} erg, a number similar in magnitude to estimates of the of the implanted impact energy for basins of this size.

We derive a thermal history for the uplift heating contribution to sub-basin heating, as we will for all subsequent models presented, by first approximating the initial temperature distribution shown in Figure 8a by a series of cylinders of constant temperature. We then compute the cooling histories of each cylinder using equation (2) and superpose the solutions to complete the problem. The resulting temperature fields at times 10, 100, and 500 m.y. after the initial uplift of isotherms ($t = 0$) are displayed in Figures 8b-8d. By 10 m.y., much of the heat in the upper 20 km of model A has left the basin region. By 100 m.y. only about 30% of the initial heat remains. By 500 m.y. less than 1% of the heat is

left. Thus most of the thermal contraction and stress contributed by isotherm uplift will take place within 100 m.y. of basin formation.

The major uncertainties in the cooling that follows isotherm uplift are the adopted pre-impact selenotherm and the extent of uplift. It is difficult to assess error in the adopted selenotherm, but the error in amount of uplift under Orientale is constrained by the uncertainty in the crustal thickness given in Chapter 5. In Chapter 5, a major assumption used to constrain both Moho uplift and the thickness of mare basalts was that the mare units are the sole source of super-isostatic mass in the basin region. Given that this assumption is correct, the error in Moho relief is probably about ± 10 km. We have computed the thermal histories for isotherms uplifted to ± 10 km from that presented in this chapter and found that temperatures vary by less than 10% from the model shown in Figure 8.

The displacements and thermal stresses accumulated at the surface in model A at times 10, 100, and 500 m.y. after basin formation are shown in Figure 9. The center of the basin subsides (vertical displacement w , Figure 9a) by about 180 m in the first 10 m.y., but subsides only an additional 175 m during the next 90 m.y. The horizontal extent of significant subsidence is controlled by the breadth of uplifted material beneath the basin. By 500 m.y., a region out to 200 km radius has experienced at least 100 m of subsidence. Subsidence at the center of the basin approaches 430 m by this time.

The amount of subsidence predicted at the center of the basin for model A (~ 430 m) is about an order of magnitude less than the observed relief of the central basin depression. Even if upper mantle isotherms were raised to the surface during basin formation, the total accumulated subsidence would not exceed 1 km. Thus, if a large portion of the relief associated with the central depression is a consequence of thermal subsidence, some additional source of initial heat is required.

The radial displacement u (Figure 9c) is toward the center of the basin. The magnitude of u is greatest between 150 and 200 km radial distance and accumulates to about -225 m by 500 m.y. This magnitude is about half that of the subsidence of the basin center over the same time interval.

The radial (σ_{rr} , Figure 9b) and azimuthal ($\sigma_{\theta\theta}$, Figure 9d) stresses in the central basin region are compressional and similar in magnitude. Because σ_{zz} and all shear stresses are zero at the surface, thrust faulting should be the dominant mode of stress release near the center of the basin. Both stresses accumulate most rapidly in the first 10 m.y. and reach 1.6 kbar by 100 m.y. The preferred direction of thrusting could vary depending on local structural heterogeneities or sources of stress other than local thermal contractions (e.g., tidal stress).

While $\sigma_{\theta\theta}$ approaches zero with increasing r , σ_{rr} passes through zero at about 200 km radial distance and becomes extensional at greater distance (Figure 9). The zone of

maximum extensional σ_{rr} is located between 250 and 350 km radial distance. By 100 m.y., maximum extensional stress in the radial direction reaches -0.4 kbar at $r = 290$ km. To understand better the style of faulting produced in regions of the basin where σ_{rr} is extensional and $\sigma_{\theta\theta}$ is compressional, we examine the Mohr failure criterion for diabase derived by Brace [1964]. The Mohr failure envelope shown in Figure 10 was calculated from the strength of diabase observed in axial extension and compression experiments in the laboratory. Coulomb-law envelopes, which are straight lines, predict a constant angle of failure under any stress state. However, a more parabolic shape to the failure envelope, also observed in the laboratory and field by others [e.g., Griggs and Handin, 1960; Muehlberger, 1961], is evident from the results of Brace [1964]. The fracture envelope intersects the normal stress (σ) axis at a right angle. This type of failure criterion was inferred from the observation that extension fractures formed normal to the direction of maximum extension when samples were subject to small confining pressures. Applied to the formation of fissures via basin cooling ($\sigma_{rr} = \sigma_3 =$ negative, $\sigma_{\theta\theta} = \sigma_1 =$ positive, $\sigma_{zz} = 0$), extensional fissures roughly concentric to the basin center should form when the Mohr circle with diameter $\sigma_{\theta\theta} - \sigma_{rr}$ has a radius of curvature less than that of the appropriate failure envelope in the extensional regime ($\sigma < 0$) and when σ_{rr} exceeds the extensional strength of material near the lunar surface. In diabase, for example, if σ_{rr} reaches about -0.4 kbar and $\sigma_{\theta\theta}$ is less than about 1.5 kbar,

extensional fractures develop (see Mohr circle a, Figure 10). However, if $\sigma_{\theta\theta}$ surpasses 1.5 kbar while σ_{rr} is negative but greater than -0.4 kbar, strike-slip failure is predicted on one or both of a pair of conjugate shear planes. For Mohr circle b (Figure 10) for instance, the predicted shear planes would trend $\pm 24^\circ$ from the direction of $\sigma_{\theta\theta}$.

The use of any single failure envelope to represent the near-surface crust in the vicinity of Orientale and to predict the existence and style of failure should be viewed with caution, for several reasons. First, the mechanism of failure most important at any given radius will depend not only on the values of σ_{rr} and $\sigma_{\theta\theta}$, but also on perturbations to the stress field caused by local structural, topographic and material heterogeneities and on the failure criterion specific to the region. Additionally, because the ring of fissuring around Orientale represents some type of brittle failure, stress was not accumulated monotonically with time but was locally relieved on perhaps an episodic basis. Once stress has been relieved through brittle failure, a new failure criterion must be defined for the region.

Laboratory studies of failure nonetheless provide information useful to our assessment of the importance of thermal stress to basin tectonics. The work of Brace [1964] and others suggests that at near-surface conditions, both conjugate strike-slip failure and extensional fracturing are possible in a stress regime where σ_{rr} is extensional and $\sigma_{\theta\theta}$ is compressional. Further, a rock will fail in either shear or

extension at or before the extensional strength of the rock is reached. For igneous rocks, measured extensional strengths range from about -0.2 to -0.4 kbar. Therefore, fissuring is most likely in regions of the basin where σ_{rr} is less than -0.2 to -0.4 kbar, though if $\sigma_{\theta\theta} - \sigma_{rr}$ is sufficiently large, strike-slip fissuring could occur wherever σ_{rr} is extensional.

In model A, the region of the basin where σ_{rr} is less than -0.2 kbar at 100 m.y. is from $r = 230$ to 400 km. Though the polarity of stresses produced by basin cooling predicts well the style of failure seen in Orientale, the zone of fissuring predicted by cooling of uplifted isotherms alone is at radial distances greater than that observed. If the fissures originated by thermal stress, the anomalous initial temperature field contributed by impact heating must have been localized within a radial distance smaller than that characterizing isotherm uplift.

Impact heating

Superimposed on the heat distribution beneath Orientale arising from isotherm uplift was an uncertain quantity of heat converted from impact kinetic energy. The loss of this impact heat gives rise to additional displacements and thermal stresses which must be added to those contributed by cooling following isotherm uplift.

As discussed previously, we assume that implanted impact heat decays exponentially with distance from the point of impact. The decay constant, s , defines the distribution of heat beneath the basin (equations 16 and 17). How much of the

implanted heat is ejected from of the excavated cavity is also highly uncertain. An inversion of gravity and topographic data for nearside lithosphere structure performed in Chapter 5 (Figure 8) indicates that the crust has been considerably thinned beneath Orientale, as it has been beneath other large lunar basins. The profile of crustal and upper mantle structure beneath Orientale (Figure 10a, Chapter 5) suggests that at least 55 km thickness of shock-heated crust was removed during the excavation of the central portion of the cavity. We assume that the upper 55 km of the hemispherical distribution of converted impact energy density was completely removed from the basin as ejecta.

In the models for Orientale presented below, we keep s as a free parameter with which the locations of maximum stress and thus of tectonic features can be modeled. The other free parameter in our models is the quantity of implanted heat (E_B) initially residing beneath the basin. The quantity E_B is weakly constrained by scaling relationships for impact kinetic energy (equation 15), studies of partitioning of impact kinetic energy [e.g., O'Keefe and Ahrens, 1976], and estimates of the excavated volumes of basins. We begin by assuming that $E_B = 10^{32}$ ergs and show later the effect of varying this highly uncertain quantity.

In Figures 11 and 12 are shown the initial temperature distributions and cooling histories produced by impact heating beneath Orientale assuming $E_B = 10^{32}$ erg and $s = 25$ km (model B) and 90 km (model C), respectively. Most of the initial heat

in model B is concentrated within a small volume near the basin surface. The initial temperature immediately beneath the central basin region exceeds 1000°C . By 100 m.y., only 27% of the initial energy remains beneath the basin. Virtually all the heat has left the basin by 500 m.y. The initial heat is distributed over a larger volume in model C relative to model B, and therefore the initial temperatures are considerably reduced. For the same reason, heat leaves the basin more slowly. About 60% of the initial energy remains buried in the subsurface after 100 m.y. After an additional 400 m.y. of cooling, only 8% remains. A substantial fraction of the deformation resulting from cooling in either model B or C will be completed by 100 m.y., a necessary condition if thermal displacement and stress are responsible for the fissuring within Orientale.

The thermoelastic effects of cooling in models B and C are illustrated in Figures 13 and 14, respectively. Shown in each figure are w and σ_{rr} versus r and t at the lunar surface. The shapes of plots of u and $\sigma_{\theta\theta}$ and their relationships to w and σ_{rr} are similar to those for model A (Figure 9).

Comparison of Figures 13 and 14 shows that the patterns of accumulated displacement and stress reflect the distribution of initial heat (defined by s) in each model. Most of the subsidence in model B is confined to the first 100 m.y. and to radial distances less than 150 km. Subsidence near $r = 0$ (Figure 13a) is almost 1.8 km by 100 m.y. Subsidence beneath the basin center for model C (Figure 14a) is lower in magnitude

(only ~ 200 m by 100 m.y.) but takes place over a broader region. Also, because heat is lost more slowly in model C, a significant proportion of the basin's total subsidence takes place between 100 and 500 m.y. Radial stress (Figure 13b) in the center of the basin is compressional and exceeds 20 kbar by the time the cooling is complete in model B. Maximum extensional stress in this model occurs between $r = 90$ and 130 km and accumulates to about -3 kbar. In model C (Figure 14b), σ_{rr} at $r = 0$ reaches 1.2 kbar by 100 m.y. The radius at which σ_{rr} becomes extensional increases with increasing basin age. The distribution of σ_{rr} in model C contains a broader region of extension compared to that of model B. For $E_B = 10^{32}$ erg, maximum extension accumulates to -0.2 kbar by 100 m.y.

Neither model B or C satisfactorily predicts the location of fissures and the magnitude and relief of the central depression within Orientale. However, the above models indicate that the form of the associated displacement and stress fields are sensitive to the value of the decay constant s . The distributions of displacement and thermal stress scale linearly with E_B , but the shapes of those distributions are independent of E_B . Therefore, if the central depression and fissures are products of thermal stress and if the elastic half-space model presented adequately represents the Moon during the time of basin formation and modification, we may constrain the quantity and distribution of heat implanted by varying s so as to match the horizontal extent of the central depression and the location of fissuring and by varying E_B so

as to match a given fraction of the relief of the central depression.

A model following these guidelines, model D, is shown in Figure 15a. For this model $E_B = 5 \times 10^{32}$ erg and $s = 50$ km. Significant impact heating extends to radial distances and depths somewhat greater than the radius of Orientale (~ 310 km). Near-surface temperatures within 100 km of basin center exceed the liquidus temperatures of most igneous rocks. Cooling proceeds at a rate intermediate to the rates in models B and C. By 100 m.y., 24% of the initial energy has been lost; less than 10% of E_B remains in the target region by 500 m.y.

Figure 16 shows the subsidence and surface radial stress for model D. The center of the basin subsides by 3.3 km after 500 m.y. This subsidence is in addition to the 0.4 km of subsidence contributed by the loss of heat from isotherm uplift (model A). Though it is difficult to assess what fraction of the relief of Orientale's central depression is due to thermal subsidence, the sum of the subsidence from models A and D accounts for most of the present relief from the basin center to the foot of the inner Rook Mountains. Also, that most of the subsidence occurs at radial distances less than 200 km is in agreement with the observed topographic profile (Figure 2).

By 100 m.y., accumulation of σ_{rr} in Model D is almost complete (Figure 15). This is in agreement with the inferred period of fissuring in the corrugated and plains facies of Orientale. Near the center of the basin, σ_{rr} is compressional and exceeds the measured strength of igneous samples under

uniaxial compression (2-5 kbar, Brace, [1964]). Radial stress is most extensional between about 100 and 250 km radial distance. Between 10 and 100 m.y., the location of maximum extensional σ_{rr} moves from $r = 140$ to $r = 180$ km. The greatest extensional stress reaches -2 kbar by 10 m.y. and -3 kbar by 100 m.y. Even by 10 m.y., σ_{rr} thus exceeds the measured strength of laboratory rocks under extension at near-surface conditions [Brace, 1964]. Thus, fissuring is likely before 10 m.y. of cooling has taken place. The location of maximum extensional σ_{rr} is consistent with the band of fissures observed at radial distances between 150 and 230 km from the center of Orientale.

Effect of an Elastic Blocking Temperature

As discussed previously, it is probable that portions of the crust and mantle at sufficiently elevated temperatures will not contribute significantly to the thermal stress field. Thermal stress models (E and F) that incorporate an elastic blocking temperature [Turcotte, 1974, 1983] are shown in Figures 17 and 18. In both models, the anomalous temperature field contains contributions from both isotherm uplift (model A) and impact heating (model B). To determine whether a parcel of material is above or below the elastic blocking temperature T_e , the selenotherm (Figure 7a) must be added to the anomalous temperature. In model E a blocking temperature of $T_e = 550^\circ$ is imposed. This is the isotherm suggested by McNutt and Menard [1982] to represent the base of the mechanically strong lithosphere in terrestrial ocean basins.

In model F, we use a blocking temperature of 800°C, corresponding to the temperature at the greatest depth of earthquakes in terrestrial intraplate settings [Chen and Molnar, 1983].

At $t = 0$ in model E, all temperatures beneath the center of the basin exceed $T_e = 550^\circ \text{C}$, even ignoring impact heating. This is evident from an examination of Figure 7. Impact heating of the sub-basin region further raises the initial temperatures beneath the basin. Therefore, thermal stress during early cooling is confined to the shallow crust exterior to the regions most extensively heated by the basin formation process. This effect is reflected in the shape of the σ_{rr} distribution at $t = 10$ m.y. (Figure 17). The zone of maximum contraction contributing to thermal stress occurs not beneath the basin center during this time interval, but near $r \sim 125\text{km}$, where temperatures are below 550°C . The cooled, elastic surface layer is pulled toward this annular region, thus producing zones of mild extension near $r = 0$ and $r = 300$ km. By 100 m.y., most of the crust beneath the basin has cooled below T_e and, as a result, has begun to contribute to and accumulate thermal stress. By this time, the distribution of σ_{rr} begins to resemble models with no blocking temperature. However, because the total volume of material cooler than T_e is small relative to the total volume heated during basin formation, the magnitude of σ_{rr} is everywhere less than in previous models. This is especially evident near the basin center where σ_{rr} in model B (Figure 13) exceeded 20 kbar at

500 m.y. Also, the zone at the surface experiencing radial extension is strongly controlled by both the value of T_e and the radial extent of isotherm uplift beneath the basin. The result is a region of extensional stress that is broader, and located at a greater radial distance from the basin center, than that which would be produced by the same model without a blocking temperature. In fact, models computed using the same value of T_e and the same amount of isotherm uplift as was used in model E, but including impact heat confined to regions smaller than the region of isotherm uplift, differ little from that shown in Figure 17. After 100 m.y. in model E, σ_{rr} exceeds the extensional strength of igneous rocks (~ -0.2 to -0.4 kbar) only for $r > 250$ km. Model E, then, poorly accounts for the observed fissuring between $r = 150$ and 230 km within the Orientale basin.

By increasing the blocking temperature to 800°C in model F (Figure 18), a larger volume of basin subsurface is permitted to contribute to thermal stress. However, the results of model F are significantly different than if no blocking temperature were assumed. This may be seen by comparing σ_{rr} from model F (Figure 18a) with the sum of σ_{rr} from models A (Figure 9) and B (Figure 13). Stress magnitudes are smaller in model F, and the region over which σ_{rr} exceeds the extensional strength of igneous rock is broader and farther from the basin center. Still, by $t = 100$ m.y., σ_{rr} is more extensional than -0.2 kbar between $r = 180$ and 450 km and $\sigma_{\theta\theta}$ is positive (Figure 18b), thus providing a state of stress sufficient to produce the

fissuring observed around Orientale. If the impact heating contribution had a value for s much larger than that used here ($s = 25$ km), the region of extensional stress would be pushed to even greater radial distances. Therefore, even though model D ($s = 50$ km) provided the best fit to the observed topography and tectonics of Orientale among models not including the effects of a blocking temperature, models with a blocking temperature as high as 800°C provide a better fit to topography and tectonics if the impact heating contribution was concentrated within a relatively smaller volume.

DISCUSSION

The exploratory models presented above suggest that the cooling of an impact basin following heating by conversion of impact kinetic energy and uplift of isotherms was an important contributor to the topography and fissuring observed within the Orientale basin. Beyond this qualitative result, however, what can we say about the quantity and distribution of initial heat beneath large lunar basins and, by inference, about the basin-forming impact process? If the Moon behaved as an elastic half-space throughout the range of temperatures encountered during the thermal evolution of a basin, model D ($E_B = 5 \times 10^{32}$ erg, $s = 50$ km, Figures 15-16) predicts best the 4 km of relief defining the central basin depression and the radial position of the ring of fissures ($r = 150$ and 230 km). We know from observations on the Earth, however, that rocks undergo ductile deformation at elevated temperatures for a given stress difference and strain rate and that a substantial

volume of the subsurface beneath newly-formed impact basins were at least initially in such a temperature range. The incorporation of an elastic blocking temperature (T_e) can account approximately for such effects, but only in an incompletely satisfying and uncertain way.

Despite uncertainties in the appropriate value of T_e and, more fundamentally, in the ability of models which include a blocking temperature to represent adequately anelastic processes in the Moon, we may derive at least rough constraints on the quantity and distribution of impact heat implanted during the formation of the Orientale basin using the calculations of this chapter. To do so, we must assume that fissuring is a product of basin cooling and thermal stress [Church et al., 1982] and that the model for the field of anomalous temperatures resulting from isotherm uplift is relatively well known [Chapter 5]. With these assumptions we may note, first, isotherm uplift alone predicts poorly the location of fissuring within Orientale. Therefore, heat converted from impact kinetic energy must be of greater significance to the early thermal budget of the basins. Put differently, E_B is probably greater than the total amount of anomalous heat contributed by isotherm uplift (1.4×10^{32} erg). When no blocking temperature has been applied, the decay constant s must be about 50 km (model D, Figures 15-16) to predict correctly the occurrence of fissuring between $r = 150$ and 230 km. Incorporation of a blocking temperature necessitates the need for an even greater concentration of

impact heat near the point of impact. We therefore suggest that the decay of impact heat density with distance from the point of projectile impact for an Orientale-size event must be rapid and that for the exponential decay assumed in this paper (equation 16), $s = 50$ km may be an upper bound on the decay constant. For comparison, the energy density distribution shown in Figure 5 of O'Keefe and Ahrens [1975] for their numerical model of Imbrium (assumed $D_c = 650$ km) falls off roughly exponentially with a decay constant s of about 20 km.

While the total uplifted heat (1.4×10^{32} erg) provides a lower bound on the quantity E_B of impact kinetic energy implanted as heat beneath the basin, the work of this chapter also yields an upper bound. First, an upper bound of 5 km can be placed on the total subsidence that has occurred within the central basin region of Orientale based on the observed thickness of mare units [~ 1 km, Head, 1974] and the relief of the depression measured from the foot of the inner Rook Mountains to the top of the mare units [Head et al., 1981]. It is likely that only a portion of this total is due to thermal subsidence. The sum of models A and D ($E_B = 5 \times 10^{32}$ erg and $s = 50$ km) accounts for 3.7 km of the observed depression. Given s , the magnitude of subsidence is proportional to E_B . Thus if $s = 50$ km, then $E_B = 6.3 \times 10^{32}$ erg can explain all 5 km of the upper bound on the subsidence. Therefore 6.3×10^{32} erg is an upper bound on the actual value of E_B . Also, if $s < 50$ km, subsidence at the center of the basin increases for a given value E_B (compare models B and C, Figures

13 and 14). Because the discussion in the paragraph above suggests that s must be less than 50 km, 5×10^{32} erg is again an upper bound on E_B .

If we assume that E_B represents about 25% [O'Keefe and Ahrens, 1976] original kinetic energy (E_K) of the impacting projectile, and we use the lower (1.4×10^{32} erg) and upper (6.3×10^{32} erg) bounds suggested by our analysis, we can constrain (though weakly) E_K for the Orientale event to be greater than about 6×10^{32} erg but less than 2.5×10^{33} erg. These bounds may act as useful constraints for models of planetary accretion by the impact of sizeable planetesimals and subsequent planetary thermal histories [e.g., Kaula, 1979] in which the fractional conversion impact kinetic energy to heat and the spatial distribution of that heat are important parameters.

CONCLUSIONS

We have explored the hypothesis that thermal stress has made a significant contribution to the topography and tectonics of multi-ringed basins. The models presented are based on the response of an elastic half-space to cooling following the emplacement of heat by dissipation of impact kinetic energy of and uplift of isotherms during basin formation. Thermal stresses and displacements have been calculated both with and without the inclusion of an elastic blocking temperature. These solutions have been compared with the magnitude of topographic relief [Head et al., 1981], the location of fissuring and the timing of fissure formation [Church et al., 1982] in the relatively well-preserved Orientale basin.

For all reasonable thermal models, basin cooling and accumulation of thermal stress is most rapid in the first 100 m.y. after basin formation, in agreement with the inferred timing of fissuring within Orientale. The predicted state of stress in the fissured region (σ_{rr} = extensional, $\sigma_{\theta\theta}$ = compressional) predicts well the form and trends of fissures whether these features are extension fractures, strike-slip faults, or both.

If the Moon's response to cooling was completely elastic, then the relief of the central depression to the base of the mare units (~ 5 km) and the occurrence of fissuring between radial distances of 150 and 230 km from the center of Orientale can be explained by postulating the quantity E_B of implanted impact energy was 6.3×10^{32} erg and its decay with a slant-range distance q from the point of impact varied as $e^{-q/s}$, where $s = 50$ km. Because of the anelastic effects of material at high temperatures on the thermal stress field, however, and the probability that only a portion of the central depression is related to thermal subsidence, it is likely that $E_B = 6.3 \times 10^{32}$ erg and $s = 50$ km are upper bounds on the actual values of these quantities. That the effect of isotherm uplift alone fails to explain the location of fissuring suggests that the contribution to the anomalous temperature field from impact heating is probably of greater magnitude than the 1.4×10^{32} erg of heat contributed by isotherm uplift. The amount of kinetic energy emplaced as heat beneath Orientale was therefore between about 1.4×10^{32} and 6.3×10^{32} erg.

There is, of course, an untested element of uncertainty in the ability of our models to represent the anelastic response of material at high temperature to cooling. The models presented here nonetheless suggest that basin cooling and thermal stress contributed significantly to the topography and tectonics of multi-ringed basins and that constraints on the initial quantity and distribution of heat may be derived from topographic and tectonic observations and the sensitivity of our models to the initial temperature field.

REFERENCES

- Baldrige, W.S. and G. Simmons, Thermal expansion of lunar rocks, Proc. Lunar Sci. Conf. 2nd, 2317-2321, 1971.
- Brace, W.F., Brittle fracture of rocks, in State of Stress in the Earth's Crust, edited by W. R. Judd, pp. 110-178, Elsevier, New York, 1964.
- Bratt, S.R., S.C. Solomon, and J.W. Head, The evolution of multi-ringed basins: Cooling, subsidence, and thermal stress (abstract), Lunar Planet. Sci., 12, 109-111, 1981.
- Carslaw, H.S., and J.C. Jaeger, Conduction of Heat in Solids, 2nd ed., 510 pp., Oxford University Press, 1959.
- Chen, W., and P. Molnar, Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res., 88, 4183-4214, 1983.
- Church, S., J.W. Head, and S.C. Solomon, Multi-ringed basin interiors: Structure and early evolution of Orientale (abstract), Lunar Planet. Sci., 13, 98-99, 1982.
- Comer, R.P., S.C. Solomon, and J.W. Head, Elastic lithosphere thickness on the moon from mare tectonic features: A formal inversion, Proc. Lunar Planet. Sci. Conf. 10th, 2441-2463, 1979.
- Dence, M.R., Impact melts, J. Geophys. Res., 76, 5552-5565, 1971.
- Gault, D.E., and E.D. Heitowit, The partitioning of energy for hypervelocity impact craters formed in rock, Proc. Sixth Hypervelocity Impact Symposium, 2, 419-456, 1963.

- Gault, D.E., and J.A. Wedekind, Experimental hypervelocity impact into quartz sand, II, Effects of gravitational acceleration, in Impact and Explosion Cratering, ed. D.H. Roddy, R.O. Pepin, and R.B. Merrill, pp. 1231-1260, Pergamon Press, NY, 1977.
- Gautschi, W., Error function and Fresnel integrals, in Handbook of Mathematical Functions with Formulas, Graphs, and Mathematical Tables, ed. M. Abramowitz and I.A. Stegun, pp. 295-329, National Bureau of Standards, Washington, D.C. 1964.
- Goodier, J.N., On the integration of the thermoelastic equations, Phil. Mag., 7, 1017-1032, 1937.
- Grieve, R.A.F., and M.J. Cintala, A method for estimating the initial impact conditions of terrestrial cratering events, exemplified by its application to Brent crater, Ontario, Proc. Lunar Planet. Sci., 12B, 1609-1621, 1981.
- Grieve, R.A.F., M.R. Dence, and P.B. Robertson, Cratering processes: As interpreted from the occurrence of impact melts, in Impact and Explosion Cratering, ed. D.J. Roddy, R.O. Pepin, R.B. Merrill, pp. 791-814, Pergamon Press, NY, 1977.
- Griggs, D., and J.W. Handin, Observations on fracture and a hypothesis of earthquakes, Geol. Soc. Am. Mem., 79, 347-364, 1960.
- Hartman, W.K., and C.A. Wood, Moon: Origin and evolution of multi-ring basins, Moon, 3, 3-78, 1971.
- Head, J.W., Orientale multi-ringed basin interior and implications for the petrogenesis of lunar highland

- samples, Moon, 11, 327-356, 1974.
- Head, J.W., Origin of outer rings in lunar multi-ringed basins: Evidence from morphology and ring spacing, in Impact and Explosion Cratering, ed. D.J. Roddy, R.O. Pepin, R.B. Merrill, pp. 563-573, Pergamon Press, NY, 1977.
- Head, J.W., E. Robinson, and R. Phillips, Topography of the Orientale basin, Lunar Planet. Sci., 12, 421-423, 1981.
- Holsapple, K.A., and R.M. Schmidt, On the scaling of crater dimensions, 2. Impact processes, J. Geophys. Res., 87, 1849-1870, 1982.
- Ivanov, B.A., The effect of gravity on crater formation: Thickness of ejecta and concentric basins, Proc. Lunar Sci. Conf. 7th, 2947-2965, 1976.
- Kaula, W.M., Thermal evolution of earth and moon growing by planetesimal impacts, J. Geophys. Res., 84, 999-1008, 1979.
- Masters, J.I., Some applications in physics of the P function, J. Chem. Phys., 23, 1865-1874, 1955.
- Maxwell, D.E., Simple Z model of cratering, ejection and overturned flap, in Impact and Explosion Cratering, ed. D.J. Roddy, R.O. Pepin, R.B. Merrill, pp. 1003-1008, Pergamon Press, NY, 1977.
- McNutt, M., and H.W. Menard, Constraints on yield strength in the oceanic lithosphere derived from observations of flexure, Geophys. J. Roy. Astron. Soc., 71, 363-394, 1982.
- Melosh, H.J., and W.B. McKinnon, The mechanics of ringed basin formation, Geophys. Res. Lett., 5, 985-988, 1978.

- Mindlin, R.D., and D.H. Cheng, Thermoelastic stress in the semi-infinite solid, J. Appl. Phys., 21, 931-933, 1950.
- Mizutani, H., and M. Osako, Elastic-wave velocities and thermal diffusivities of Apollo 17 rocks and their geophysical implications, Proc. Lunar Planet. Sci. 5th, 2891-2901, 1974.
- Moon, W., A new method of computing geopotential fields, Geophys. J. Roy. Astron. Soc., 67, 735-746, 1981.
- Moore, H.J., C.A. Hodges, and D.H. Scott, Multiringed basin - illustrated by Orientale and associated features, Proc. Lunar Sci. Conf. 5th, 71-100, 1974.
- Moulton, F.R., Celestial Mechanics, 2nd ed., 437 pp., Macmillan, New York, 1914.
- Muehlberger, W.R., Conjugate joint sets at small dihedral angle, J. Geol., 69, 211-219, 1961.
- Nordyke, M.D., Nuclear craters and preliminary theory of mechanics of explosive crater formation, J. Geophys. Res., 66, 3439-3459, 1961.
- Nordyke, M.D., Nuclear cratering experiments: United States and Soviet Union, in Impact and Explosion Cratering, ed. D.J. Roddy, R.O. Pepin, R.B. Merrill, pp. 103-124, Pergamon Press, NY, 1977.
- O'Keefe, J.D., and T.J. Ahrens, Shock effects from a large impact on the moon, Proc. Lunar Sci. Conf. 6th, 2831-2844, 1975.
- O'Keefe, J.D., and T.J. Ahrens, Impact ejecta on the moon, Proc. Lunar Sci. Conf. 7th, 3007-3025, 1976.

- O'Keefe, J.D., and T.J. Ahrens, Impact-induced energy partitioning, melting, and vaporization on terrestrial planets, Proc. Lunar Sci. Conf. 8th, 3357-3374, 1977.
- Olver, F.W.J., Bessel functions of integer order, in Handbook of Mathematical Functions with Formulas, Graphs, and Mathematical Tables, ed. M. Abramowitz and I.A. Stegun, pp. 355-434, National Bureau of Standards, Washington, D.C., 1964.
- Onorato, P.I.K., D.R. Uhlmann, C.H. Simonds, The thermal history of the Manicouagan impact melt sheet, Quebec, J. Geophys. Res., 83, 2789-2798, 1978.
- Royden, L., F. Horváth, and J. Rumpler, Evolution of the Pannonian Basin system, 1. Tectonics, Tectonics, 2, 63-90, 1983.
- Schatz, J.F., and G. Simmons, Thermal conductivity of earth materials at high temperatures, J. Geophys. Res., 77, 6966-6983, 1972.
- Schmidt, R.M., A centrifuge cratering experiment: Development of a gravity-scaled yield parameter, in Impact and Explosion Cratering, ed. D.J. Roddy, R.O. Pepin, R.B. Merrill, pp. 1261-1278, Pergamon Press, NY, 1977.
- Shoemaker, E.M., Penetration mechanics at high velocity meteorites, illustrated by Meteorite crater, Arizona, Proc. Internat. Geol. Cong. 21st, Copenhagen, 418-434, 1960.
- Simmons, G., and W.F. Brace, Comparison of static and dynamic measurements of compressibility of rocks, J. Geophys. Res., 70, 5649-5656, 1965.

- Skinner, B.J., Thermal expansion, in Handbook of Physical Constants, ed. S.P. Clark, Jr., pp. 75-96, Geological Society of America, New York, 1966.
- Solomon, S.C., Mare volcanism and lunar crustal structure, Proc. Lunar Sci. Conf. 6th, 1021-1042, 1975.
- Solomon, S.C., R.P. Comer, and J.W. Head, The evolution of impact basins: Viscous relaxation of topographic relief, J. Geophys. Res., 87, 3975-3992, 1982.
- Solomon, S.C., and J.W. Head, Vertical movement in mare basins: Relation to mare emplacement, basin tectonics, and lunar thermal history, J. Geophys. Res., 84, 1667-1682, 1979.
- Solomon, S.C., and J.W. Head, Lunar mascon basins: Lava filling, tectonics, and evolution of the lithosphere, Rev. Geophys. Space Phys., 18, 107-141, 1980.
- Solomon, S.C., and J. Longhi, Magma oceanography: 1. Thermal evolution, Proc. Lunar Sci. Conf. 8th, 583-599, 1977.
- Timoshenko, S.P., and J.N. Goodier, Theory of Elasticity, 3rd ed., 567 pp., McGraw-Hill Book Co., New York, 1970.
- Turcotte, D.L., Are transform faults thermal contraction cracks?, J. Geophys. Res., 79, 2573-2577, 1974.
- Turcotte, D.L., Thermal stress in planetary elastic lithospheres, Proc. 13th Lunar Sci. Conf., Part 2, J. Geophys. Res., 88, Suppl., A585-A587, 1983.
- Vaile, R.B., Jr., Pacific craters and scaling laws, J. Geophys. Res., 66, 3413-3438, 1961.
- Wilhelms, D.E., Comparison of Martian and lunar multi-ringed circular basins, J. Geophys. Res., 78, 4084-4095, 1973.

Wilhelms, D.E., Relative ages of lunar basins (abstract),
Reports of Planetary Geology Program, 1978-1979, NASA
Technical Memorandum 80339, 135-137, 1979.

Wood, C.A., and J.W. Head, Comparison of impact basins on
Mercury, Mars and the Moon, Proc. Lunar Sci. Conf. 7th,
3629-3651, 1976.

FIGURE CAPTIONS

Figure 1. The Orientale basin as seen from Lunar Orbiter IV.

The basin's center is located at 20°S , 95°W . The diameter of the outer ring (Cordillera Mountains) is about 900 km.

Figure 2. Generalized topographic profile of Orientale, after Head et al. [1981]. Also shown are the locations of rings, facies, and structural features, after Church et al. [1982].

Figure 3. Map view of Orientale showing rings, fissures (thin lines) and graben (double, hatched lines).

Figure 4. Lunar Orbiter IV photograph of fissures with V-shaped cross sections located in the corrugated and plains facies southwest of the basin center. Photo number LOIV 195H1.

Figure 5. Geometry of the thermoelastic stress problem for a cooling impact basin. Cylinders of constant temperature change can be superposed to form any axisymmetric distribution of temperature change.

Figure 6. Scenario for the early stages of basin formation and heating.

Figure 7. (a) Pre-impact selenotherm 3.9 b.y. ago [Solomon and Head, 1979] and thermal profile after the excavation of Orientale and uplift of isotherms by 55 km beneath the center of the basin.

(b) Anomalous temperature beneath the center of the basin contributed by isotherm uplift.

Figure 8. Basin thermal evolution for model A. The initial field of anomalous temperature $t=0$) is that produced only by isotherm uplift. Also shown are the temperature fields 10, 100, and 500 m.y. after basin formation.

Figure 9. Displacements and thermal stresses for model A (Figure 8). Values shown are accumulated over times of 10, 100, and 500 m.y. after basin formation. (a) Subsidence w ; (b) radial stress σ_{rr} , (c) radial displacement u ; and (d) azimuthal or hoop stress $\sigma_{\theta\theta}$.

Figure 10. Mohr failure envelope for Frederick diabase derived from the laboratory experiments of Brace [1964].

Figure 11. Basin thermal evolution for model B. The initial anomalous temperature field is due to implanted impact kinetic energy; the total heat E_b buried beneath the basin is 10^{32} erg and the decay constant s is 25 km (see equations 16 and 17). Also shown are the temperature fields 10, 100, and 500 m.y. after basin formation.

Figure 12. Basin thermal evolution for model C. Parameters identical to the model of Figure 11 except that $s = 90$ km.

Figure 13. Subsidence w and surface radial thermal stress σ_{rr} for basin thermal model B (Figure 11). Values shown are accumulated after times of 10, 100, and 500 m.y. after basin formation.

Figure 14. Subsidence w and surface radial stress σ_{rr} for basin thermal model C (Figure 12).

Figure 15. Basin thermal evolution for model D. The model includes impact heating with $E_B = 5 \times 10^{32}$ erg and $s = 50$ km. See Figure 11 for further explanation.

Figure 16. Subsidence w and surface radial stress σ_{rr} for model D (Figure 15).

Figure 17. Surface radial stress σ_{rr} for model E, a combination of uplift heating from model A (Figure 8) and impact heating from model B ($E_B = 10^{32}$ erg and $s = 25$ km, Figure 11). An elastic blocking temperature T_e of 550°C has been imposed.

Figure 18. Surface thermal stresses for model F, a combination of uplift heating from model A and impact heating from model B. An elastic blocking temperature $T_e = 800^\circ\text{C}$ has been assumed.

Table 1. Adopted values of parameters used in thermal stress models

Variable	Description	Value	Source
ρ_c	crustal density	2.9 g/cm ³	<u>Solomon</u> [1975]
ρ_m	upper mantle density	3.4 g/cm ³	<u>Solomon</u> [1975]
C_p	specific heat	1.2×10^7 erg/g-°C	<u>Solomon and Longhi</u> [1977]
k	heat capacity	0.01 cm ² /s	<u>Schatz and Simmons</u> [1972]
α	volumetric thermal expansion coefficient	2.7×10^{-5} °C ⁻¹	<u>Baldrige and Simmons</u> [1971], <u>Skinner</u> [1966]
E	Young's modulus	7×10^{11} dyne/cm ²	<u>Mizutani and Osako</u> [1974], <u>Simmons and Brace</u> [1965]

Table 2. Models of thermal stress arising from the cooling of the Orientale basin

Model Number	Buried Heat (E_B), 10^{32} erg	Decay Constant (s), km	Blocking Temperature T_e , °C	Uplift heating included?	Figure Number(s)
A	0	-	-	yes	8,9
B	1	25	-	no	11,13
C	1	90	-	no	12,14
D	5	50	-	no	15,16
E	1	25	550	yes	17
F	1	25	800	yes	18

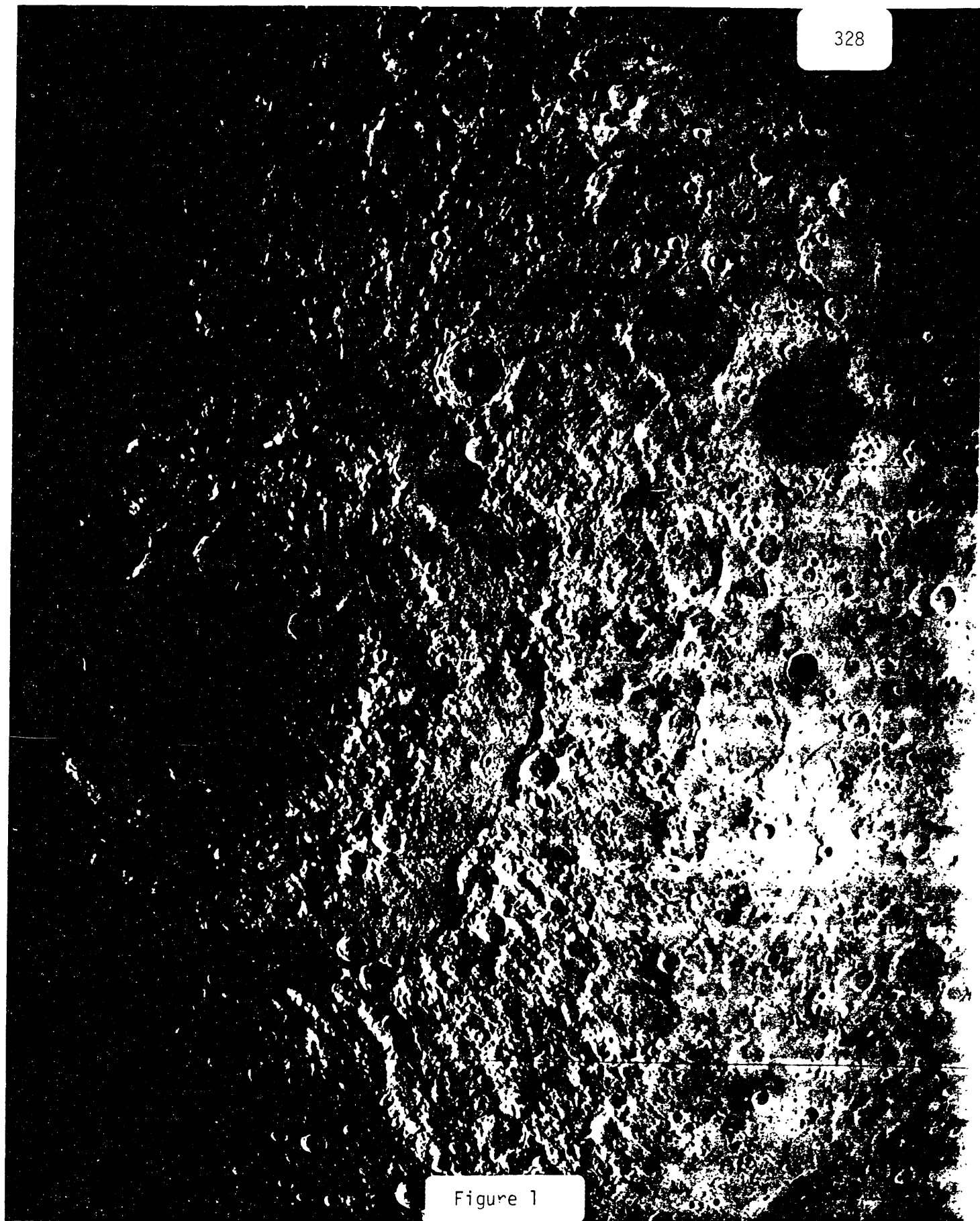


Figure 1

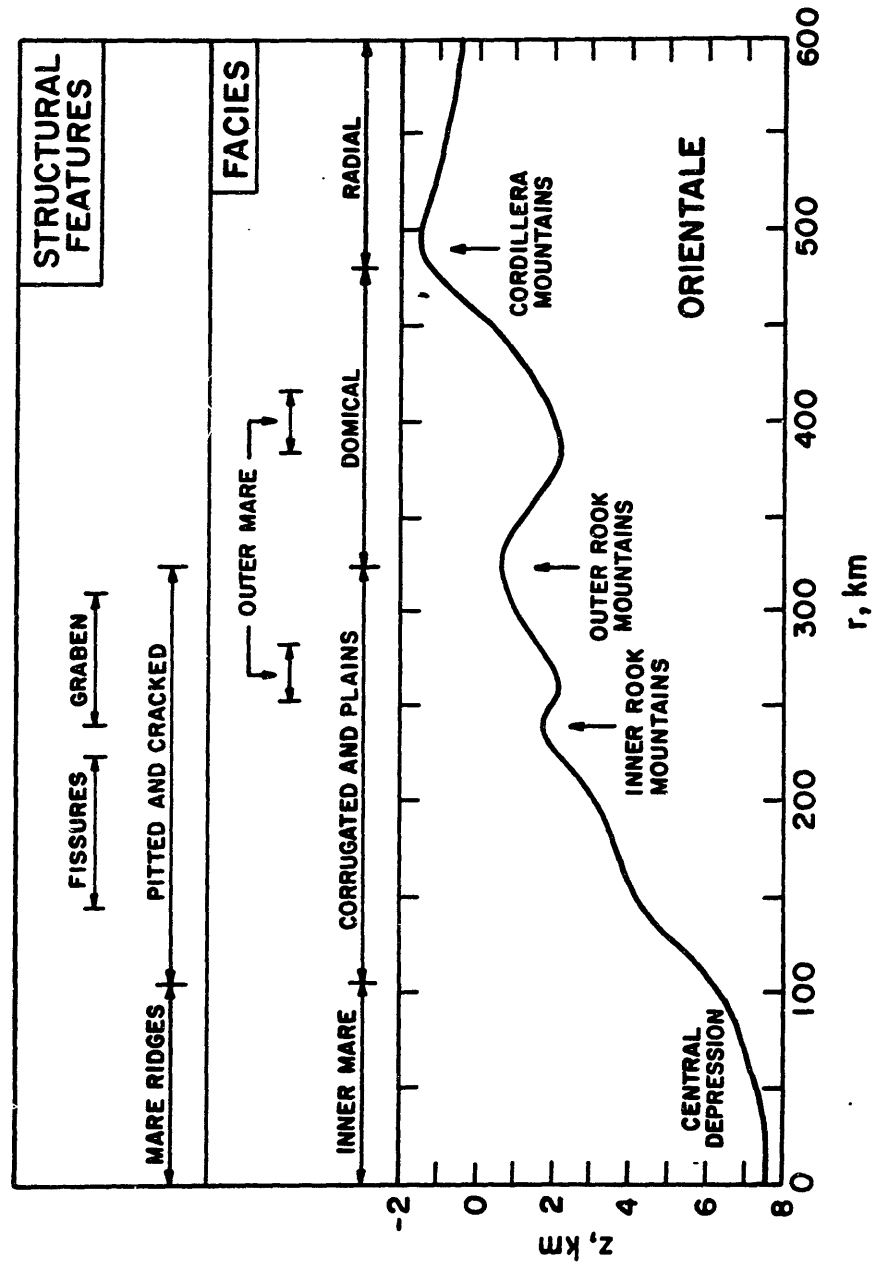


Figure 2

ORIENTALE STRUCTURE

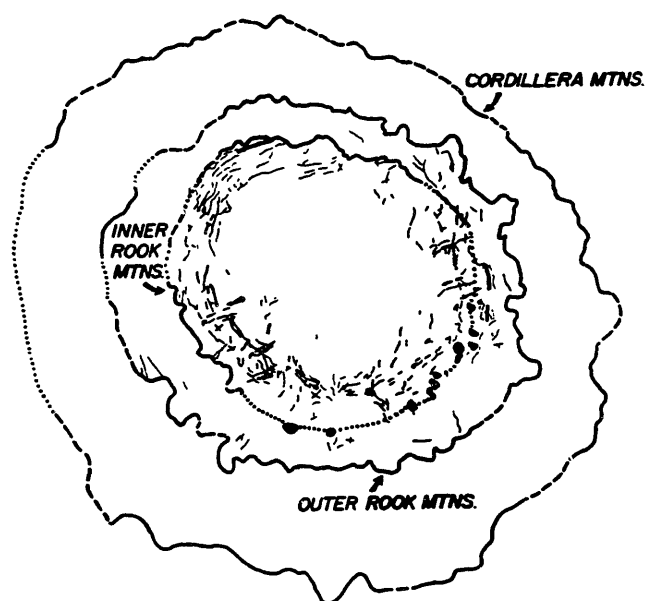


Figure 3

NASA-S-71-2266-V

331

Figure 4

50 KM



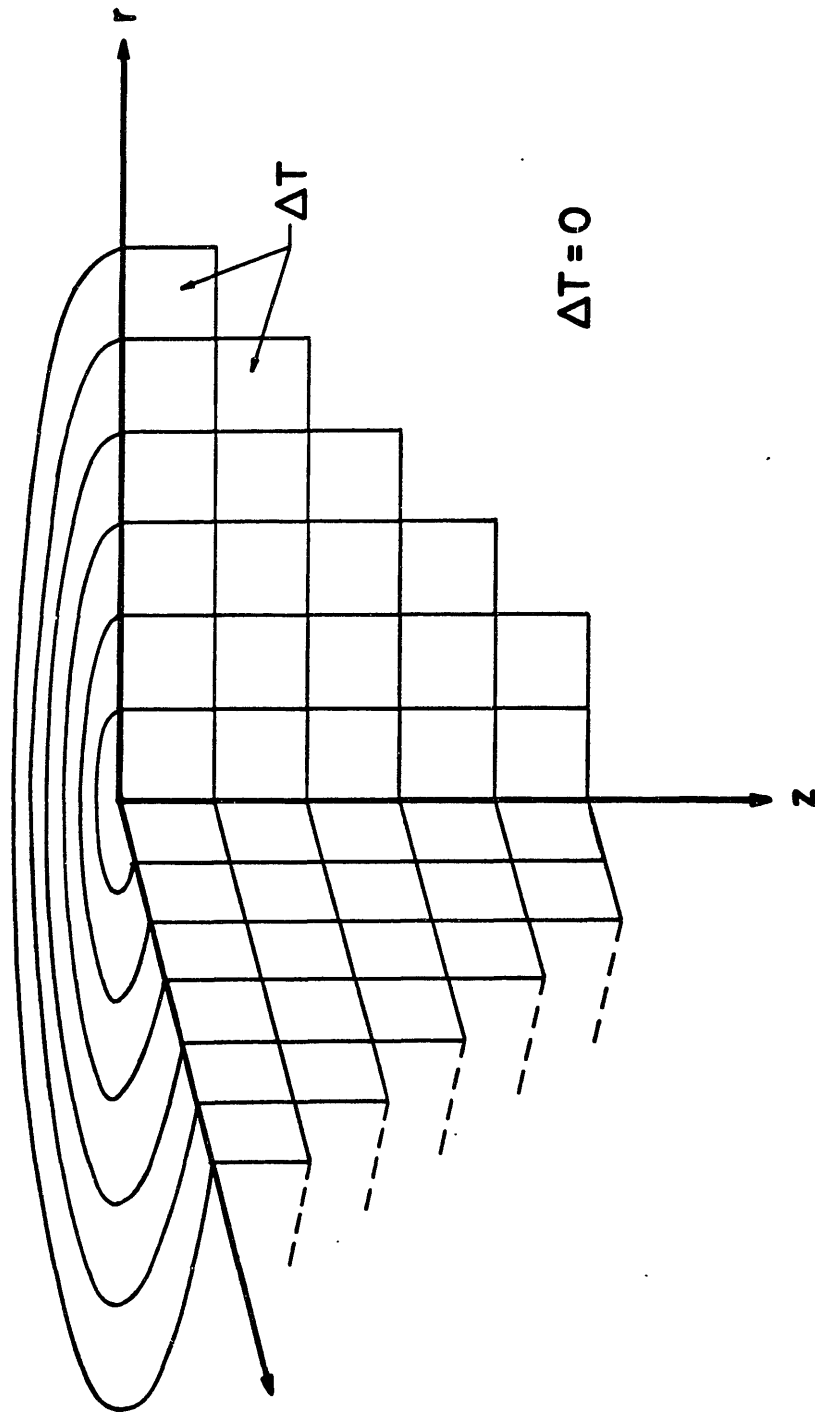


Figure 5

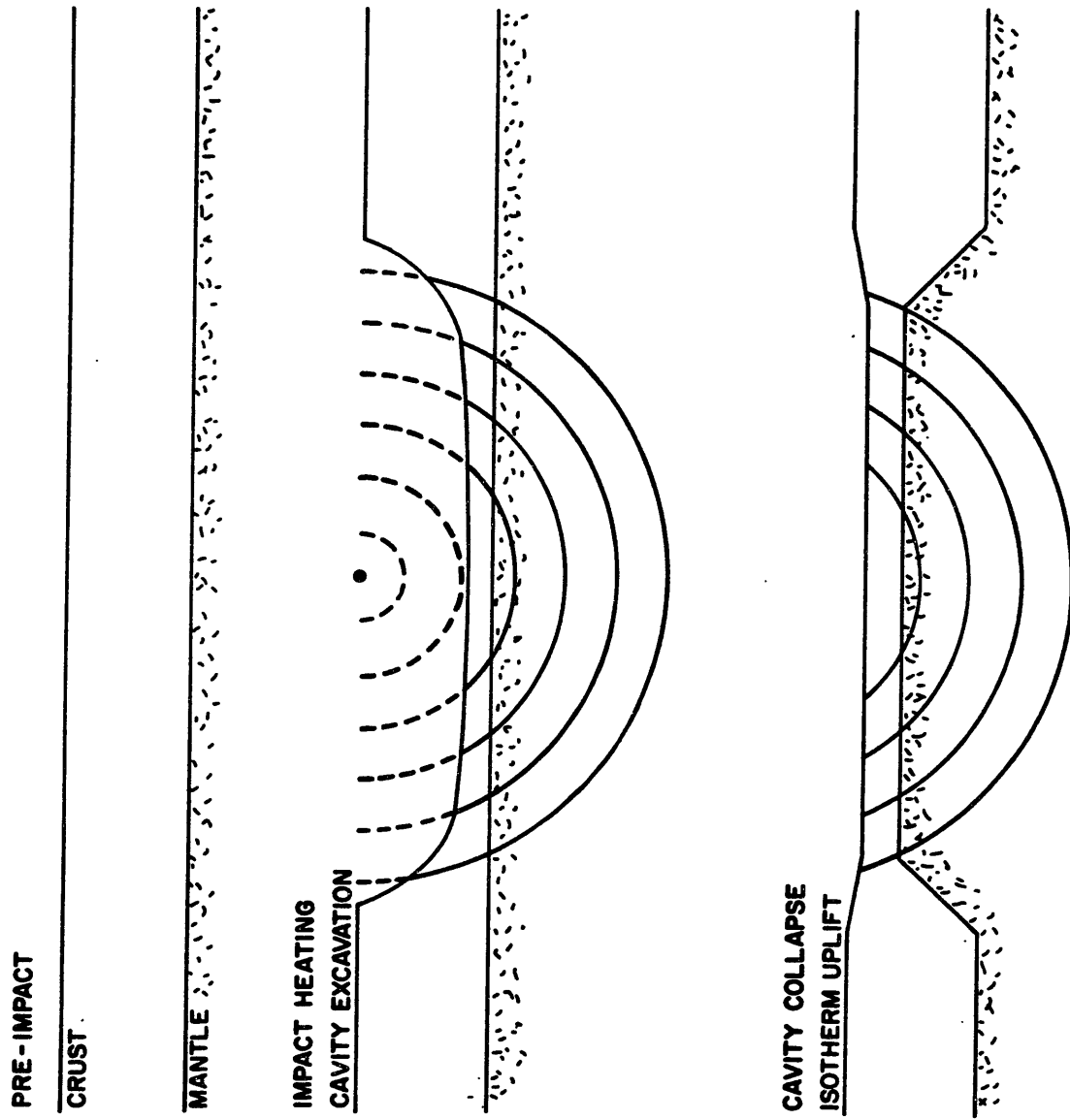


Figure 6

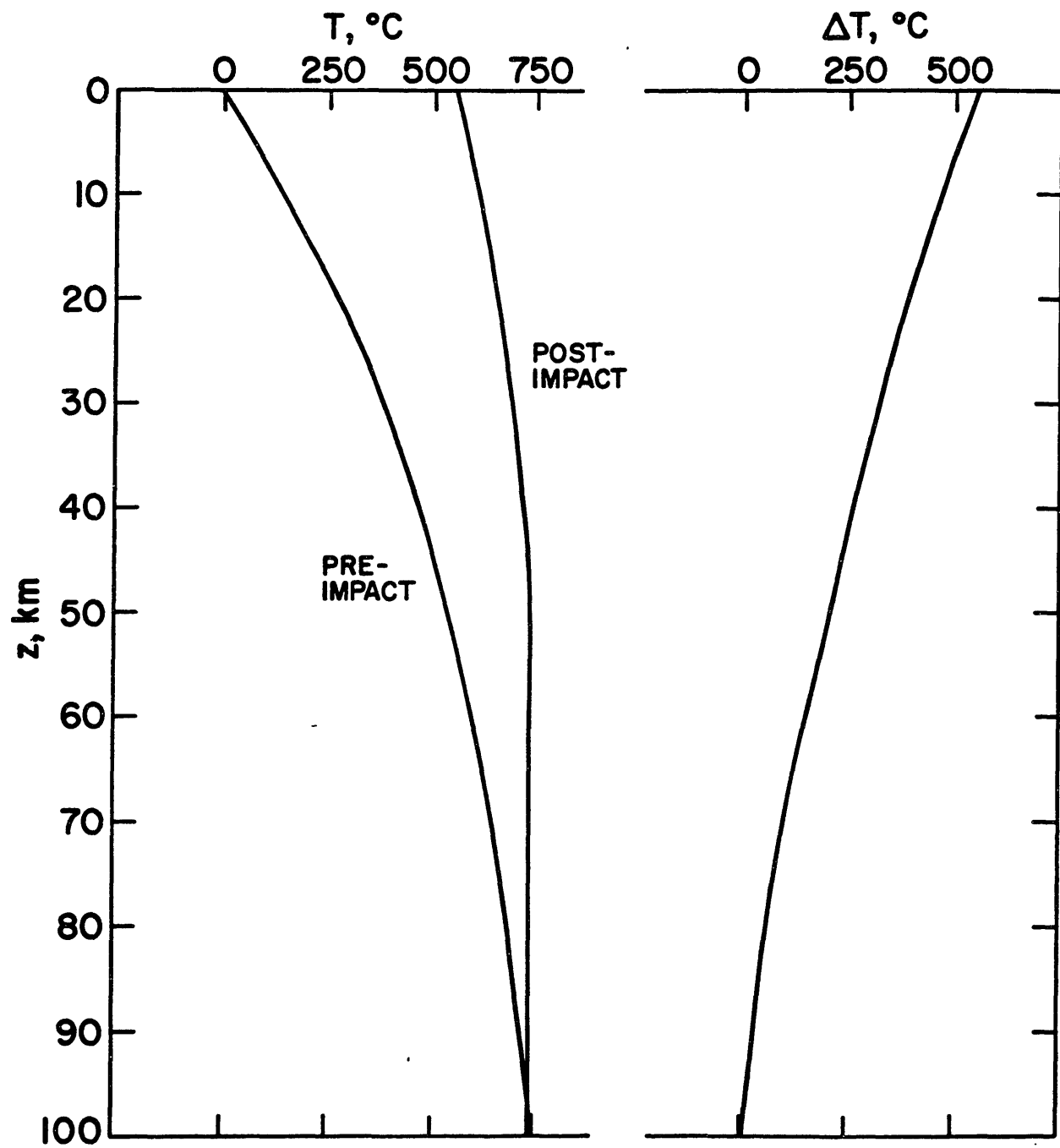


Figure 7

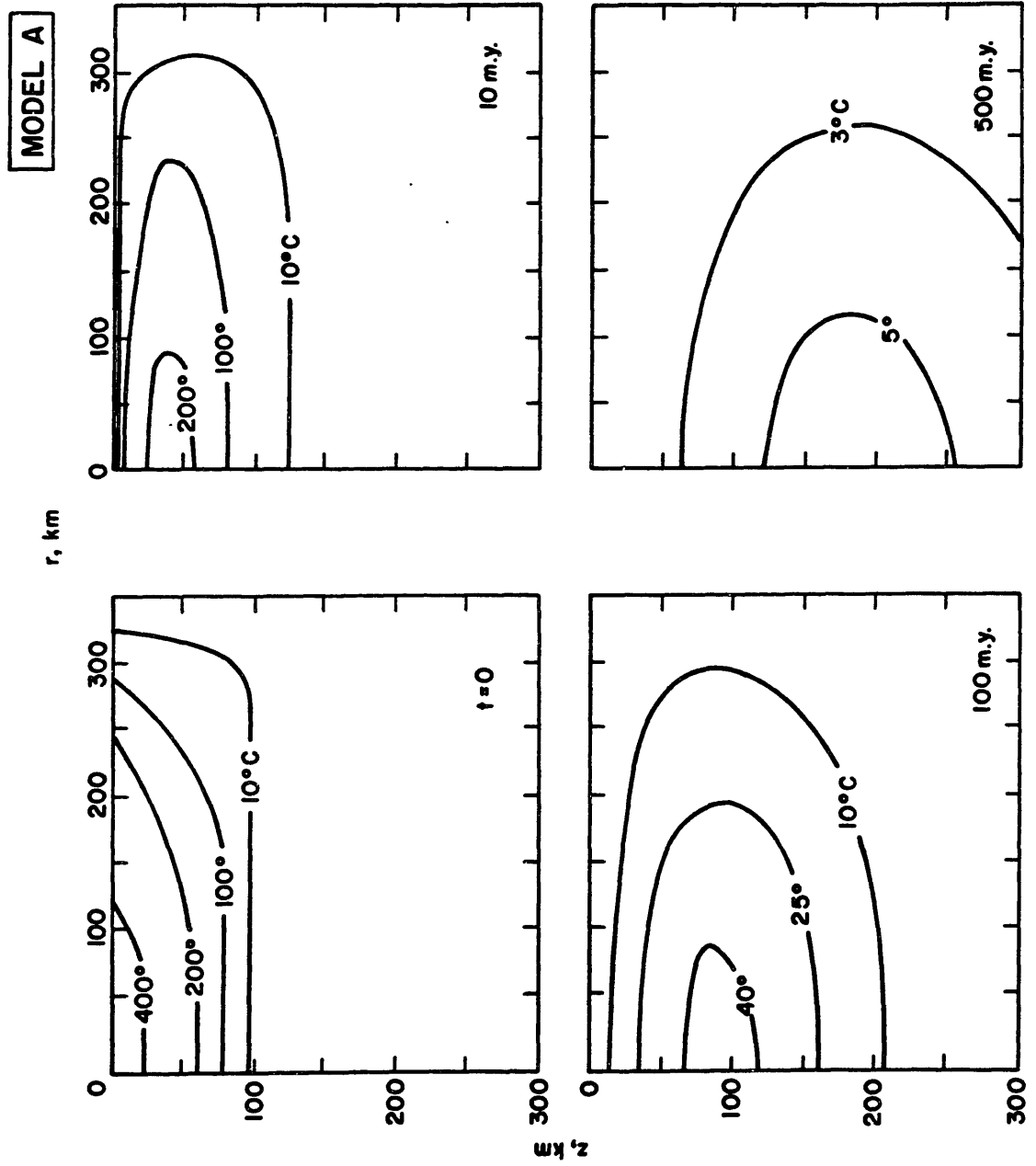


Figure 8

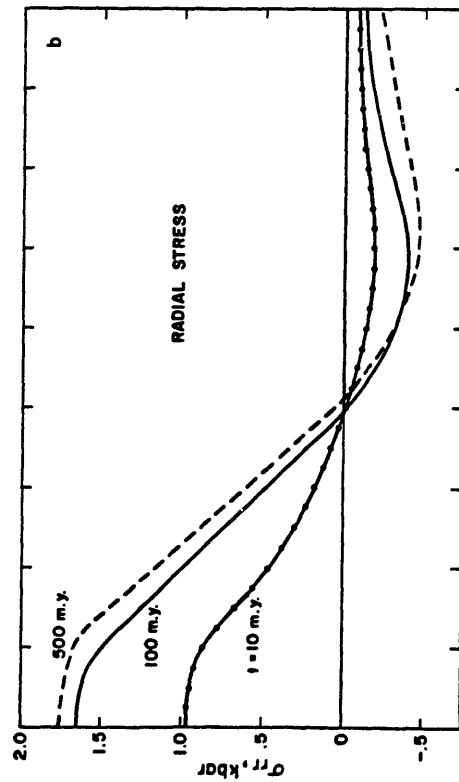
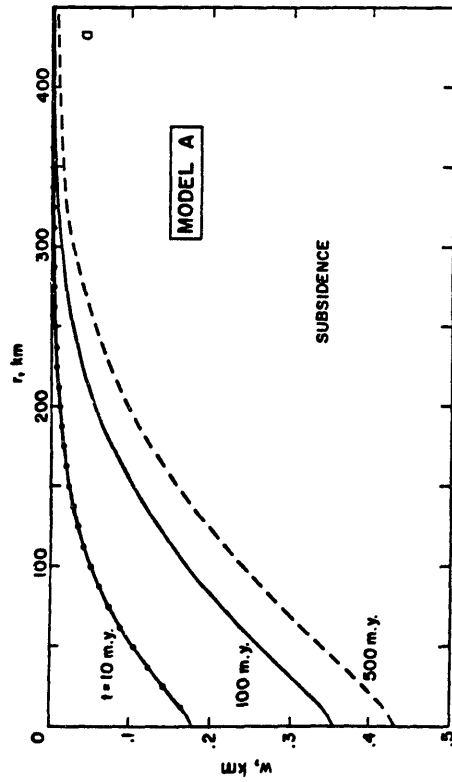
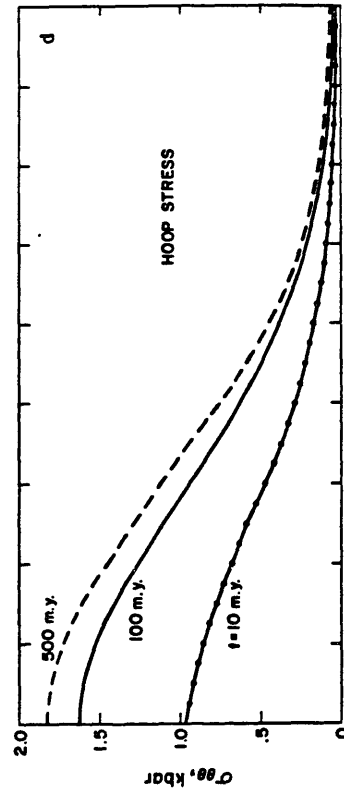
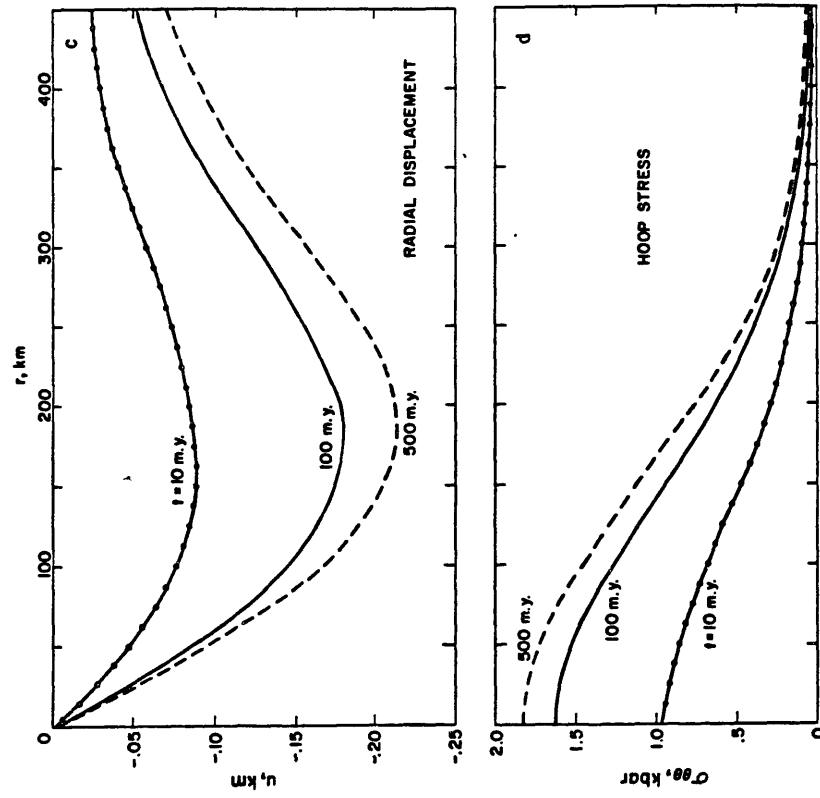


Figure 9

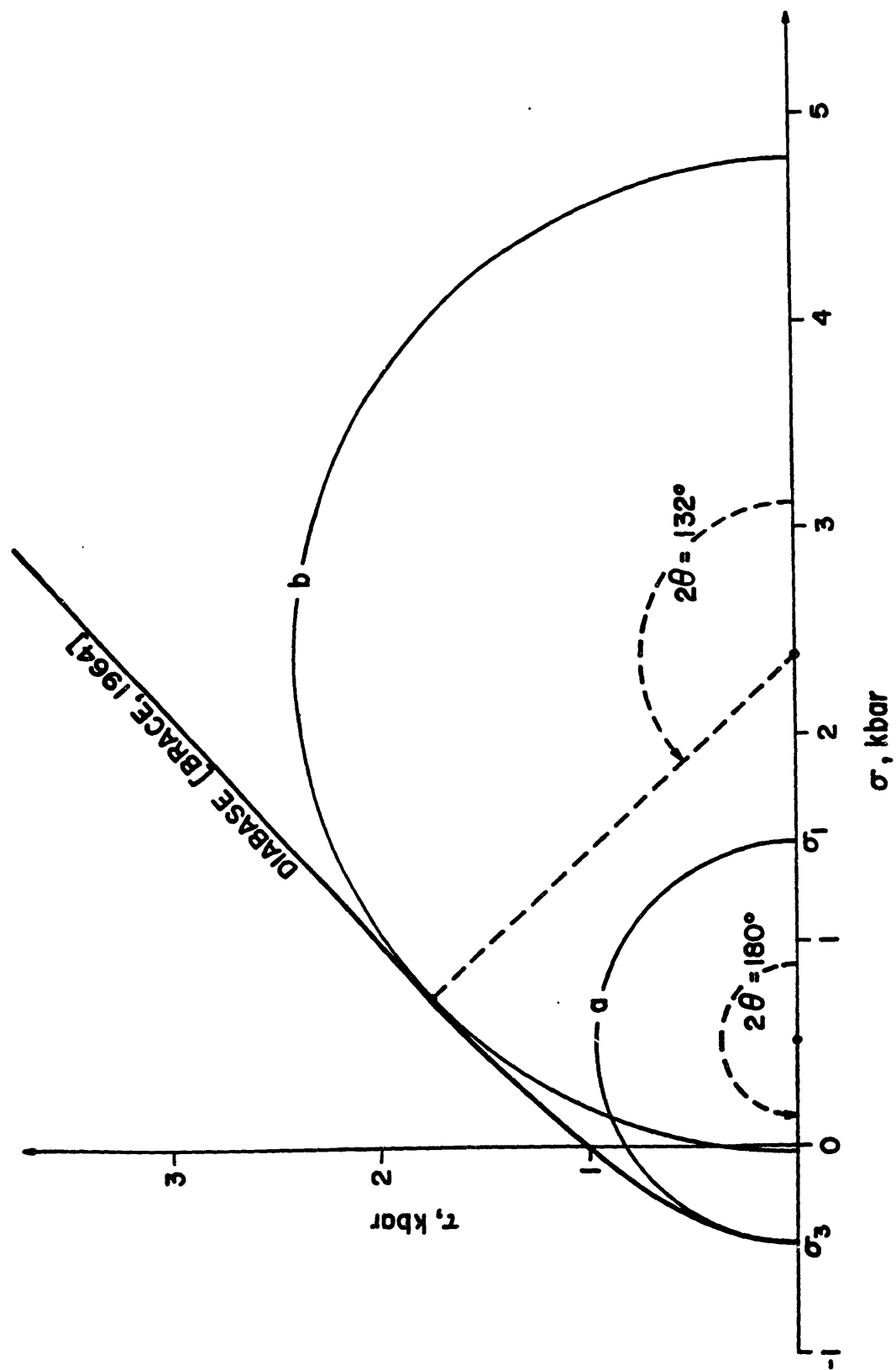


Figure 10

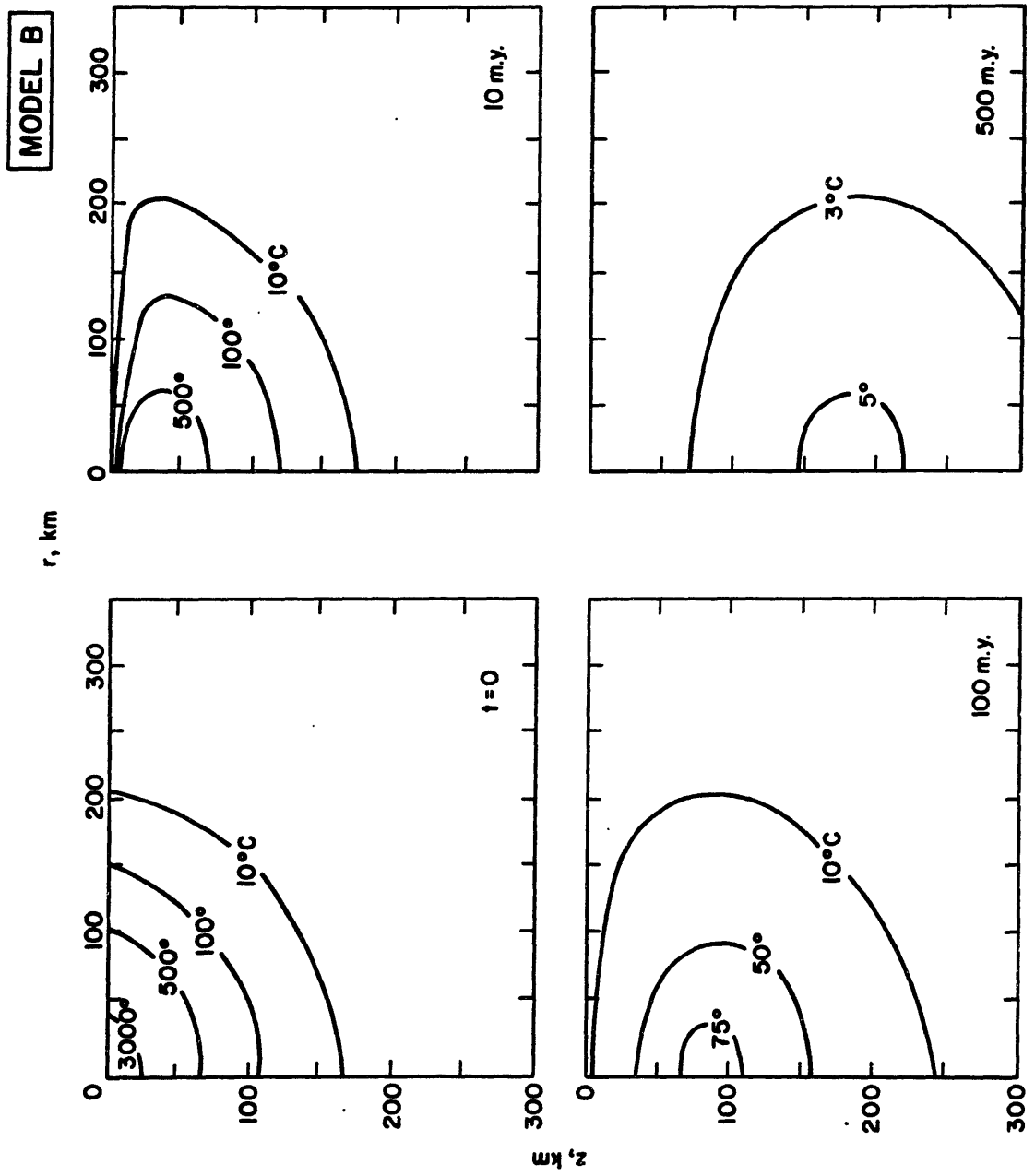


Figure 11

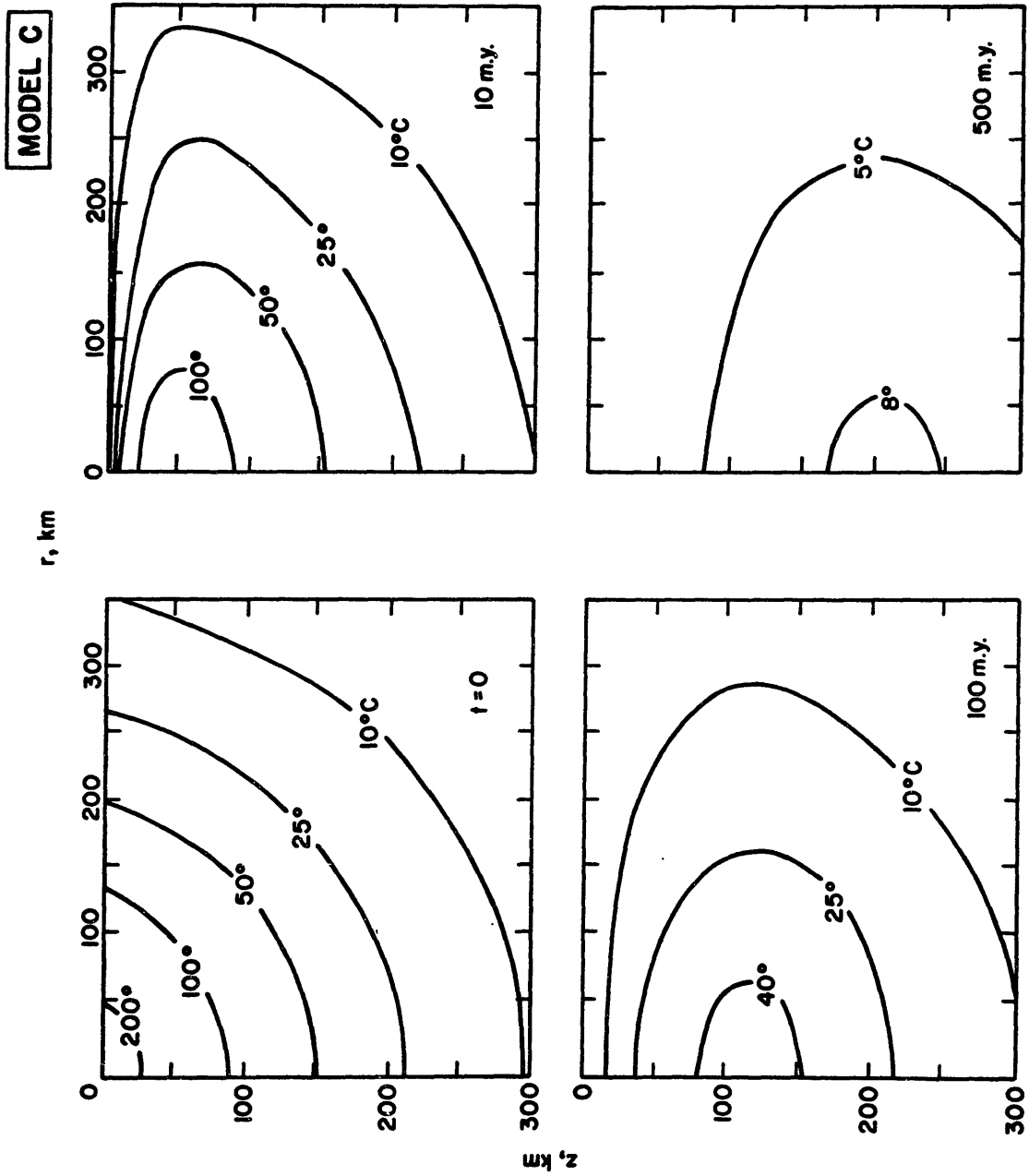


Figure 12

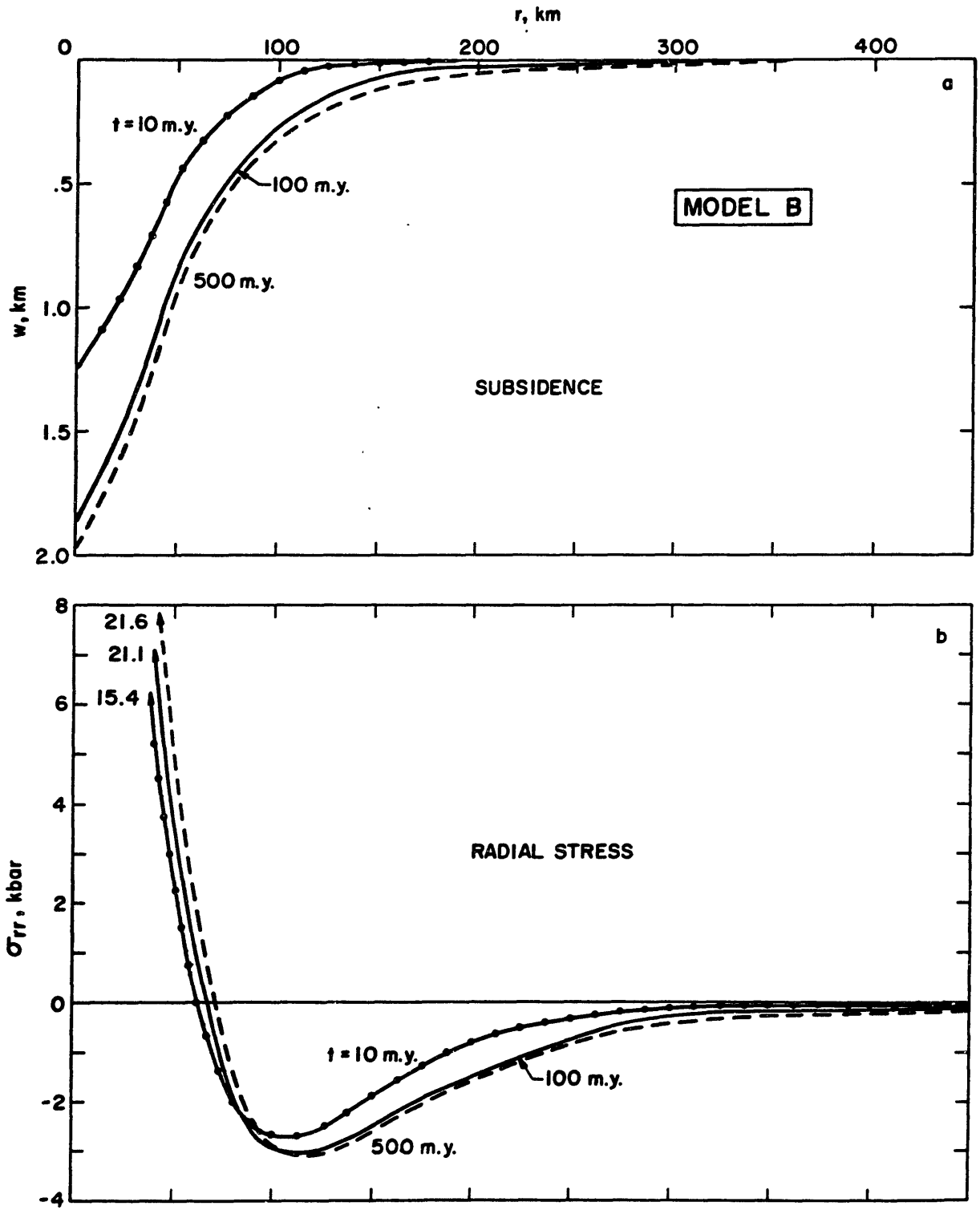


Figure 13

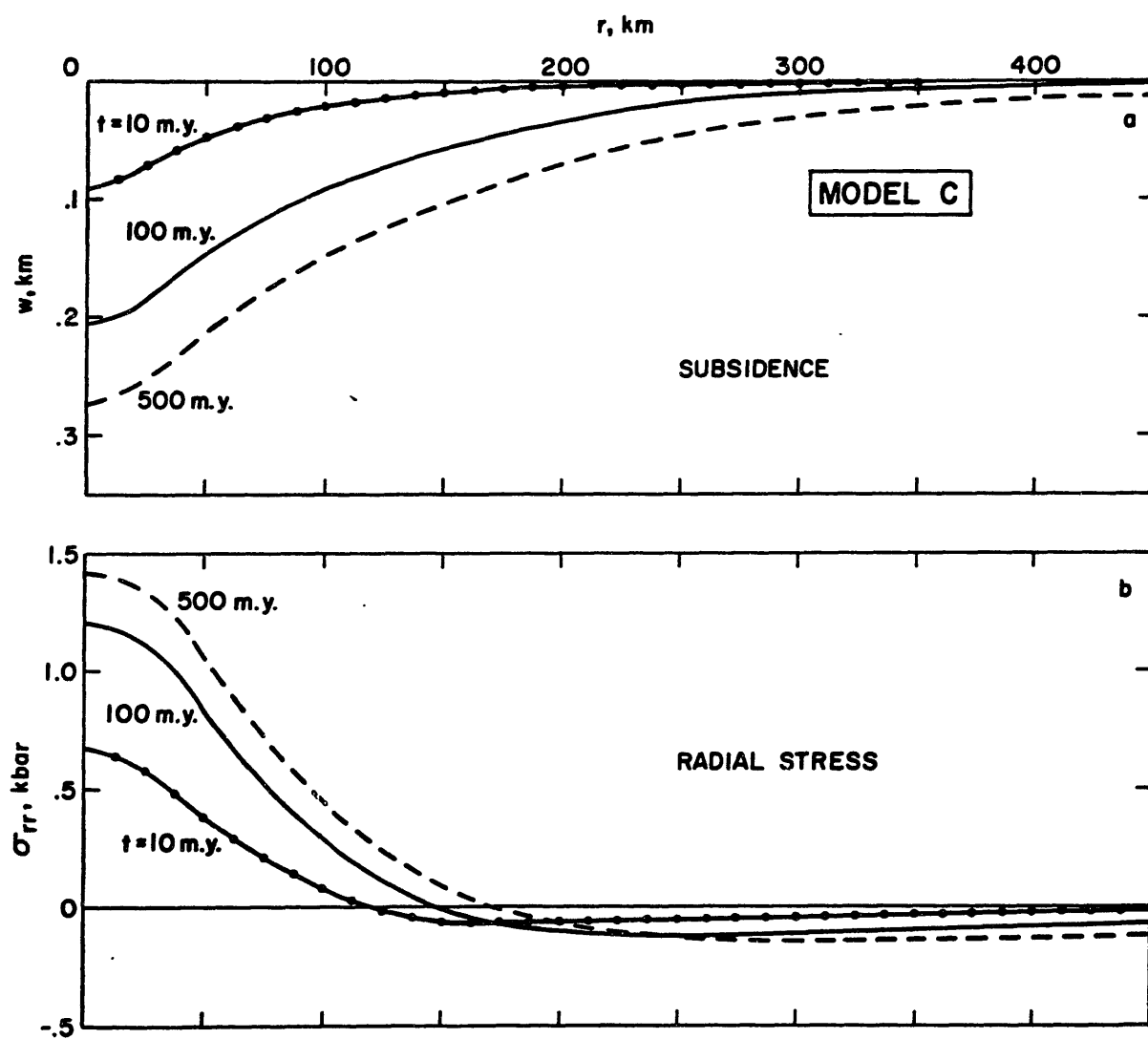


Figure 14

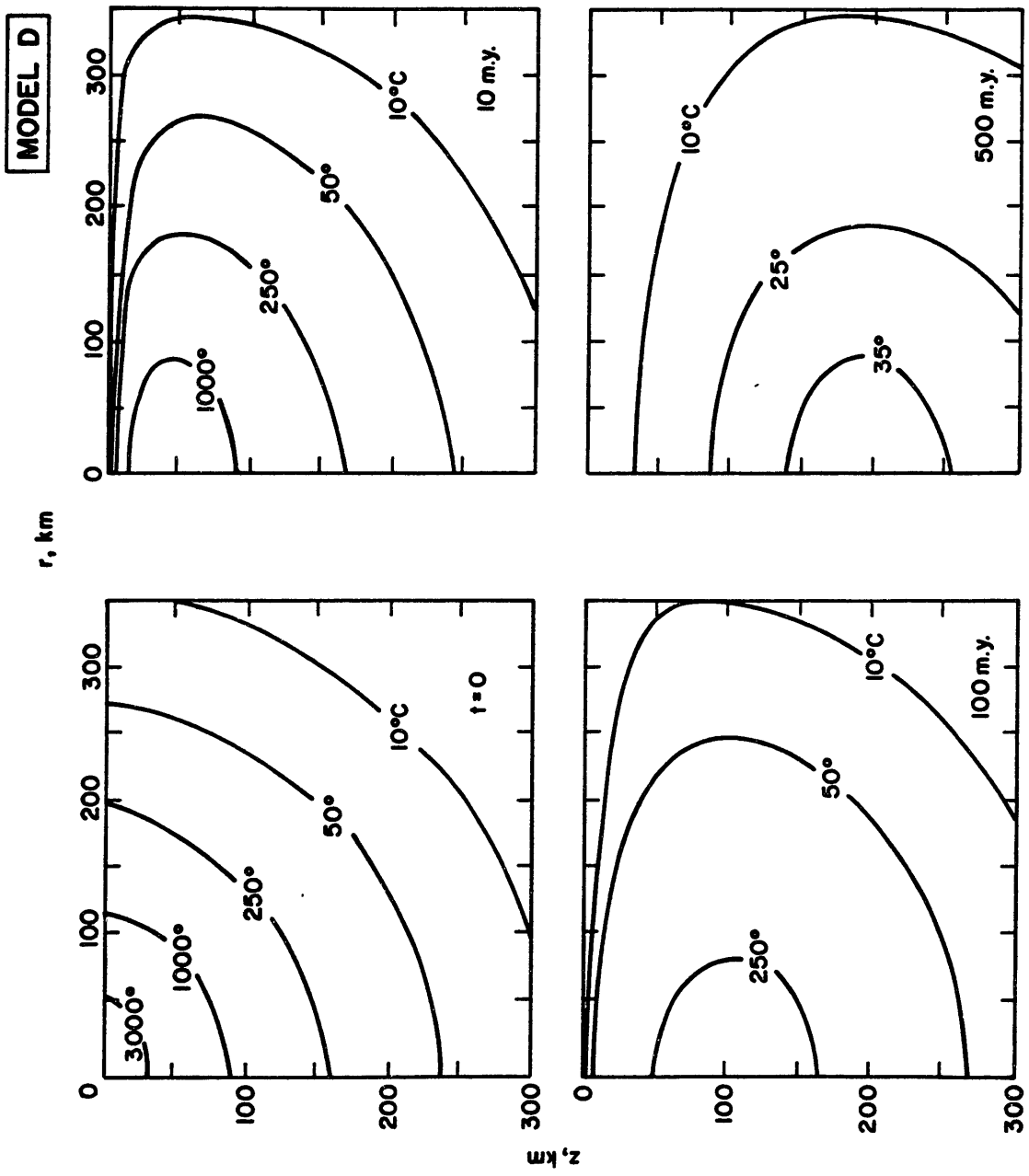


Figure 15

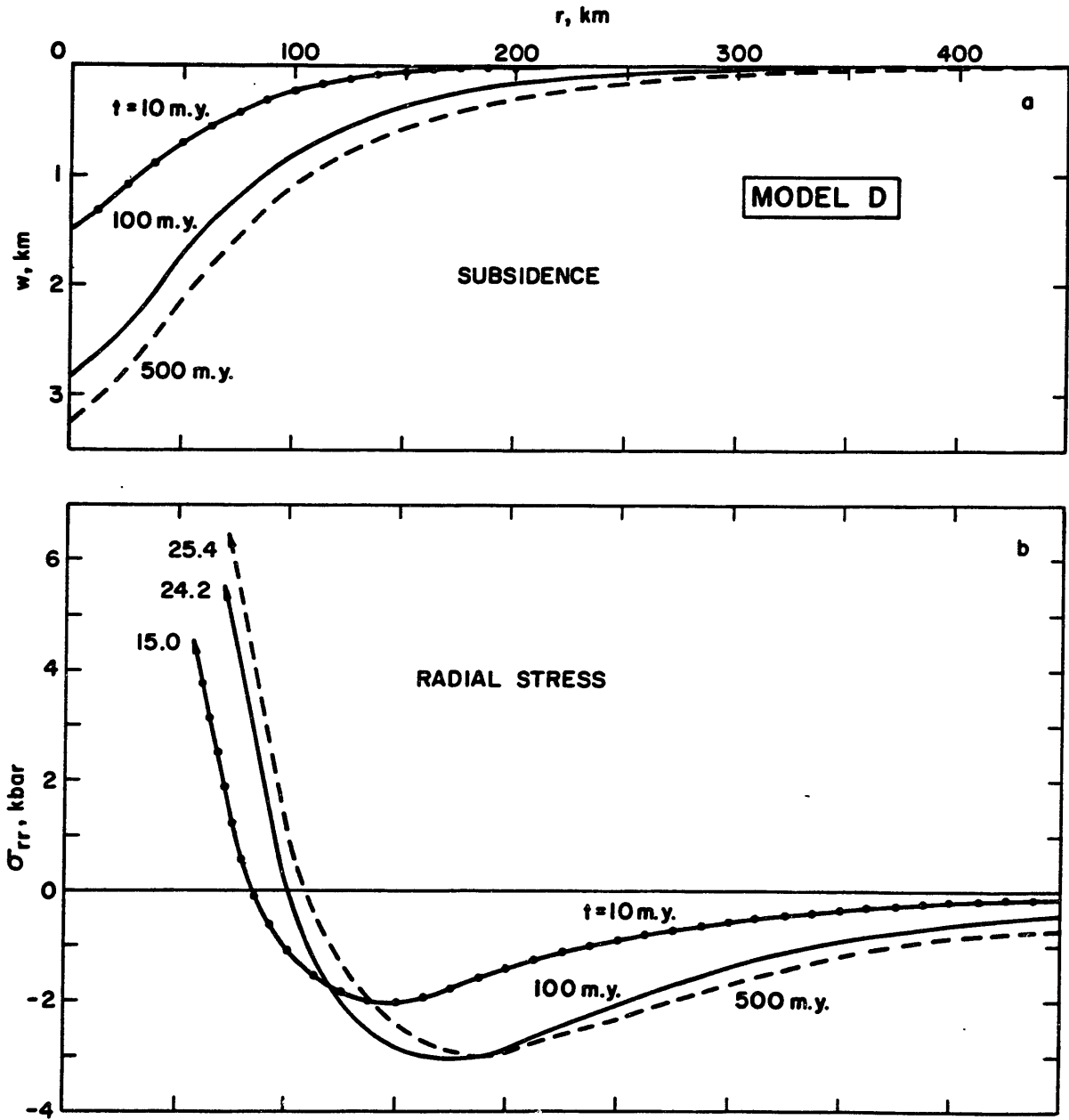


Figure 16

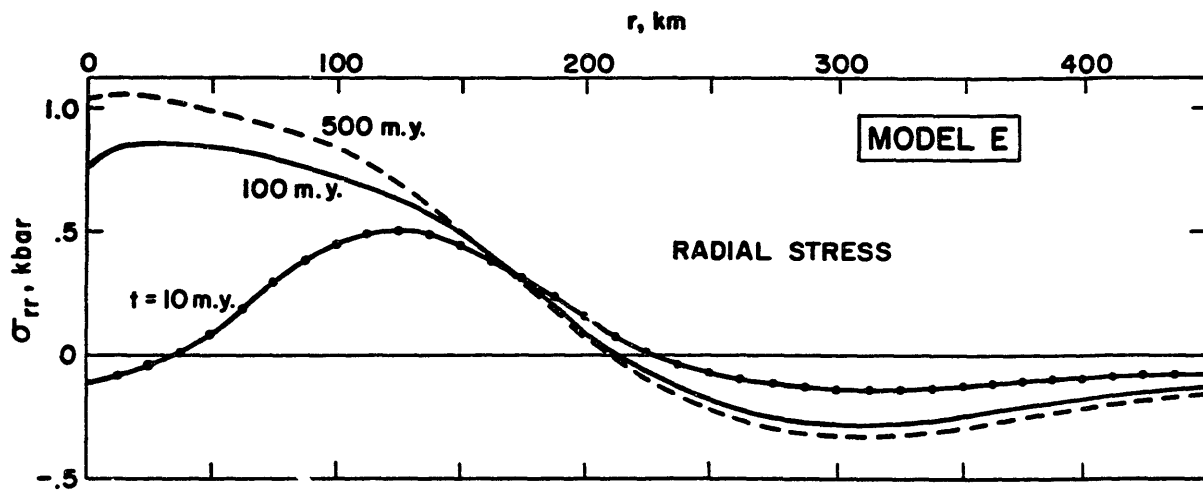


Figure 17

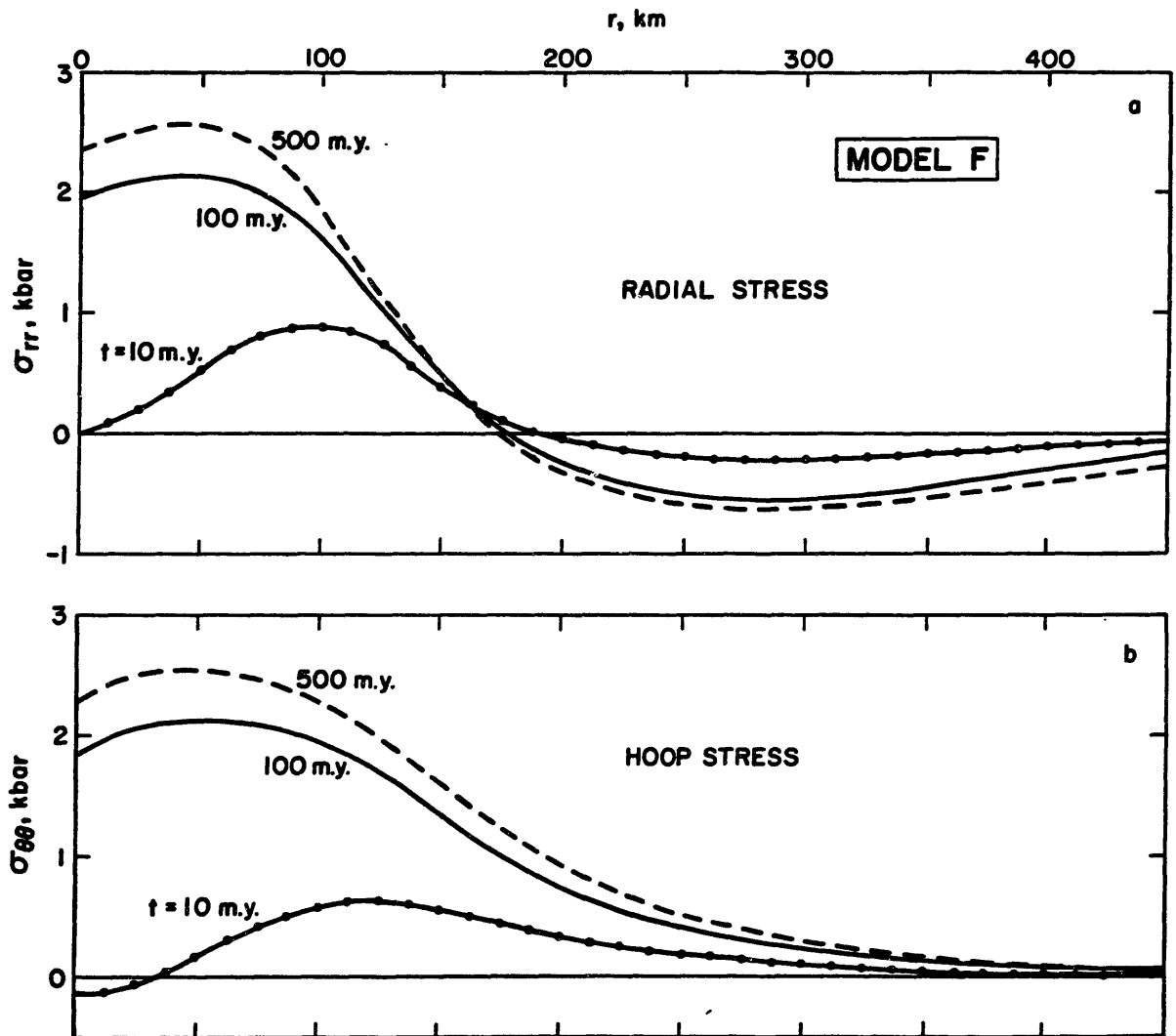


Figure 18

Chapter 7

CONCLUDING REMARKS

The major objectives of this thesis have been to examine the structure of young oceanic lithosphere and lunar impact basins, to interpret these structures in terms of the processes responsible for their formation, and to assess the contribution of thermal stress to the tectonics of oceanic and lunar lithosphere. In the broadest sense, it was found that the lithosphere in both environments is definably heterogenous and that the processes which acted to shape them varied in time and space. Also, the response of simple elastic models to oceanic and basin cooling suggests that thermal stress is an important factor in the tectonic history of the lithosphere in the vicinity of both mid-ocean ridges and lunar impact basins.

The examination of the seismic velocity structure of crust near the East Pacific Rise detailed in Chapter 2 provides new data on the nature and origin of young oceanic crust, but leaves many questions unanswered. Are thermal models for crustal accretion at fast-spreading ridges [e.g., Sleep, 1975], which generally predict a steady-state magma chamber along the rise axis, inadequate? Or is the absence of a sizable axial magma body at crustal levels in the ROSE area anomalous? Is the 5- to 10-km-wide magma chamber inferred to exist 2 km beneath the East Pacific Rise crest at 9°N [e.g., Herron et

al., 1978] real? Could magma be stored in crustal dikes or sills and still explain seismic and petrologic observations? The use of three-component ocean-bottom seismometers to examine shear and compressional energy that has traversed a mid-ocean ridge greatly increases the information available with which the presence and volume of crustal magma bodies can be assessed. The conversion of compressional to shear energy at the seafloor is, however, not well understood. Can crustal anisotropy explain the preferential conversion of P to S waves on a rise-crossing line discussed in Chapter 2 as compared to a rise-flanking line? A study of the energetics of mode conversion at the cracked and porous seafloor typical of young oceanic crust would provide insight into the observed variability in shear wave arrivals seen in the ROSE area and elsewhere. Reflection seismograms, SEABEAM bathymetric data, and observations from submersibles would have greatly enhanced our understanding of the tectonics of the ROSE area and of the ocean bottom environment which affects mode conversion beneath the shots and receivers and coupling of the receivers with the seafloor.

The advantages of recording refraction data with a series of identical, ocean-bottom instruments is evident from the data recorded on 0.5 m.y. old crust and presented in Chapter 3. The amplitude patterns and small changes in those patterns would have been difficult to detect and confirm had a variety of instruments with different coupling characteristics and

instrument responses been employed in this study. Comparison of the work of Chapter 3 to a similar study completed on 140 m.y. old crust [Purdy, 1983] provided strong evidence for an increase in the velocity of the uppermost crust with increasing age. Studies identical to that presented here and in Purdy [1983], but conducted along isochrons both younger and older than 0.5 m.y., could disclose structural trends not seen in previous studies using fewer or less uniform instrumentation. Questions to be answered by such studies include: At what age do velocities near the surface begin to increase? Over what period of time do these changes take place? Is the structural evolution of the crust dependent on plate velocity, plate age, or both? What would these findings tell us about hydrothermal circulation? A final question motivated by the work of Chapter 3 is the relative importance of metamorphism versus original igneous lithology in determining the seismic layering of oceanic crust. By deploying several ocean bottom instruments and conducting a thorough refraction experiment in conjunction with deep drilling, many questions about the geologic analogs of the observed seismic layering could be answered.

In Chapter 3, the thickness of seismic layer 2 was found to vary by no more than a kilometer over a 200 km long refraction line. However, the thickness of the crust varies by up to 80 km over the same lateral distance on the nearside of the Moon (Chapter 5). That the lunar lithosphere was thinned

by tens of kilometers in the seconds to minutes required for the formation of large basins attests to the great power of the impact process and to the mechanical and thermal energy transferred to all planets in the solar system during the period of intensive bombardment which ended almost 4 b.y. ago. More can be learned about the impact process by exploiting the full resolution of the lunar line-of-sight gravity data to improve upon the models of the deep structure of impact basins presented in Chapter 5. The results of this chapter pose questions that might be answered by a re-examination of the lunar photographic data set: Do differences in basin morphology (e.g., ring development) as a function of regional crustal thickness exist? Do these differences represent the preserved effects of a strength transition across the lunar Moho on the expanding impact cavity at the time of basin formation? Is the subdued relief of irregular mare basins a product of a difference in the basin formation process early in lunar history or a reflection of a relaxation process acting over longer time scales?

A major conclusion of Chapters 4 and 6, is that thermal stress is or has been a significant contributor to the topography and tectonics of young oceanic crust and lunar impact basins. This assessment was made using a series of simple, elastic half-space models in which the effects of temperature and pressure on the rheology of material at depth were accounted for only incompletely and artificially.

Future work will include visco-elastic models of the thermal stress field in cooling lithosphere and models in which stress is episodically relieved according to strength envelopes based on the calculated strain rates and differential stresses and an assumed rheology. Thermal stress may also be important to the state of stress in such other geologic settings as oceanic fracture zones, ridge-transform intersections, igneous intrusions and volcanic extrusions. An effort should be mounted to document geological and geophysical evidence (e.g., fracture patterns, seismicity, or physiographic features formed in response to contraction) for the contribution of thermal stress to the tectonics in such settings in order that the modeling of the thermal stress fields be worthwhile.

REFERENCES

- Herron, T.J., W.J. Ludwig, P.L. Stoffa, T.K. Kan, and P. Buhl, Structure of the East Pacific Rise crest from multichannel seismic reflection data, J. Geophys. Res., 83, 798-804, 1978.
- Purdy, G.M., The seismic structure of 140 m.y. old crust in the western central Atlantic Ocean, Geophys. J. R. Astron. Soc., 72, 115-138, 1983.
- Sleep, N.H., Formation of ocean crust: Some thermal constraints, J. Geophys. Res., 80, 4037-4042, 1975.

APPENDIX A
MASS DISK PARAMETERS AND TRIAXIAL GRAVITY COEFFICIENTS
USED TO COMPUTE THE LUNAR FREE-AIR GRAVITY ANOMALY

The free-air gravity anomaly field for the lunar nearside is calculated as the sum of the contributions from 350 mass disks [Wong et al., 1975] superimposed on a triaxial gravity field [Liu and Laing, 1971].

Presented in Table 1 of this appendix are the following disk parameters given by Wong et al. [1975]: latitude, longitude, mass of the disk (m divided by the mass of the Moon $m = 7.35 \times 10^{25}$ g), disk radius (a), and the radial location of the center of the disk relative to the lunar center of mass. the components of attraction of a single disk at the origin in the x-y plane (Cartesian coordinates) on a point in space at x, y, z can be approximated by taking the limit as the vertical thickness approaches zero of an oblate spheroid [Moulton, 1914]. The resulting components of attraction (A_x, A_y, A_z) as given by Wong et al. [1971] are:

$$\frac{A_x}{x} = \frac{A_y}{y} = \frac{3Gm}{2a^3} \{ -[k^{1/2}/(1+k)] + \sin^{-1}[1/(1+k)^{1/2}] \} \quad (A1)$$

$$\frac{A_z}{z} = -\frac{3Gm}{a^3} \{ 1/k^{1/2} - \sin^{-1}[1/(1+k)^{1/2}] \} \quad (A2)$$

where G is the gravitational constant and k is the solution of the quadratic equation:

$$k^2 a^2 + [a^2 - (x^2 + y^2 + z^2)]k - z^2 = 0. \quad (A3)$$

The gravitational attraction in space due to a distribution of disks in spherical coordinates can be easily computed from equations (A1-3) by rotating coordinates and summing the radial components.

The triaxial gravity field (ignoring the zero order term) at radius r from the Moon's center of mass and at latitude θ and longitude ϕ is given by

$$\frac{3GMR_m^2}{r^4} [C_{20}P_{20} + C_{22}P_{22} \cos 2\phi + S_{22}P_{22} \sin 2\phi] \quad (A4)$$

where R_m is the mean lunar radius (1737.42 km) and the non-normalized coefficients are given by Liu and Laing [1971] as

$$\begin{aligned} C_{20} &= -0.19963 \times 10^{-3} \\ C_{22} &= 0.23595 \times 10^{-4} \\ S_{22} &= 0.45383 \times 10^{-5} \end{aligned} \quad (A5)$$

and the Legendre functions are:

$$\begin{aligned} P_{20} &= \frac{3}{2} \sin^2 \theta - \frac{1}{2} \\ P_{22} &= 3 \cos^2 \theta \end{aligned} \quad (A6)$$

REFERENCES

- Liu, A.S., and P.A. Laing, Lunar gravity analysis from long-term effects, Science, 173, 1017-1020, 1971.
- Moulton, F.R., Celestial Mechanics, 2nd ed., 437 pp., Macmillan, New York, 1914.
- Wong, L., G. Buechler, W. Downs, W. Sjogren, P. Muller, and P. Gottlieb, A surface-layer representation of the lunar gravitational field, J. Geophys. Res., 76, 6220-6236, 1971.
- Wong, L., G. Buechler, W. Downs, W. Sjogren, and P. Muller, A lunar mass distribution from Apollo 14, 15, 16 CSM-LM and subsatellite radio Doppler data, Jet Propulsion Laboratory, 1975.

Appendix A, Table 1. Mass disk parameters (Wong et al., 1975).

number	latitude (deg)	longitude (deg)	mass (/lunar mass)	disk radius (km)	at lunar radius (km)
1	-2.0	86.0	0.44414E-05	180.	1729.
2	9.0	85.0	-0.12877E-05	65.	1736.
3	-18.0	96.0	0.35532E-07	60.	1736.
4	-23.0	92.0	-0.95662E-07	50.	1736.
5	-27.0	81.0	-0.61448E-07	90.	1735.
6	-21.5	79.0	-0.77084E-06	45.	1736.
7	-12.5	79.5	-0.92770E-07	40.	1736.
8	-20.0	70.0	-0.31595E-06	70.	1736.
9	-24.5	60.5	0.53336E-06	70.	1736.
10	-8.5	61.0	-0.23476E-05	50.	1736.
11	-16.0	61.5	-0.11561E-05	45.	1736.
12	17.0	58.5	0.12502E-04	210.	1726.
13	-16.0	34.0	0.80516E-05	200.	1727.
14	-11.5	26.0	-0.41135E-07	45.	1736.
15	-13.0	24.0	-0.12205E-05	40.	1736.
16	-21.0	33.3	-0.11274E-05	50.	1736.
17	-24.0	17.0	-0.53734E-07	40.	1736.
18	-24.0	-39.0	0.70379E-05	176.	1732.
19	9.5	-20.0	0.23808E-07	40.	1736.
20	-9.0	-1.5	-0.51451E-07	70.	1736.
21	-5.0	5.0	-0.25764E-07	65.	1736.
22	-11.5	4.0	-0.44419E-06	45.	1736.
23	-30.0	-13.5	-0.65777E-06	45.	1736.
24	-5.0	-68.0	0.79345E-06	65.	1736.
25	-20.0	-95.0	0.62353E-05	150.	1732.
26	-20.0	-95.0	-0.59017E-04	450.	1680.
27	-20.0	-95.0	0.70597E-04	630.	1620.
28	37.0	-19.0	0.28723E-04	450.	1680.
29	26.0	19.0	0.19507E-04	240.	1721.
30	2.0	-67.0	-0.92347E-06	40.	1736.
31	-2.5	-73.5	0.94833E-06	60.	1736.
32	-29.5	32.5	0.50127E-06	40.	1736.
33	-29.5	55.7	0.56165E-06	40.	1736.
34	-30.0	100.0	-0.17117E-05	75.	1736.
35	-30.0	95.0	0.76646E-06	75.	1736.
36	-30.0	90.0	0.34045E-05	75.	1736.
37	-30.0	85.0	-0.19041E-05	75.	1736.
38	-30.0	75.0	-0.25998E-06	75.	1736.
39	-30.0	70.0	-0.70436E-06	75.	1736.
40	-30.0	65.0	0.76885E-06	75.	1736.

```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)          (deg)          (/lunar mass) (km)          (km)
*****
41          -30.0          60.0          -0.25318E-08 75.          1736.
42          -30.0          50.0          0.14769E-06 75.          1736.
43          -30.0          45.0          0.12987E-06 75.          1736.
44          -30.0          40.0          0.93366E-06 75.          1736.
45          -30.0          25.0          0.82455E-06 75.          1736.
46          -30.0          20.0          0.47538E-06 75.          1736.
47          -30.0          15.0          0.66065E-06 75.          1736.
48          -30.0          10.0          0.86220E-07 75.          1736.
49          -30.0          5.0          0.27163E-06 75.          1736.
50          -30.0          0.0          0.20992E-06 75.          1736.
51          -30.0          -5.0          -0.23945E-07 75.          1736.
52          -30.0          -10.0          -0.10431E-06 75.          1736.
53          -30.0          -20.0          -0.63109E-06 75.          1736.
54          -30.0          -25.0          0.11710E-07 75.          1736.
55          -30.0          -30.0          -0.18155E-06 75.          1736.
56          -30.0          -35.0          -0.63939E-06 75.          1736.
57          -30.0          -40.0          -0.46329E-06 75.          1736.
58          -30.0          -45.0          -0.86306E-06 75.          1736.
59          -30.0          -50.0          0.96083E-06 75.          1736.
60          -30.0          -55.0          0.25390E-06 75.          1736.
61          -30.0          -60.0          -0.16806E-06 75.          1736.
62          -30.0          -65.0          -0.19986E-06 75.          1736.
63          -30.0          -70.0          0.15595E-05 75.          1736.
64          -25.0          100.0          0.16166E-05 75.          1736.
65          -23.5          86.5          -0.58390E-08 75.          1736.
66          -25.0          75.0          -0.84811E-07 75.          1736.
67          -25.0          70.0          0.95891E-06 75.          1736.
68          -25.0          65.0          -0.64162E-06 75.          1736.
69          -25.0          55.0          -0.31593E-06 75.          1736.
70          -25.0          50.0          0.53796E-06 75.          1736.
71          -25.0          45.0          0.44196E-06 75.          1736.
72          -25.0          40.0          -0.43924E-06 75.          1736.
73          -25.0          35.0          -0.39671E-07 75.          1736.
74          -25.0          30.0          0.48771E-07 75.          1736.
75          -25.0          25.0          0.87887E-06 75.          1736.
76          -25.0          20.0          0.73101E-06 75.          1736.
77          -25.0          10.0          0.29742E-06 75.          1736.
78          -25.0          5.0          -0.17162E-07 75.          1736.
79          -25.0          0.0          0.61126E-07 75.          1736.
80          -25.0          -5.0          -0.41723E-07 75.          1736.
*****

```

```

*****
number      latitude      longitude      mass          disk radius      at lunar radius
            (deg)         (deg)         (/lunar mass) (km)             (km)
*****
81          -25.0         -10.0         0.20596E-06   75.             1736.
82          -25.0         -15.0         0.13544E-07   75.             1736.
83          -25.0         -20.0         0.31346E-06   75.             1736.
84          -25.0         -25.0         -0.68794E-07   75.             1736.
85          -25.0         -30.0         -0.19800E-06   75.             1736.
86          -25.0         -45.0         -0.70859E-07   75.             1736.
87          -25.0         -50.0         -0.45405E-06   75.             1736.
88          -25.0         -55.0         0.43407E-06   75.             1736.
89          -25.0         -60.0         -0.37664E-07   75.             1736.
90          -25.0         -65.0         0.11786E-05   75.             1736.
91          -25.0         -70.0         -0.23018E-06   75.             1736.
92          -20.0         100.0         0.99282E-06   75.             1736.
93          -20.0         90.0         0.56463E-06   75.             1736.
94          -20.0         85.0         0.10855E-05   75.             1736.
95          -20.0         75.0         0.56127E-06   75.             1736.
96          -20.0         65.0         0.51673E-06   75.             1736.
97          -20.0         60.0         -0.51222E-06   75.             1736.
98          -20.0         55.0         0.55253E-07   75.             1736.
99          -20.0         50.0         -0.21240E-06   75.             1736.
100         -20.0         45.0         -0.49369E-06   75.             1736.
101         -20.0         25.0         -0.12960E-05   75.             1736.
102         -20.0         20.0         0.16775E-05   75.             1736.
103         -20.0         15.0         0.24122E-06   75.             1736.
104         -20.0         10.0         0.29389E-06   75.             1736.
105         -20.0         5.0         0.56383E-06   75.             1736.
106         -20.0         0.0         0.24152E-06   75.             1736.
107         -20.0         -5.0         -0.41866E-06   75.             1736.
108         -20.0         -10.0         -0.87283E-06   75.             1736.
109         -20.0         -15.0         -0.19821E-06   75.             1736.
110         -20.0         -20.0         -0.30292E-06   75.             1736.
111         -20.0         -25.0         0.36196E-06   75.             1736.
112         -20.0         -30.0         -0.55155E-06   75.             1736.
113         -20.0         -45.0         -0.20461E-06   75.             1736.
114         -20.0         -50.0         -0.56890E-06   75.             1736.
115         -20.0         -55.0         0.91066E-06   75.             1736.
116         -20.0         -60.0         0.22837E-05   75.             1736.
117         -20.0         -65.0         -0.71655E-06   75.             1736.
118         -20.0         -70.0         -0.15453E-05   75.             1736.
119         -15.0         100.0         -0.19908E-07   75.             1736.
120         -15.0         90.0         -0.74930E-06   75.             1736.
*****

```

```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)          (deg)          (/lunar mass) (km)            (km)
*****
121         -15.0          85.0         -0.95150E-07 75.             1736.
122         -15.0          80.0          0.32041E-06 75.             1736.
123         -15.0          75.0          0.65526E-06 75.             1736.
124         -15.0          70.0          0.27131E-06 75.             1736.
125         -15.0          65.0          0.11029E-06 75.             1736.
126         -15.0          55.0          0.29059E-07 75.             1736.
127         -15.0          50.0          0.49578E-06 75.             1736.
128         -15.0          45.0         -0.89961E-06 75.             1736.
129         -15.0          20.0          0.84215E-06 75.             1736.
130         -15.0          15.0          0.67683E-06 75.             1736.
131         -15.0          10.0          0.14712E-06 75.             1736.
132         -15.0           5.0          0.64517E-07 75.             1736.
133         -15.0           0.0          -0.81406E-07 75.             1736.
134         -15.0          -5.0         -0.56776E-06 75.             1736.
135         -15.0          -10.0        -0.47453E-06 75.             1736.
136         -15.0          -15.0        -0.56438E-06 75.             1736.
137         -15.0          -20.0        -0.57680E-06 75.             1736.
138         -15.0          -25.0        0.75284E-06 75.             1736.
139         -15.0          -30.0        0.45889E-06 75.             1736.
140         -15.0          -35.0        0.97720E-08 75.             1736.
141         -15.0          -40.0        -0.13403E-05 75.             1736.
142         -15.0          -45.0        -0.25801E-06 75.             1736.
143         -15.0          -50.0        0.80150E-06 75.             1736.
144         -15.0          -55.0        -0.12722E-06 75.             1736.
145         -15.0          -60.0        -0.27984E-05 75.             1736.
146         -15.0          -65.0        0.11114E-05 75.             1736.
147         -15.0          -70.0        -0.23416E-07 75.             1736.
148         -10.0          100.0       0.27994E-06 75.             1736.
149         -10.0          95.0        0.14372E-05 75.             1736.
150         -10.0          90.0        0.11758E-05 75.             1736.
151         -10.0          85.0        0.11212E-05 75.             1736.
152         -10.0          80.0        -0.32114E-06 75.             1736.
153         -10.0          75.0        0.98279E-06 75.             1736.
154         -10.0          70.0        -0.26857E-06 75.             1736.
155         -10.0          65.0        -0.56340E-06 75.             1736.
156         -10.0          55.0        -0.58034E-06 75.             1736.
157         -10.0          50.0        -0.74614E-06 75.             1736.
158         -10.0          45.0        -0.45620E-06 75.             1736.
159         -10.0          40.0        -0.19354E-05 75.             1736.
160         -10.0          35.0        -0.50966E-07 75.             1736.
*****

```

```

*****
number      latitude      longitude      mass          disk radius      at lunar radius
            (deg)         (deg)         (/lunar mass) (km)             (km)
*****
161         -10.0          30.0          0.65004E-06    75.             1736.
162         -10.0          20.0          0.26582E-06    75.             1736.
163         -10.0          15.0          -0.14262E-06   75.             1736.
164         -10.0          10.0          0.80109E-06    75.             1736.
165         -10.0          10.0          -0.13232E-05   75.             1736.
166         -10.0          15.0          0.50491E-06    75.             1736.
167         -10.0          20.0          -0.96157E-06   75.             1736.
168         -10.0          25.0          0.27706E-06    75.             1736.
169         -10.0          30.0          0.76342E-07    75.             1736.
170         -10.0          35.0          -0.60457E-08   75.             1736.
171         -10.0          40.0          0.73581E-06    75.             1736.
172         -10.0          45.0          0.11419E-06    75.             1736.
173         -10.0          50.0          -0.13664E-06   75.             1736.
174         -10.0          55.0          0.82892E-06    75.             1736.
175         -10.0          60.0          0.99958E-06    75.             1736.
176         -10.0          65.0          -0.14514E-07   75.             1736.
177         -10.0          70.0          -0.71630E-06   75.             1736.
178         -5.0           100.0         -0.50819E-06    75.             1736.
179         -5.0           95.0          -0.14403E-05    75.             1736.
180         -5.0           80.0          0.66699E-06    75.             1736.
181         -5.0           75.0          0.38758E-06    75.             1736.
182         -5.0           70.0          0.19524E-06    75.             1736.
183         -5.0           65.0          -0.40828E-06    75.             1736.
184         -5.0           60.0          0.15171E-06    75.             1736.
185         -5.0           55.0          0.60426E-06    75.             1736.
186         -5.0           50.0          0.54872E-06    75.             1736.
187         -5.0           45.0          0.27489E-06    75.             1736.
188         -5.0           40.0          -0.20180E-07    75.             1736.
189         -5.0           35.0          -0.94948E-06    75.             1736.
190         -5.0           30.0          -0.25337E-08    75.             1736.
191         -5.0           25.0          0.69333E-06    75.             1736.
192         -5.0           20.0          -0.84404E-06    75.             1736.
193         -5.0           15.0          0.16288E-06    75.             1736.
194         -5.0           10.0          0.80807E-07    75.             1736.
195         -5.0           0.0           -0.10735E-07    75.             1736.
196         -5.0           -5.0          -0.35860E-06    75.             1736.
197         -5.0           -10.0         -0.24067E-06    75.             1736.
198         -5.0           -15.0         -0.58673E-06    75.             1736.
199         -5.0           -20.0         -0.36266E-06    75.             1736.
200         -5.0           -25.0         -0.39576E-06    75.             1736.
*****

```

```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)         (deg)         (/lunar mass) (km)             (km)
*****
201         -5.0          -30.0         0.11357E-06      75.             1736.
202         -5.0          -35.0         0.97579E-06      75.             1736.
203         -5.0          -40.0         -0.15530E-07      75.             1736.
204         -5.0          -45.0         0.21258E-06      75.             1736.
205         -5.0          -50.0         -0.83809E-06      75.             1736.
206         -5.0          -55.0         0.19659E-06      75.             1736.
207         -5.0          -60.0         -0.15725E-05      75.             1736.
208         0.0           100.0         0.34240E-06      75.             1736.
209         0.0           95.0          0.14441E-07      75.             1736.
210         0.0           80.0         -0.28788E-06      75.             1736.
211         0.0           75.0          0.15771E-06      75.             1736.
212         0.0           70.0          0.14916E-06      75.             1736.
213         0.0           65.0         -0.42977E-06      75.             1736.
214         0.0           60.0         -0.13828E-07      75.             1736.
215         0.0           55.0          0.13814E-06      75.             1736.
216         0.0           50.0          0.16078E-06      75.             1736.
217         0.0           45.0         -0.13618E-06      75.             1736.
218         0.0           40.0         -0.59143E-06      75.             1736.
219         0.0           35.0         -0.56236E-06      75.             1736.
220         0.0           30.0         -0.26109E-06      75.             1736.
221         0.0           25.0         -0.18155E-06      75.             1736.
222         0.0           20.0         -0.37934E-06      75.             1736.
223         0.0           15.0         -0.48964E-06      75.             1736.
224         0.0           10.0         -0.35493E-06      75.             1736.
225         0.0            5.0          0.12885E-06      75.             1736.
226         0.0            0.0          0.58310E-06      75.             1736.
227         0.0           -5.0          0.78234E-07      75.             1736.
228         0.0          -10.0         0.63932E-07      75.             1736.
229         0.0          -15.0         -0.18293E-06      75.             1736.
230         0.0          -20.0         -0.10745E-06      75.             1736.
231         0.0          -25.0         -0.20388E-06      75.             1736.
232         0.0          -30.0         0.26959E-06      75.             1736.
233         0.0          -35.0         0.37159E-06      75.             1736.
234         0.0          -40.0         0.36082E-06      75.             1736.
235         0.0          -45.0         0.70664E-06      75.             1736.
236         0.0          -50.0        -0.58651E-06      75.             1736.
237         0.0          -55.0        -0.23523E-07      75.             1736.
238         0.0          -60.0         0.14433E-05      75.             1736.
239         0.0          -80.0         0.36621E-06      75.             1736.
240         5.0          100.0        0.20886E-05      75.             1736.
*****

```

```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)         (deg)         (/lunar mass) (km)             (km)
*****
241         5.0           95.0         -0.70782E-06      75.             1736.
242         5.0           80.0          0.12257E-06      75.             1736.
243         5.0           75.0          0.20380E-06      75.             1736.
244         5.0           70.0          0.69154E-08      75.             1736.
245         5.0           65.0          -0.81347E-07      75.             1736.
246         5.0           60.0          -0.35838E-06      75.             1736.
247         5.0           55.0          -0.38838E-06      75.             1736.
248         5.0           50.0          0.65996E-07      75.             1736.
249         5.0           45.0          0.12767E-06      75.             1736.
250         5.0           40.0          -0.40499E-06      75.             1736.
251         5.0           35.0          -0.45801E-06      75.             1736.
252         5.0           30.0          -0.67911E-06      75.             1736.
253         5.0           25.0          0.33758E-06      75.             1736.
254         5.0           20.0          0.20298E-06      75.             1736.
255         5.0           15.0          -0.62869E-06      75.             1736.
256         5.0           10.0          -0.39230E-06      75.             1736.
257         5.0            5.0          0.12648E-06      75.             1736.
258         5.0            0.0          -0.34489E-06      75.             1736.
259         5.0           -5.0          -0.16459E-07      75.             1736.
260         5.0          -10.0          0.39811E-06      75.             1736.
261         5.0          -15.0          0.42002E-06      75.             1736.
262         5.0          -20.0          0.19570E-06      75.             1736.
263         5.0          -25.0          0.69614E-08      75.             1736.
264         5.0          -30.0          0.16070E-06      75.             1736.
265         5.0          -35.0          0.17659E-07      75.             1736.
266         5.0          -40.0          -0.42583E-07      75.             1736.
267         5.0          -45.0          0.30164E-06      75.             1736.
268         5.0          -50.0          0.27021E-06      75.             1736.
269         5.0          -55.0          0.66759E-07      75.             1736.
270         5.0          -65.0          0.11473E-06      75.             1736.
271         5.0          -75.0          -0.33847E-06      75.             1736.
272         5.0          -80.0          0.47005E-06      75.             1736.
273         5.0          -85.0          0.28109E-06      75.             1736.
274         5.0          -90.0          0.15724E-06      75.             1736.
275         5.0          -95.0          -0.31664E-06      75.             1736.
276         5.0          -100.0         0.16109E-05      75.             1736.
277         5.0          -105.0         0.24359E-05      75.             1736.
278         10.0          100.0         0.68085E-06      75.             1736.
279         10.0           95.0          0.48653E-07      75.             1736.
280         10.0           90.0          -0.17529E-06      75.             1736.
*****

```

```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)          (deg)          (/lunar mass) (km)             (km)
*****
281          10.0           85.0          0.10777E-05      75.             1736.
282          10.0           80.0          0.28197E-06      75.             1736.
283          10.0           75.0          0.57635E-06      75.             1736.
284          10.0           70.0          0.10523E-06      75.             1736.
285          10.0           65.0          -0.12382E-05      75.             1736.
286          10.0           50.0          -0.11493E-05      75.             1736.
287          10.0           45.0          -0.52378E-06      75.             1736.
288          10.0           40.0          -0.59526E-06      75.             1736.
289          10.0           35.0          -0.64304E-06      75.             1736.
290          10.0           30.0          -0.33926E-06      75.             1736.
291          10.0           25.0          0.20163E-06      75.             1736.
292          10.0           20.0          -0.47099E-06      75.             1736.
293          10.0           15.0          -0.88621E-06      75.             1736.
294          10.0           10.0          -0.44779E-06      75.             1736.
295          10.0           5.0           0.31778E-07      75.             1736.
296          10.0           0.0           -0.35299E-06      75.             1736.
297          10.0           -5.0          -0.20467E-06      75.             1736.
298          10.0           -10.0         0.99764E-06      75.             1736.
299          10.0           -15.0         0.30674E-06      75.             1736.
300          10.0           -20.0         -0.31390E-06      75.             1736.
301          10.0           -25.0         -0.21688E-06      75.             1736.
302          10.0           -30.0         -0.15359E-06      75.             1736.
303          10.0           -35.0         -0.51674E-06      75.             1736.
304          10.0           -40.0         -0.83573E-06      75.             1736.
305          10.0           -45.0         0.13350E-06      75.             1736.
306          10.0           -50.0         -0.59781E-07      75.             1736.
307          10.0           -55.0         0.45897E-06      75.             1736.
308          10.0           -60.0         -0.39840E-06      75.             1736.
309          10.0           -65.0         -0.75584E-06      75.             1736.
310          10.0           -70.0         -0.17353E-06      75.             1736.
311          10.0           -75.0         -0.32902E-06      75.             1736.
312          10.0           -80.0         0.16356E-06      75.             1736.
313          10.0           -85.0         -0.18199E-06      75.             1736.
314          10.0           -90.0         0.53477E-06      75.             1736.
315          10.0           -95.0         -0.36804E-06      75.             1736.
316          10.0           -100.0        -0.27556E-05      75.             1736.
317          15.0           -105.0        -0.10912E-06      75.             1736.
318          15.0           -40.0         0.11280E-05      75.             1736.
319          20.0           -50.0         0.46313E-06      75.             1736.
320          20.0           85.0         0.16860E-05      75.             1736.
*****

```



```

*****
number      latitude      longitude      mass      disk radius      at lunar radius
            (deg)          (deg)          (/lunar mass) (km)             (km)
*****
321         20.0           75.0          0.12872E-05    75.             1736.
322         20.0           5.0           -0.15477E-05   75.             1736.
323         20.0          -5.0           -0.11989E-05   75.             1736.
324         22.0          50.0           -0.11432E-06   75.             1736.
325         23.0          45.0           0.27501E-06    75.             1736.
326         24.0          40.0           0.65365E-06    75.             1736.
327         25.0          35.0           -0.72726E-06   75.             1736.
328         25.0          30.0           -0.24487E-06   75.             1736.
329         25.0           0.0           -0.19688E-06   75.             1736.
330         25.0          -5.0           -0.30619E-06   75.             1736.
331         25.0          -10.0          -0.70132E-06   75.             1736.
332         30.0           5.0           -0.35853E-06   75.             1736.
333         20.0          -15.0          -0.27656E-06   150.            1732.
334         20.0          -25.0          0.13972E-05    150.            1732.
335         20.0          -35.0          0.20697E-05    150.            1732.
336         20.0          -45.0          0.38081E-06    150.            1732.
337         20.0          -55.0          -0.85253E-07   150.            1732.
338         20.0          -65.0          -0.90106E-06   150.            1732.
339         20.0          -75.0          -0.21569E-06   150.            1732.
340         20.0          -85.0          -0.11857E-05   150.            1732.
341         20.0          -95.0          -0.57200E-06   150.            1732.
342         20.0          -105.0         -0.24441E-05   150.            1732.
343         30.0          -35.0          0.35236E-06    150.            1732.
344         30.0          -45.0          0.34998E-06    150.            1732.
345         30.0          -55.0          0.12964E-06    150.            1732.
346         30.0          -65.0          -0.30242E-07   150.            1732.
347         30.0          -75.0          0.13339E-05    150.            1732.
348         30.0          -85.0          0.40594E-05    150.            1732.
349         30.0          -95.0          0.14428E-05    150.            1732.
350         30.0          -105.0         0.51751E-05    150.            1732.
*****

```

APPENDIX B

5° BY 5° AVERAGES OF LUNAR NEAR SIDE TOPOGRAPHY

The topographic data set used in the study of lunar crustal structure in Chapter 5 is given in Table 1 of this appendix. These data are presented in the form of 5° latitude by 5° longitude averages of observations of nearside topography from a variety of sources (see text of Chapter 5). The synthesis and averaging of most of the measurements were done by B.G. Bills [personal communication, 1982]. Bills' data are indicated by the number 1 under the "source" column in Table 1. Also given is the number of observations incorporated into each of his averages. In regions where the topography is poorly resolved, values from the harmonic representation of Bills and Ferrari [1977] (source = 2) supplement the block averages. Topography over Orientale is taken from Head et al. [1981] (source = 3), but required a downward adjustment of about 2 km to fit smoothly with the topographic measurements of the surrounding terrain. Discussion of the uncertainty of these data is given in Chapter 5. Finally, the percent of each block covered by mare units as used in the inversion procedure of Chapter 5 is given in Table 1.

REFERENCES

- Bills, B.G. and A.J. Ferrari, A harmonic analysis of lunar topography, Icarus, 31, 244-259, 1977.
- Head, J.W., E. Robinson, and R.J. Phillips, Topography of the Orientale basin (abstract), Lunar Planet. Sci., 12, 421-423, 1981.

Appendix B, Table 1. Averages of lunar topography over 5 by 5 blocks centered at latitude and longitude given.

number	latitude (deg)	longitude (deg)	lunar radius (km)	maria cover (%)	number of observations	source
1	37.5	-102.5	1737.517	0.	0	2
2	37.5	-97.5	1737.441	0.	2	1
3	37.5	-92.5	1737.616	0.	2	1
4	37.5	-87.5	1737.626	0.	1	1
5	37.5	-82.5	1737.311	0.	0	2
6	37.5	-77.5	1737.332	25.	0	2
7	37.5	-72.5	1736.105	100.	0	2
8	37.5	-67.5	1735.120	100.	1	1
9	37.5	-62.5	1734.835	100.	0	2
10	37.5	-57.5	1734.195	100.	5	1
11	37.5	-52.5	1734.530	100.	1	1
12	37.5	-47.5	1734.738	75.	1	1
13	37.5	-42.5	1735.109	0.	0	2
14	37.5	-37.5	1735.560	50.	5	1
15	37.5	-32.5	1735.184	100.	4	1
16	37.5	-27.5	1735.324	100.	7	1
17	37.5	-22.5	1734.944	100.	9	1
18	37.5	-17.5	1735.120	100.	9	1
19	37.5	-12.5	1734.582	100.	8	1
20	37.5	-7.5	1735.034	100.	12	1
21	37.5	-2.5	1734.767	100.	10	1
22	37.5	2.5	1734.574	75.	6	1
23	37.5	7.5	1735.667	25.	6	1
24	37.5	12.5	1735.747	50.	8	1
25	37.5	17.5	1734.969	25.	4	1
26	37.5	22.5	1735.183	50.	7	1
27	37.5	27.5	1735.177	100.	4	1
28	37.5	32.5	1736.104	50.	5	1
29	37.5	37.5	1735.698	50.	8	1
30	37.5	42.5	1736.072	0.	5	1
31	37.5	47.5	1737.052	0.	1	1
32	37.5	52.5	1736.808	0.	0	2
33	37.5	57.5	1736.788	0.	0	2
34	37.5	62.5	1736.828	0.	0	2
35	37.5	67.5	1736.176	0.	4	1
36	37.5	72.5	1735.998	0.	1	1
37	37.5	77.5	1736.725	0.	0	2
38	37.5	82.5	1736.419	0.	0	2
39	37.5	87.5	1736.117	0.	5	1
40	37.5	92.5	1736.337	0.	5	1

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
41          37.5           97.5           1736.069           0.              6              1
42          37.5           102.5          1735.916           0.              5              1
43          32.5          -102.5          1737.751           0.              0              2
44          32.5           -97.5           1738.116           0.              0              2
45          32.5           -92.5           1738.367           0.              0              2
46          32.5           -87.5           1738.444           0.              0              1
47          32.5           -82.5           1737.471           0.              0              2
48          32.5           -77.5           1737.032           75.             0              2
49          32.5           -72.5           1735.839           100.            0              2
50          32.5           -67.5           1734.815           75.             0              2
51          32.5           -62.5           1734.901           75.             5              1
52          32.5           -57.5           1735.028           100.            5              1
53          32.5           -52.5           1735.156           100.            10              1
54          32.5           -47.5           1735.041           100.            10              1
55          32.5           -42.5           1735.229           100.            15              1
56          32.5           -37.5           1735.307           100.            23              1
57          32.5           -32.5           1734.930           100.            13              1
58          32.5           -27.5           1735.234           100.            11              1
59          32.5           -22.5           1735.202           100.            20              1
60          32.5           -17.5           1734.943           100.            14              1
61          32.5           -12.5           1734.950           100.            10              1
62          32.5           -7.5            1735.099           75.             17              1
63          32.5           -2.5            1734.996           50.             18              1
64          32.5            2.5            1735.273           25.             12              1
65          32.5            7.5            1734.858           50.              7              1
66          32.5           12.5           1734.678           100.            13              1
67          32.5           17.5           1734.865           100.             6              1
68          32.5           22.5           1734.882           100.            11              1
69          32.5           27.5           1734.863           50.            15              1
70          32.5           32.5           1736.683           25.              4              1
71          32.5           37.5           1736.213           50.             11              1
72          32.5           42.5           1736.641           0.              2              1
73          32.5           47.5           1736.214           0.             17              1
74          32.5           52.5           1736.484           0.              9              1
75          32.5           57.5           1736.427           0.              0              2
76          32.5           62.5           1735.668           25.              3              1
77          32.5           67.5           1735.840           25.              2              1
78          32.5           72.5           1736.260           0.              0              2
79          32.5           77.5           1736.419           0.              0              1
80          32.5           82.5           1736.851           0.              5              1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
81          32.5           87.5           1735.863           0.              6              1
82          32.5           92.5           1736.117           0.              10             1
83          32.5           97.5           1736.053           0.              12             1
84          32.5           102.5          1736.072           0.              12             1
85          27.5           -102.5         1737.970           0.              0              2
86          27.5           -97.5         1737.942           0.              2              1
87          27.5           -92.5         1737.559           0.              2              1
88          27.5           -87.5         1737.754           0.              1              1
89          27.5           -82.5         1736.611           50.             2              1
90          27.5           -77.5         1736.630           75.             0              2
91          27.5           -72.5         1735.658           75.             0              2
92          27.5           -67.5         1735.462           75.             1              1
93          27.5           -62.5         1734.864           100.            21             1
94          27.5           -57.5         1735.112           100.            57             1
95          27.5           -52.5         1735.535           50.             50             1
96          27.5           -47.5         1735.603           50.             52             1
97          27.5           -42.5         1735.691           50.             58             1
98          27.5           -37.5         1735.641           100.            64             1
99          27.5           -32.5         1735.701           100.            43             1
100         27.5           -27.5         1735.590           100.            84             1
101         27.5           -22.5         1735.384           100.           102             1
102         27.5           -17.5         1735.255           100.            82             1
103         27.5           -12.5         1735.093           75.             56             1
104         27.5           -7.5          1735.596           25.             71             1
105         27.5           -2.5          1735.690           50.             77             1
106         27.5           2.5           1735.769           25.             89             1
107         27.5           7.5           1735.502           75.             89             1
108         27.5           12.5          1734.857           100.            63             1
109         27.5           17.5          1735.003           100.            60             1
110         27.5           22.5          1734.627           100.            39             1
111         27.5           27.5          1734.510           100.            32             1
112         27.5           32.5          1735.848           100.            16             1
113         27.5           37.5          1735.814           0.              0              2
114         27.5           42.5          1736.857           0.              11             1
115         27.5           47.5          1736.322           0.              0              2
116         27.5           52.5          1736.465           0.              0              1
117         27.5           57.5          1735.750           0.              17             1
118         27.5           62.5          1735.438           0.              14             1
119         27.5           67.5          1735.514           25.             9              1
120         27.5           72.5          1735.514           0.              0              2
120         27.5           72.5          1735.754           0.              0              2
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)            (%)              observations
*****
121         27.5           77.5          1735.941          0.              7              1
122         27.5           82.5          1736.075          0.              0              2
123         27.5           87.5          1736.714          0.             13              1
124         27.5           92.5          1736.820          0.             18              1
125         27.5           97.5          1736.276          0.             35              1
126         27.5          102.5          1736.103          0.             45              1
127         22.5          -102.5          1738.262          0.              0              2
128         22.5          -97.5          1737.624          0.              2              1
129         22.5          -92.5          1737.985          0.              2              1
130         22.5          -87.5          1737.544          0.              0              2
131         22.5          -82.5          1737.407          25.             2              1
132         22.5          -77.5          1736.382          75.             2              2
133         22.5          -72.5          1735.706          50.             0              2
134         22.5          -67.5          1735.505          75.             2              1
135         22.5          -62.5          1734.851          100.            10              1
136         22.5          -57.5          1735.155          100.            16              1
137         22.5          -52.5          1735.495          75.             24              1
138         22.5          -47.5          1735.910          50.             36              1
139         22.5          -42.5          1735.843          100.            60              1
140         22.5          -37.5          1735.878          100.            88              1
141         22.5          -32.5          1735.984          100.           96              1
142         22.5          -27.5          1735.877          75.           107              1
143         22.5          -22.5          1735.601          100.          110              1
144         22.5          -17.5          1735.380          100.           90              1
145         22.5          -12.5          1735.156          100.          101              1
146         22.5           -7.5          1735.367          75.           96              1
147         22.5           -2.5          1735.849          25.          110              1
148         22.5            2.5          1736.421          0.              0              2
149         22.5            7.5          1736.179          25.             0              1
150         22.5           12.5          1734.945          100.            18              1
151         22.5           17.5          1735.184          100.            75              1
152         22.5           22.5          1734.998          100.            75              1
153         22.5           27.5          1734.682          100.            63              1
154         22.5           32.5          1735.941          25.             61              1
155         22.5           37.5          1736.052          50.             46              1
156         22.5           42.5          1736.582          25.             36              1
157         22.5           47.5          1736.495          0.              30              1
158         22.5           52.5          1735.974          50.              32              1
159         22.5           57.5          1735.144          75.              21              1
160         22.5           62.5          1735.188          50.              17              1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
161          22.5           67.5           1735.751          50.              17              1
162          22.5           72.5           1736.773           0.              19              1
163          22.5           77.5           1736.082           0.              0               2
164          22.5           82.5           1736.076           0.              38              1
165          22.5           87.5           1736.755           25.             43              1
166          22.5           92.5           1736.808           0.              0               2
167          22.5           97.5           1736.916           0.              32              1
168          22.5          102.5          1737.843           0.              27              1
169          17.5          -102.5          1738.764           0.              0               2
170          17.5          -97.5           1738.363           0.              0               2
171          17.5          -92.5           1737.908           0.              0               2
172          17.5          -87.5           1737.112           0.              0               2
173          17.5          -82.5           1736.228           0.              0               2
174          17.5          -77.5           1735.571           25.             0               2
175          17.5          -72.5           1735.758           75.             2               1
176          17.5          -67.5           1735.370           100.            25              1
177          17.5          -62.5           1735.288           100.            39              1
178          17.5          -57.5           1735.247           100.            49              1
179          17.5          -52.5           1735.584           100.            37              1
180          17.5          -47.5           1735.688           100.            27              1
181          17.5          -42.5           1735.861           100.            19              1
182          17.5          -37.5           1735.901           100.            15              1
183          17.5          -32.5           1736.312           100.            26              1
184          17.5          -27.5           1736.057           75.             42              1
185          17.5          -22.5           1736.207           75.             54              1
186          17.5          -17.5           1735.411           100.            49              1
187          17.5          -12.5           1735.560           75.             50              1
188          17.5           -7.5           1735.818           50.             48              1
189          17.5           -2.5           1736.606           0.              0               2
190          17.5           2.5           1736.727           50.             99              1
191          17.5           7.5           1736.482           50.            123              1
192          17.5          12.5           1735.956           75.            106              1
193          17.5          17.5           1735.499           75.            130              1
194          17.5          22.5           1735.186           75.            103              1
195          17.5          27.5           1735.164           75.            113              1
196          17.5          32.5           1735.956           25.            102              1
197          17.5          37.5           1736.063           100.            99              1
198          17.5          42.5           1736.379           0.              78              1
199          17.5          47.5           1736.625           0.              93              1
200          17.5          52.5           1733.453          100.             80              1
*****

```



```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)         (deg)         (km)             (%)             observations
*****
201         17.5          57.5          1734.340          100.            92            1
202         17.5          62.5          1734.216          100.            40            1
203         17.5          67.5          1735.034          75.             57            1
204         17.5          72.5          1736.476          0.              66            1
205         17.5          77.5          1736.376          0.              56            1
206         17.5          82.5          1736.525          50.             49            1
207         17.5          87.5          1736.621          25.             31            1
208         17.5          92.5          1737.313          0.              23            1
209         17.5          97.5          1737.067          0.              26            1
210         17.5          102.5         1738.653          0.              29            1
211         12.5          -102.5         1738.821          0.              0             2
212         12.5          -97.5          1738.255          0.              2             1
213         12.5          -92.5          1737.910          0.              0             2
214         12.5          -87.5          1737.409          0.              2             1
215         12.5          -82.5          1736.477          0.              22            1
216         12.5          -77.5          1735.817          25.             33            1
217         12.5          -72.5          1735.557          50.             38            1
218         12.5          -67.5          1735.586          100.            23            1
219         12.5          -62.5          1735.677          100.            18            1
220         12.5          -57.5          1736.285          100.             6            1
221         12.5          -52.5          1736.471          100.             9            1
222         12.5          -47.5          1736.007          100.            14            1
223         12.5          -42.5          1736.237          100.            26            1
224         12.5          -37.5          1736.165          100.            21            1
225         12.5          -32.5          1736.181          100.            31            1
226         12.5          -27.5          1736.058          50.             65            1
227         12.5          -22.5          1735.768          25.             51            1
228         12.5          -17.5          1735.610          75.             58            1
229         12.5          -12.5          1735.662          50.             45            1
230         12.5           -7.5          1735.746          75.             44            1
231         12.5           -2.5          1736.572          0.              68            1
232         12.5           2.5          1736.708          100.            77            1
233         12.5           7.5          1737.108          25.             70            1
234         12.5          12.5          1737.386          25.             91            1
235         12.5          17.5          1737.338          50.             88            1
236         12.5          22.5          1736.372          75.             65            1
237         12.5          27.5          1735.690          75.             65            1
238         12.5          32.5          1736.042          100.            86            1
239         12.5          37.5          1736.083          100.            104           1
240         12.5          42.5          1736.273          50.             98            1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)         (deg)         (km)             (%)             observations
*****
241         12.5          47.5          1736.469          0.              86            1
242         12.5          52.5          1735.355          50.             91            1
243         12.5          57.5          1734.528          100.            123            1
244         12.5          62.5          1734.725          50.             89            1
245         12.5          67.5          1735.559          0.              83            1
246         12.5          72.5          1736.532          0.              69            1
247         12.5          77.5          1736.429          25.             62            1
248         12.5          82.5          1736.005          25.             49            1
249         12.5          87.5          1736.359          25.             51            1
250         12.5          92.5          1736.525          0.              50            1
251         12.5          97.5          1737.820          0.              49            1
252         12.5          102.5         1737.639          0.              43            1
253         7.5          -102.5         1739.384          0.              0             2
254         7.5          -97.5          1739.119          0.              8             1
255         7.5          -92.5          1738.310          0.              0             2
256         7.5          -87.5          1737.471          0.              40            1
257         7.5          -82.5          1736.627          0.              24            1
258         7.5          -77.5          1736.575          0.              11            1
259         7.5          -72.5          1736.372          25.             15            1
260         7.5          -67.5          1736.147          0.              14            1
261         7.5          -62.5          1735.916          75.             6             1
262         7.5          -57.5          1735.591          75.            13            1
263         7.5          -52.5          1736.074          100.            14            1
264         7.5          -47.5          1736.056          100.            14            1
265         7.5          -42.5          1735.876          100.             7            1
266         7.5          -37.5          1735.925          100.             9            1
267         7.5          -32.5          1736.320          100.            16            1
268         7.5          -27.5          1736.685          75.            21            1
269         7.5          -22.5          1736.570          25.            13            1
270         7.5          -17.5          1736.487          50.            18            1
271         7.5          -12.5          1736.315          100.            18            1
272         7.5           -7.5          1736.592          25.            13            1
273         7.5           -2.5          1736.784          0.              0             2
274         7.5           2.5          1737.024          0.              41            1
275         7.5           7.5          1736.871          0.              46            1
276         7.5          12.5          1737.041          0.              38            1
277         7.5          17.5          1736.585          50.              29            1
278         7.5          22.5          1735.387          100.             34            1
279         7.5          27.5          1735.863          100.             36            1
280         7.5          32.5          1736.615          100.             43            1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)            (%)            observations
*****
281          7.5            37.5            1736.750          100.            56            1
282          7.5            42.5            1736.324          25.             57            1
283          7.5            47.5            1735.675          0.              71            1
284          7.5            52.5            1735.469          0.              0             2
285          7.5            57.5            1735.890          0.              71            1
286          7.5            62.5            1736.755          0.              41            1
287          7.5            67.5            1736.691          50.             45            1
288          7.5            72.5            1737.306          25.             43            1
289          7.5            77.5            1735.936          0.              60            1
290          7.5            82.5            1735.487          25.             49            1
291          7.5            87.5            1735.086          0.              43            1
292          7.5            92.5            1735.098          0.              47            1
293          7.5            97.5            1736.520          0.              52            1
294          7.5           102.5            1737.083          0.              61            1
295          2.5           -102.5            1739.820          0.              39            1
296          2.5           -97.5            1739.185          0.              45            1
297          2.5           -92.5            1738.853          0.              0             2
298          2.5           -87.5            1738.491          0.              14            1
299          2.5           -82.5            1738.026          0.              21            1
300          2.5           -77.5            1737.879          0.              14            1
301          2.5           -72.5            1736.821          0.              0             2
302          2.5           -67.5            1736.586          0.              10            1
303          2.5           -62.5            1735.765          100.            11            1
304          2.5           -57.5            1735.869          100.            7             1
305          2.5           -52.5            1735.511          100.            8             1
306          2.5           -47.5            1736.056          100.           10            1
307          2.5           -42.5            1735.814          100.            8             1
308          2.5           -37.5            1736.605          75.             8             1
309          2.5           -32.5            1736.253          100.            7             1
310          2.5           -27.5            1736.450          100.            0             2
311          2.5           -22.5            1736.839          50.             4             1
312          2.5           -17.5            1736.382          75.            27            1
313          2.5           -12.5            1736.258          75.            33            1
314          2.5            -7.5            1736.553          50.            31            1
315          2.5            -2.5            1736.704          50.            36            1
316          2.5             2.5            1736.868          75.            52            1
317          2.5             7.5            1737.559          0.             46            1
318          2.5            12.5            1737.515          0.             37            1
319          2.5            17.5            1736.949          0.             47            1
320          2.5            22.5            1735.875          100.            41            1
*****

```

number	latitude (deg)	longitude (deg)	lunar radius (km)	maria cover (%)	number of observations	source
321	2.5	27.5	1735.519	100.	55	1
322	2.5	32.5	1736.328	25.	54	1
323	2.5	37.5	1736.545	75.	50	1
324	2.5	42.5	1736.130	25.	54	1
325	2.5	47.5	1735.304	75.	44	1
326	2.5	52.5	1735.038	75.	53	1
327	2.5	57.5	1735.409	50.	44	1
328	2.5	62.5	1736.045	25.	87	1
329	2.5	67.5	1736.341	25.	103	1
330	2.5	72.5	1736.204	0.	85	1
331	2.5	77.5	1736.420	0.	80	1
332	2.5	82.5	1734.987	0.	104	1
333	2.5	87.5	1733.536	75.	84	1
334	2.5	92.5	1734.482	25.	85	1
335	2.5	97.5	1736.469	0.	78	1
336	2.5	102.5	1738.122	0.	83	1
337	-2.5	-102.5	1739.000	0.	0	3
338	-2.5	-97.5	1739.000	0.	0	3
339	-2.5	-92.5	1739.100	0.	0	2
340	-2.5	-87.5	1739.000	0.	0	3
341	-2.5	-82.5	1739.000	0.	0	3
342	-2.5	-77.5	1737.829	0.	0	2
343	-2.5	-72.5	1737.235	0.	15	1
344	-2.5	-67.5	1736.121	25.	14	1
345	-2.5	-62.5	1736.050	25.	15	1
346	-2.5	-57.5	1735.787	100.	11	1
347	-2.5	-52.5	1735.325	100.	5	1
348	-2.5	-47.5	1735.362	100.	3	1
349	-2.5	-42.5	1735.649	100.	0	2
350	-2.5	-37.5	1735.850	100.	0	2
351	-2.5	-32.5	1736.041	100.	0	2
352	-2.5	-27.5	1736.131	100.	0	2
353	-2.5	-22.5	1736.316	75.	5	1
354	-2.5	-17.5	1736.195	100.	30	1
355	-2.5	-12.5	1736.291	75.	31	1
356	-2.5	-7.5	1736.413	50.	40	1
357	-2.5	-2.5	1737.040	25.	40	1
358	-2.5	2.5	1737.357	0.	35	1
359	-2.5	7.5	1737.758	0.	32	1
360	-2.5	12.5	1737.697	0.	29	1

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)            (%)              observations
*****
361         -2.5           17.5          1738.066          0.              38             1
362         -2.5           22.5          1736.357          25.             38             1
363         -2.5           27.5          1735.209          75.             52             1
364         -2.5           32.5          1735.913          25.             46             1
365         -2.5           37.5          1736.148          0.              69             1
366         -2.5           42.5          1735.732          75.             63             1
367         -2.5           47.5          1735.401          100.            107             1
368         -2.5           52.5          1735.216          100.            87             1
369         -2.5           57.5          1735.058          75.             96             1
370         -2.5           62.5          1735.647          0.              66             1
371         -2.5           67.5          1735.964          0.              61             1
372         -2.5           72.5          1736.621          0.              82             1
373         -2.5           77.5          1736.527          0.              87             1
374         -2.5           82.5          1734.209          0.             107             1
375         -2.5           87.5          1733.768          25.             98             1
376         -2.5           92.5          1735.407          0.              75             1
377         -2.5           97.5          1737.174          0.              87             1
378         -2.5          102.5          1739.031          0.              47             1
379         -2.5          -102.5         1739.000          0.              0              3
380         -7.5          -97.5          1737.500          0.              0              3
381         -7.5          -92.5          1738.325          0.             10             1
382         -7.5          -87.5          1738.783          0.              0              2
383         -7.5          -82.5          1740.000          0.              0              3
384         -7.5          -77.5          1738.600          0.              0              3
385         -7.5          -72.5          1737.186          0.              9              1
386         -7.5          -67.5          1736.574          0.              0              2
387         -7.5          -62.5          1736.059          25.             0              2
388         -7.5          -57.5          1735.566          0.              0              2
389         -7.5          -52.5          1735.432          50.             14             1
390         -7.5          -47.5          1735.419          100.            14             1
391         -7.5          -42.5          1735.644          100.            20             1
392         -7.5          -37.5          1735.665          100.            29             1
393         -7.5          -32.5          1735.635          100.            30             1
394         -7.5          -27.5          1735.602          75.             28             1
395         -7.5          -22.5          1735.377          50.             30             1
396         -7.5          -17.5          1735.913          75.             32             1
397         -7.5          -12.5          1735.913          100.            46             1
398         -7.5           -7.5          1736.270          100.            43             1
399         -7.5           -2.5          1737.019          0.              62             1
400         -7.5           2.5          1738.079          0.              51             1
*****
0
*****

```

number	latitude (deg)	longitude (deg)	lunar radius (km)	maria cover (%)	number of observations	source
401	-7.5	7.5	1738.472	0.	66	1
402	-7.5	12.5	1737.878	0.	77	1
403	-7.5	17.5	1738.145	0.	71	1
404	-7.5	22.5	1736.202	25.	54	1
405	-7.5	27.5	1735.033	50.	43	1
406	-7.5	32.5	1735.824	25.	60	1
407	-7.5	37.5	1735.917	0.	54	1
408	-7.5	42.5	1735.734	75.	41	1
409	-7.5	47.5	1735.588	100.	46	1
410	-7.5	52.5	1735.489	100.	77	1
411	-7.5	57.5	1735.364	0.	75	1
412	-7.5	62.5	1735.115	0.	38	1
413	-7.5	67.5	1735.992	0.	29	1
414	-7.5	72.5	1736.965	0.	21	1
415	-7.5	77.5	1736.966	0.	33	1
416	-7.5	82.5	1736.098	0.	64	1
417	-7.5	87.5	1735.016	0.	70	1
418	-7.5	92.5	1736.856	0.	78	1
419	-7.5	97.5	1738.105	0.	56	1
420	-7.5	102.5	1739.098	0.	27	1
421	-12.5	-102.5	1740.000	0.	0	2
422	-12.5	-97.5	1739.000	0.	0	3
423	-12.5	-92.5	1739.000	0.	0	3
424	-12.5	-87.5	1740.000	25.	0	3
425	-12.5	-82.5	1740.500	25.	0	3
426	-12.5	-77.5	1738.900	0.	0	3
427	-12.5	-72.5	1737.313	0.	0	2
428	-12.5	-67.5	1736.743	0.	0	2
429	-12.5	-62.5	1736.124	0.	2	1
430	-12.5	-57.5	1735.888	0.	3	1
431	-12.5	-52.5	1736.251	50.	0	2
432	-12.5	-47.5	1736.136	50.	0	2
433	-12.5	-42.5	1736.116	25.	6	1
434	-12.5	-37.5	1736.487	75.	5	1
435	-12.5	-32.5	1735.483	100.	6	1
436	-12.5	-27.5	1735.683	100.	10	1
437	-12.5	-22.5	1735.907	75.	8	1
438	-12.5	-17.5	1736.050	75.	11	1
439	-12.5	-12.5	1735.589	100.	11	1
440	-12.5	-7.5	1736.330	75.	10	1

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
441         -12.5          -2.5          1737.063          0.              6              1
442         -12.5          2.5          1736.998          0.              5              1
443         -12.5          7.5          1737.676          0.              10             1
444         -12.5          12.5         1737.658          0.              10             1
445         -12.5          17.5         1738.329          0.              12             1
446         -12.5          22.5         1736.782          0.              8              1
447         -12.5          27.5         1736.740          0.              2              1
448         -12.5          32.5         1735.237          100.            10             1
449         -12.5          37.5         1735.699          75.             5              1
450         -12.5          42.5         1735.937          0.              0              2
451         -12.5          47.5         1736.463          50.             6              1
452         -12.5          52.5         1735.873          75.             0              2
453         -12.5          57.5         1735.616          0.              0              2
454         -12.5          62.5         1735.867          0.              0              2
455         -12.5          67.5         1736.020          0.              1              1
456         -12.5          72.5         1737.060          0.              0              2
457         -12.5          77.5         1737.208          0.              5              1
458         -12.5          82.5         1737.999          0.              19             1
459         -12.5          87.5         1737.983          0.              17             1
460         -12.5          92.5         1737.402          0.              19             1
461         -12.5          97.5         1738.237          0.              0              2
462         -12.5          102.5        1737.289          0.              23             1
463         -17.5          -102.5        1738.500          0.              0              3
464         -17.5          -97.5         1735.000          50.             0              3
465         -17.5          -92.5         1735.000          50.             0              3
466         -17.5          -87.5         1738.500          0.              0              3
467         -17.5          -82.5         1740.500          25.             0              3
468         -17.5          -77.5         1738.829          0.              0              2
469         -17.5          -72.5         1737.292          0.              0              2
470         -17.5          -67.5         1737.003          0.              0              2
471         -17.5          -62.5         1736.886          0.              4              1
472         -17.5          -57.5         1736.806          0.              4              1
473         -17.5          -52.5         1737.030          0.              6              1
474         -17.5          -47.5         1736.673          0.              0              2
475         -17.5          -42.5         1736.300          0.              0              2
476         -17.5          -37.5         1735.560          50.             3              1
477         -17.5          -32.5         1735.993          50.             3              1
478         -17.5          -27.5         1735.515          50.             0              2
479         -17.5          -22.5         1735.577          75.             5              1
480         -17.5          -17.5         1735.322          75.             0              2
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)         (deg)         (km)             (%)             observations
*****
481         -17.5         -12.5         1735.843         100.           5             1
482         -17.5         -7.5         1736.043         50.            0             2
483         -17.5         -2.5         1737.012         0.             0             2
484         -17.5         2.5         1736.938         0.            12             1
485         -17.5         7.5         1737.385         0.            17             1
486         -17.5         12.5        1737.770         0.            13             1
487         -17.5         17.5        1738.415         0.             0             2
488         -17.5         22.5        1737.490         0.             0             2
489         -17.5         27.5        1737.268         0.             0             1
490         -17.5         32.5        1735.848         100.           10             1
491         -17.5         37.5        1736.051         75.            3             1
492         -17.5         42.5        1735.362         25.            5             1
493         -17.5         47.5        1736.544         0.             5             1
494         -17.5         52.5        1736.723         100.           3             1
495         -17.5         57.5        1735.807         25.            2             1
496         -17.5         62.5        1736.217         0.             9             1
497         -17.5         67.5        1737.037         0.             0             2
498         -17.5         72.5        1737.890         0.             0             2
499         -17.5         77.5        1738.581         0.             0             2
500         -17.5         82.5        1739.092         0.             2             1
501         -17.5         87.5        1738.706         0.             0             2
502         -17.5         92.5        1738.357         0.             1             1
503         -17.5         97.5        1738.256         0.             0             2
504         -17.5        102.5       1738.179         0.             0             2
505         -22.5        -102.5      1739.000         0.             0             3
506         -22.5        -97.5      1733.000         25.            0             3
507         -22.5        -92.5      1734.000         25.            0             3
508         -22.5        -87.5      1739.000         0.             0             3
509         -22.5        -82.5      1740.500         0.             0             3
510         -22.5        -77.5      1738.900         0.             0             3
511         -22.5        -72.5      1737.216         0.             0             2
512         -22.5        -67.5      1737.446         0.             0             2
513         -22.5        -62.5      1737.275         0.             0             2
514         -22.5        -57.5      1737.119         0.             3             1
515         -22.5        -52.5      1737.284         0.             2             1
516         -22.5        -47.5      1736.983         0.             0             1
517         -22.5        -42.5      1734.979         100.           8             2
518         -22.5        -37.5      1736.089         100.          11             1
519         -22.5        -32.5      1736.133         50.            5             1
520         -22.5        -27.5      1736.803         50.            3             1
*****

```



```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
521         -22.5          -22.5          1736.962           75.             2             1
522         -22.5          -17.5          1735.745           100.            6             1
523         -22.5          -12.5          1734.989           100.            4             1
524         -22.5          -7.5           1735.848           75.             0             2
525         -22.5          -2.5           1737.036           0.              12            1
526         -22.5          2.5            1737.595           0.              0             2
527         -22.5          7.5            1738.509           0.              0             2
528         -22.5          12.5           1738.150           0.              11            1
529         -22.5          17.5           1738.027           0.              11            1
530         -22.5          22.5           1739.039           0.              8             1
531         -22.5          27.5           1737.607           0.              0             2
532         -22.5          32.5           1737.363           0.              0             2
533         -22.5          37.5           1736.957           0.              7             1
534         -22.5          42.5           1737.516           0.              0             2
535         -22.5          47.5           1736.612           25.             0             2
536         -22.5          52.5           1736.653           50.             3             1
537         -22.5          57.5           1736.910           0.              5             1
538         -22.5          62.5           1737.040           0.              0             2
539         -22.5          67.5           1737.620           0.              0             2
540         -22.5          72.5           1738.413           0.              0             2
541         -22.5          77.5           1738.600           0.              0             3
542         -22.5          82.5           1739.157           0.              2             1
543         -22.5          87.5           1738.206           0.              2             1
544         -22.5          92.5           1738.046           0.              0             2
545         -22.5          97.5           1737.504           0.              2             1
546         -22.5          102.5          1737.386           0.              0             2
547         -27.5          -102.5          1739.000           0.              0             3
548         -27.5          -97.5           1736.500           0.              0             3
549         -27.5          -92.5           1736.500           0.              0             3
550         -27.5          -87.5           1739.000           0.              0             3
551         -27.5          -82.5           1740.500           0.              0             3
552         -27.5          -77.5           1738.900           0.              0             3
553         -27.5          -72.5           1737.045           0.              0             3
554         -27.5          -67.5           1737.588           0.              3             2
555         -27.5          -62.5           1737.150           0.              0             1
556         -27.5          -57.5           1737.331           0.              0             2
557         -27.5          -52.5           1737.230           0.              0             2
558         -27.5          -47.5           1736.950           0.              0             2
559         -27.5          -42.5           1736.868           50.             0             2
560         -27.5          -37.5           1735.219           75.             6             1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
561         -27.5          -32.5          1737.573          50.              3              1
562         -27.5          -27.5          1735.428          50.              5              1
563         -27.5          -22.5          1735.837          100.             4              1
564         -27.5          -17.5          1735.831          75.              4              1
565         -27.5          -12.5          1735.902          75.              9              1
566         -27.5          -7.5           1735.966          50.              0              2
567         -27.5          -2.5           1736.255          0.               0              2
568         -27.5          2.5           1737.219          0.               7              1
569         -27.5          7.5           1737.484          0.               7              1
570         -27.5          12.5          1738.850          0.               9              1
571         -27.5          17.5          1739.142          0.               13             1
572         -27.5          22.5          1738.775          0.               6              1
573         -27.5          27.5          1737.831          0.              10              1
574         -27.5          32.5          1738.171          0.               0              2
575         -27.5          37.5          1738.119          0.               0              2
576         -27.5          42.5          1738.062          0.               0              2
577         -27.5          47.5          1737.866          0.               0              2
578         -27.5          52.5          1737.557          0.               0              2
579         -27.5          57.5          1737.335          0.               0              2
580         -27.5          62.5          1737.421          0.               0              2
581         -27.5          67.5          1737.851          0.               0              2
582         -27.5          72.5          1738.633          0.               0              2
583         -27.5          77.5          1738.469          0.               1              1
584         -27.5          82.5          1738.546          0.               0              2
585         -27.5          87.5          1737.908          0.               0              2
586         -27.5          92.5          1736.975          0.               0              1
587         -27.5          97.5          1736.628          0.               1              1
588         -27.5          102.5         1736.445          25.              0              2
589         -32.5          -102.5        1739.762          0.               0              2
590         -32.5          -97.5         1739.500          0.               0              2
591         -32.5          -92.5         1740.000          0.               0              3
592         -32.5          -87.5         1740.500          0.               0              3
593         -32.5          -82.5         1741.000          0.               0              3
594         -32.5          -77.5         1738.900          0.               0              3
595         -32.5          -72.5         1736.849          0.               0              2
596         -32.5          -67.5         1737.021          0.               0              2
597         -32.5          -62.5         1737.017          0.               0              2
598         -32.5          -57.5         1736.811          0.               2              1
599         -32.5          -52.5         1736.866          0.               0              2
600         -32.5          -47.5         1736.731          0.               0              2
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)         (deg)         (km)             (%)             observations
*****
601         -32.5         -42.5         1736.667         50.             3             1
602         -32.5         -37.5         1736.693         0.              6             1
603         -32.5         -32.5         1737.506         50.             0             2
604         -32.5         -27.5         1737.931         75.             0             2
605         -32.5         -22.5         1738.011         50.             0             2
606         -32.5         -17.5         1737.636         25.             0             2
607         -32.5         -12.5         1737.326         0.              4             1
608         -32.5         -7.5          1736.384         0.              0             2
609         -32.5         -2.5          1736.200         0.              0             2
610         -32.5         2.5           1736.410         0.              6             2
611         -32.5         7.5           1737.355         0.              9             1
612         -32.5         12.5          1737.768         0.              5             1
613         -32.5         17.5          1738.543         0.              8             1
614         -32.5         22.5          1738.364         0.              12            1
615         -32.5         27.5          1738.906         0.              6             1
616         -32.5         32.5          1738.447         0.              0             2
617         -32.5         37.5          1738.388         0.              8             1
618         -32.5         42.5          1738.034         0.              0             2
619         -32.5         47.5          1737.754         0.              0             2
620         -32.5         52.5          1737.225         0.              3             1
621         -32.5         57.5          1737.335         0.              0             2
622         -32.5         62.5          1736.890         0.              0             2
623         -32.5         67.5          1737.733         0.              0             2
624         -32.5         72.5          1738.123         0.              0             2
625         -32.5         77.5          1738.309         0.              0             2
626         -32.5         82.5          1737.704         0.              0             2
627         -32.5         87.5          1737.379         25.             2             1
628         -32.5         92.5          1736.347         50.             0             2
629         -32.5         97.5          1735.555         25.             0             1
630         -32.5         102.5         1735.929         0.              0             2
631         -37.5         -102.5        1741.500         0.              0             3
632         -37.5         -97.5         1739.500         0.              0             3
633         -37.5         -92.5         1740.057         0.              0             3
634         -37.5         -87.5         1740.500         0.              0             3
635         -37.5         -82.5         1741.000         0.              0             3
636         -37.5         -77.5         1738.500         0.              0             3
637         -37.5         -72.5         1736.831         0.              0             3
638         -37.5         -67.5         1736.629         0.              1             1
639         -37.5         -62.5         1736.777         0.              1             1
640         -37.5         -57.5         1737.051         0.              3             1
*****

```

```

*****
number      latitude      longitude      lunar radius      maria cover      number of      source
            (deg)          (deg)          (km)              (%)              observations
*****
641         -37.5          -52.5          1736.024           0.              3              1
642         -37.5          -47.5          1736.596           0.              0              2
643         -37.5          -42.5          1736.826           50.             0              2
644         -37.5          -37.5          1737.350           0.              0              2
645         -37.5          -32.5          1738.044           0.              0              2
646         -37.5          -27.5          1738.391           25.             5              1
647         -37.5          -22.5          1738.795           0.              0              2
648         -37.5          -17.5          1738.432           0.              0              2
649         -37.5          -12.5          1737.618           0.              7              1
650         -37.5          -7.5           1736.738           0.              8              1
651         -37.5          -2.5           1736.262           0.              7              1
652         -37.5           2.5           1735.812           0.              5              1
653         -37.5           7.5           1737.137           0.              4              1
654         -37.5          12.5           1737.588           0.             13              1
655         -37.5          17.5           1738.159           0.             10              1
656         -37.5          22.5           1737.622           0.             11              1
657         -37.5          27.5           1738.089           0.              8              1
658         -37.5          32.5           1738.176           0.              7              1
659         -37.5          37.5           1737.304           0.              4              1
660         -37.5          42.5           1737.879           0.              3              1
661         -37.5          47.5           1736.577           0.              2              1
662         -37.5          52.5           1736.975           0.              0              2
663         -37.5          57.5           1736.991           0.              0              2
664         -37.5          62.5           1737.191           0.              0              2
665         -37.5          67.5           1737.542           0.              0              2
666         -37.5          72.5           1738.009           0.              0              2
667         -37.5          77.5           1738.007           0.              0              2
668         -37.5          82.5           1737.237           0.              0              2
669         -37.5          87.5           1736.868           25.             2              1
670         -37.5          92.5           1736.313           75.             0              2
671         -37.5          97.5           1735.780           25.             2              1
672         -37.5          102.5          1736.738           25.             0              2
*****

```

APPENDIX C

THE POTENTIAL OF A THIN DISK

The potential of an infinitely thin disk has been developed by Thurber [1980, personal communication]. We his procedure to derive a formula with which the displacements and stresses near a cooling, axisymmetric body can be computed.

The integral representation for the potential of an infinitely-thin disk of radius a , constant temperature change ΔT , and centered at $(0,2)$ is given by:

$$\phi_1(r,z) = \beta \int_0^a \int_0^{2\pi} \frac{1}{R_1} r' d\theta' dr' \quad (C1)$$

where β is defined in equation (3), R_1 is as defined in equations (7) of Chapter 6 (to compute ϕ_2 , insert R_2 from equation (8) in place of R_1), and θ is the angle swept out in the plane of the disk. If we assume then the observation point is at $\theta=0^\circ$, equation (C1) can be rearranged to the form

$$\phi_1(r,z) = \beta \int_0^a \int_0^{2\pi} \frac{r'}{(r'^2 + r^2 + (z-z')^2)^{1/2}} \frac{1}{(1-2x)^{1/2}} d\theta' dr' \quad (C2)$$

where,

$$x = \frac{rr' \cos\theta'}{r'^2 + r^2 + (z-z')^2} \quad (C3)$$

We then proceed with a binomial expansion of the second term in the integrand:

$$(1-2x)^{-1/2} = 1 + \frac{1}{2}(2x) + \frac{\frac{1}{2} \cdot \frac{3}{2}}{2!} (2x)^2 + \frac{\frac{1}{2} \cdot \frac{3}{2} \cdot \frac{5}{2}}{3!} (2x)^3 + \dots \quad (C4)$$

which condenses to

$$(1-2x)^{-1/2} = \sum_{n=0}^{\infty} \frac{(2n-1)(2n-3)\dots(1)}{n!} x^n \quad (C5)$$

Inserting (C3) and (C5) into (C2), removing from the integrand terms not involved in integrations, and separating the remaining terms into integrations over θ' and over r' results in:

$$\begin{aligned} \phi(r,z) = & \beta \sum_{n=0}^{\infty} \frac{(2n-1)(2n-3)\dots(1)}{n!} r^n \int_0^{2\pi} \cos n\theta' d\theta' \\ & \cdot \int_0^a \frac{r'^{n+1}}{(r'^2 + r^2 + (z-z')^2)^{n+1/2}} dr' \end{aligned} \quad (C6)$$

Integrating the cosine term in (C6) produces:

$$\begin{aligned} & \int_0^{2\pi} \cos n\theta' d\theta' = 0, \quad \text{for odd } n \\ & 2\pi \frac{(n-1)(n-3)\dots(1)}{(n)(n-2)\dots(2)}, \quad \text{for even } n \end{aligned} \quad (C7)$$

Thus we substitute $2k$ for n throughout (C6) and sum over all integral k :

$$\begin{aligned} \phi_1(r,z) = & 2\pi\beta \sum_{k=0}^{\infty} \frac{(4k-1)(4k-3)\dots(1)}{(2k)!} \cdot \frac{(2k-1)(2k-3)\dots(1)}{(2k)(2k-2)\dots(2)} \cdot r^{2k} \\ & \cdot \int_0^a \frac{r'^{2k+1}}{(r'^2 + r^2 + (z-z')^2)^{2k+1/2}} dr' \end{aligned} \quad (C8)$$

and combining the two factorial terms at the front of the summation in (C8) we get:

$$\phi_1(r,z) = 2\pi\beta \sum_{k=0}^{\infty} \frac{(4k)!}{2^{4k}(k!)^2(2k)!} r^{2k} \int_0^a \frac{r'^{2k+1}}{(r'^2+r^2+(z-z')^2)^{2k+1/2}} dr' \quad (C9)$$

The remaining integration can be solved by parts for any given k . For example, the $k=0$ term of (C9) reduces to:

$$2\pi\beta[(a^2+r^2+(z-z')^2)^{1/2} - (r^2+(z-z')^2)^{1/2}] \quad (C10)$$

which, for $r=0$, is simply the potential of a thin disk at a point on the axis [Moulton, 1914, p. 113].

The number of terms necessary for convergence of (C9) is a function of the location of the observation point $(r, z-z')$ or $(r, z+z')$ relative to the disk. More terms are required when the observation point is both near the plane of the disk and close to the disk's edge. Less terms are necessary close to the axis of the disk or at observation points far from the disk (i.e. $r^2+(z-z')^2 > 2a$).

We numerically integrate (C9) over z' to derive the potential of a cylinder of finite thickness. This potential can then be inserted into equation (4), Chapter 6 to calculate the thermoelastic displacement field produced by a uniformly cooling cylinder.

BIOGRAPHICAL NOTE

The author was born to Janet and Richard Bratt in the Brooklyn (NY) Naval Hospital on February 10, 1957. By the time Sputnik had completed its first orbit of the Earth later the same year, the infant and his family had moved to the Great White North: Syracuse, NY. There, he grew up a child of the space age. He spent the few nights during which it neither snowed nor rained camped in the backyard searching for falling stars and E.T.

In the 10th grade, the then rotund youth found his athletic passion in placing one foot in front of the other at an increasingly rapid rate. Later that year, a slimmer youth found his scholarly passion in the Silurian and Devonian near the Westhill High School during a course in Earth sciences. Convinced that he could become the first geoscientist on Mars, he began a study of geophysics at Penn State. His education in geophysics was continued at M.I.T. where space was drawn nearer to him by the study of the planets.

During the past year, he has completed a thesis, was a step away from a job as an astronaut, and married the woman who herself wrote a thesis in which he was a subject... it has been an eventful year. He also has published or submitted the following papers:

Bratt, S.R., and S.C. Solomon, Compressional and shear wave structure of the East Pacific Rise at 11°20'N: Constraints from three-component ocean bottom seismometer data, J. Geophys. Res., 89, 6095-6110, 1984.

Bratt, S.R., and G.M. Purdy, Structure and variability of oceanic crust on the flanks of the East Pacific Rise between 11° and 13°N, J. Geophys. Res., 89, 6111-6125, 1984.

Bratt, S.R., S.C. Solomon, J.W. Head, and C.H. Thurber, The deep structure of lunar basins: Implications for basin formation and modification, J. Geophys. Res., in press 1984.

Bratt, S.R., E.A. Bergman, and S.C. Solomon, Thermoelastic stress: How important as a cause of earthquakes in young oceanic lithosphere?, J. Geophys. Res., submitted, 1984.