INFORMATION TO USERS

This manuscript has been reproduced from the microfilm master. UMI films the text directly from the original or copy submitted. Thus, some thesis and dissertation copies are in typewriter face, while others may be from any type of computer printer.

The quality of this reproduction is dependent upon the quality of the copy submitted. Broken or indistinct print, colored or poor quality illustrations and photographs, print bleedthrough, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send UMI a complete manuscript and there are missing pages, these will be noted. Also, if unauthorized copyright material had to be removed, a note will indicate the deletion.

Oversize materials (e.g., maps, drawings, charts) are reproduced by sectioning the original, beginning at the upper left-hand corner and continuing from left to right in equal sections with small overlaps. Each original is also photographed in one exposure and is included in reduced form at the back of the book.

Photographs included in the original manuscript have been reproduced xerographically in this copy. Higher quality 6" x 9" black and white photographic prints are available for any photographs or illustrations appearing in this copy for an additional charge. Contact UMI directly to order.
Computer simulation of stratigraphy

Bowman, Scott Andrew, Ph.D.
Rice University, 1994

Copyright ©1994 by Bowman, Scott Andrew. All rights reserved.
RICE UNIVERSITY
COMPUTER SIMULATION OF STRATIGRAPHY

by

SCOTT ANDREW BOWMAN

A THESIS SUBMITTED
IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE
DOCTOR OF PHILOSOPHY

APPROVED, THESIS COMMITTEE

[Signatures]

PETER R. VAIL, W. Maurice Ewing Professor, Chairman
Geology and Geophysics

ALBERT W. BALLY, Harry C. others Wiess Professor
Geology and Geophysics

ANDRE W. DROXLER, Associate Professor
Geology and Geophysics

DALE S. SAWYER, Associate Professor
Geology and Geophysics

WILLIAM W. SYMES, Professor
Mathematical Sciences

Houston, Texas
April, 1994
COPYRIGHT

Scott Andrew Bowman

April, 1994
ABSTRACT

Computer Simulation of Stratigraphy

by

Scott Andrew Bowman

Simulation of stratigraphy coupled with sequence stratigraphic and backstripping analysis quantitatively defines the timing and magnitude of geologic events, including the history of sediment supply, tectonism and eustasy. This provides a quantitative basis for interpreting the mechanisms causing these variations. This computer simulation comprises algorithms that model subsidence and uplifts, eustasy, flexural response of sediment and water loads, compaction, traction- and suspension-load deposition, gravity-flow sedimentation, carbonate production and redistribution, and erosion.

Backstripping analysis can provide a geohistory, burial history, sediment accumulation history, porosity history, and a first approximation of the tectonic subsidence or uplift history. A backstripping analysis of a stratigraphic section produced by the two-dimensional simulator demonstrates the error due to overcompensating for the flexural response to sediment loading with a calculation that assumes local isostasy. These errors reinforce the necessity to use a two- or three-dimensional simulation or backstripping technique to accurately define the eustatic and tectonic history of a region.

Simulation results of the Last Chance Canyon study show that documented stratal patterns are a product of the interaction of a dynamic depositional system, with constant parameters, fed by alternating silicilastic sand and carbonate production, a constant subsidence rate of 0.4 cm/ky, and a eustatic sea-level history that contains “third-order” and higher periodicity cycles.

A hierarchy of stratigraphic packaging is presented that include continental encroachment megasequences, transgressive-regressive facies-superequences, complete and incomplete sequences, component groups and components. Sediment supply,
tectonism, and eustatic fluctuations produce these packages by changing the accommodation space with characteristic rates and patterns. The geometry of the substrate and bathymetric changes strongly influence the geometry of stratal surfaces and distribution of lithofacies. The response of these variables is simulated independently to identify their unique stratal signatures. Examples from different settings (passive margin, mixed carbonate-silicilastic sediment supplies, carbonate ramp, and steep carbonate platform margin, and others) demonstrate how siliciclastic and carbonate depositional systems interact with the bathymetric conditions produced by these variables. Simulation results show that the relative change of sea level is the sum of total subsidence (tectonic subsidence, flexure loading and compaction) and eustasy.
Acknowledgment

I would like to thank Dr. Peter R. Vail for sharing his time and overwhelming insight into the intricacies of the stratigraphic record. I would also like to thank the members of my committee, fellow graduate students, and faculty at Rice for their stimulating discussions and increasing my awareness many critical observations concerning the nature of strata, seismic data, and tectonic processes.

I also would like to thank the people at the following universities, government organizations and corporations for their considerable technical suggestions and support:

Aarhus University, Denmark
Apple Computer
Arco
Bourgogne University
BMR, Australia
CNRS, Dijon, France
Dupont E&P, France
Elf Aquitaine, France
Lamar University
Marathon
Middlebury College
Royal Dutch Shell
SNEAP, France
Statoil
Total
Tubingen University
Union College
UCLA

Amoco
Aramco
Baylor University
British Petroleum
Chevron
Conoco
Edinburg University
Exxon
Karl University
McGill University
Mobil
Saga Petroleum
Stanford University
Texaco
Toyo Engineering, Japan
USGS
University of Bern
University of California, Riverside
University of Missouri, Rolla          University of Nebraska
University of Texas, Arlington        Unocal
Vassar University                      Winona State University
Virginia Polytechnic Institute

I am indebted to Conoco for providing data and printing facilities. I am indebted to Rice University for providing a scholarship and the environment to pursue this research.

Most of all, this project would not have been possible without Maria Carmenza. She is my backbone, my heart and my life.
# Table of Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>i</td>
</tr>
<tr>
<td>Acknowledgment</td>
<td>iii</td>
</tr>
<tr>
<td>Chapter 1. Introduction</td>
<td>1</td>
</tr>
<tr>
<td>Chapter 2. Computer Simulation of Stratigraphy</td>
<td>5</td>
</tr>
<tr>
<td>Chapter 3. Response of Stratal Patterns to Sediment Supply, Tectonic, and Eustatic Variations</td>
<td>58</td>
</tr>
<tr>
<td>Chapter 4. Subsidence Analysis of Stratigraphic Sections -- Comparison of One- and Two-Dimensional Data</td>
<td>132</td>
</tr>
<tr>
<td>Chapter 5. Early Guadalupian (middle Permian) Sequence Stratigraphy of the Last Chance Canyon, Delaware Basin, New Mexico: Quantified with a Computer Simulation</td>
<td>154</td>
</tr>
<tr>
<td>Glossary of Terms</td>
<td>207</td>
</tr>
<tr>
<td>References</td>
<td>211</td>
</tr>
</tbody>
</table>
Chapter 1.

Introduction

The purpose of this research is to improve our understanding of the controls on stratal pattern geometries and their relation to depositional environments and lithofacies. This goal was accomplished by developing a set of tools to apply to stratigraphic databases to extract sedimentation and tectonic histories, as well as reconstruct a cross-sectional view of paleogeography. This information is integrated with a computer simulation (PHIL: Process and History Integrated Layers) that recreates a stratigraphic profile described by the depositional system and historical data.

Most models are "mentally tested" ideas that are presented as idealized cartoons. We have developed technology to rigorously build and test ideas with tested quantitative algorithms that closely approximate our understanding of climatic, sedimentologic and tectonic processes. These algorithms can be applied to any variation in bathymetry produced by antecedent bathymetry, and history of tectonism, eustasy, and sediment supply. Eustatic, sediment supply, and tectonic histories are rarely measured and are often simply drafted to represent the interpreters concept.

Primary seismic reflections follow geologic time lines (Vail et al., 1977) or stratal surfaces correlated in well logs and outcrop sections. These surfaces record the present geometry or remnant geometries of depositional interfaces at the time of deposition. These geometries can be used to constrain the tectonic, eustatic, sedimentation, and erosion history. Careful measurement of stratal geometries coupled with simulation studies document the character of the events recorded in the stratigraphic record. These quantitative results are critical for inferring the mechanisms that formed the stratal geometry as well as providing insights into natural processes.

PHIL provides the technology to document the history of eustasy, tectonism, and sediment supply by adjusting these values to match progradation, sediment thickness and
paleobathymetry through time as recorded in the stratal geometry and lithologic record. The simulator can be used to model sections from different basins during the same time period to determine if the same eustatic history works in all areas. The Haq et al., (1988) curve was digitized and applied to simulation studies of passive margins in the Neogene to model the response of these long-term and short-term variations given the subsidence and sediment supply history of various settings. We continually update this new curve to reflect stratigraphic geometries from sequence stratigraphic studies from studying Neogene, Cretaceous, and Jurassic. To date, the curve was modified in the Neogene to reduce the Early Pliocene highstand and include the high-frequency variations of the Pleistocene.

Our siliciclastic depositional system shows the importance of transportation by traction and suspension mechanisms. The timing and nature of gravity-flow sediments within a eustatic cycle is understood in terms of basin bathymetry and changes in relative changes of sea-level.

The carbonate depositional systems and lithofacies are modeled by both sediment production and redistribution process. Carbonate production is a function of water depth as well as the depositional environment. A refinement of the carbonate production model includes highly-productive carbonate growth proximal to open-marine basins, non-restricted production, production of fine-grained sediment on the platform that is deposited around the platform by suspension processes, and pelagic sediment production and calcite dissolution at depth. The highly-productive growth produces isolated buildups on the rims of carbonate platforms.

The interaction of carbonate and siliciclastic deposition shows the importance of siliciclastic sediment in reducing carbonate productivity. Models demonstrate the interplay of bathymetry, eustasy, and gravity-flow sedimentation during carbonate and siliciclastic sedimentation.
The addition of surface beveling, channel incision, sediment redistribution by marine currents, and shoreface erosion to stratigraphic modeling shows the role of base level in controlling erosive surfaces within system tracts. The timing of valley incision is restricted to falls in base-level, whereas ravinement surfaces form during periods of rapid sea-level rise during the formation of the lowstand prograding complex, transgressive, and early highstand system tract. Ravinement surfaces are minimum during the late highstand and maximum during the transgressive system tract.

This work has contributed to the following changes in the sequence stratigraphic model.

(1) total subsidence rather than tectonic subsidence controls the history of relative changes of sea-level;
(2) gravity-flow sediments are controlled by relative changes of sea level and the physiography of the basin;
(3) equilibrium point is not important in controlling the bay line.
(4) bayline is controlled by the interaction of fluvial and marine processes;
(5) fluvial gradient is dynamic and decreases with progradation
(6) carbonate productivity is strongly dependent on bathymetry and basin margin geometry. Total production varies with different phases of the relative changes of sea level curve and bathymetric configurations;
(7) ravinement develops during the lowstand prograding complex, transgressive and early highstand system tracts because rapid increases in accommodation are necessary to position storm-weather wavebase on coastal plain sediments;
(8) recognizing highstand, lowstand and forced regression prograding patterns from the aggradational and progradational stratal patterns;
(9) criteria for choosing the peak of the maximum flooding surface;
(10) character of sequences and system tracts within long-term transgressive and regressive cycles;
(11) role of tectonics in enhancing or subduing sequence boundaries;
(12) define the relative magnitudes of third-order eustatic variation to fourth and fifth order cyclicity;
(13) introduced the concept of complete and incomplete sequences;
(14) demonstrated how variations in sediment supply, eustasy, and tectonism influence the timing of sequence and system tract boundaries;
(15) demonstrated how changes in precipitation/runoff has a larger impact on the fluvial system than changes in gradient due to relative changes of sea level.

Our computer simulation is valuable as a platform to test and illustrate the nature of stratigraphic processes. Its comprehensive nature and controlled processes make it an important teaching tool for both the expert and inexperienced student of stratigraphy.
Chapter 2.

Computer Simulation of Stratigraphy

ABSTRACT

Computer simulation of stratigraphy allows the geologist to generate detailed stratigraphic sections. These sections approximate the stratal geometries and lithofacies distribution as determined from petroleum exploration data (seismic sections, well logs, lithologic and paleontologic control). This computer simulation comprises algorithms that model tectonism, eustasy, flexural response of sediment and water loads, compaction, traction and suspension-load deposition, gravity-flow sedimentation, carbonate production and redistribution, and erosion. This simulation is sufficiently flexible to allow one to test most stratigraphic hypotheses by adjusting variables controlling tectonism, eustasy, and sediment supply.

Interactive modeling of stratigraphic sections quantitatively defines the variables controlling stratal patterns and lithofacies distribution. It includes procedures that provide realistic values for the model: stratigraphic analysis, calculation of tectonic subsidence and sedimentation supply rates, carbonate production rates, definition of stability conditions, erosion rates, and development of a eustatic sea-level curve. The user adjusts the inputs to bring the simulation successively closer to the observed stratigraphy.

So far we have applied the model to calibrate the depositional model, and eustatic history in twenty varied geologic settings. We demonstrate the model response to the Neogene sea level history with results from the Neogene of Baltimore Canyon, offshore New Jersey. The best simulation results come from integrating varying sediment-supply histories with tectonic events and eustatic sea-level curves of varying rate (frequency).
INTRODUCTION

Six principal factors control the geometry of sedimentary sequences: 1) rates and distribution of tectonism along the margin, 2) rate of eustatic sea-level fluctuation, 3) rates of sediment flux into the available space, 4) rate and location of erosion, 5) climate, and 6) antecedent topography and bathymetry. However, the relative importance of each factor has been a topic of much debate. Many of the differences are based on theoretical limitations of rate, duration, distribution in space, and magnitude of their effects. This paper is an outline of the core algorithms of a stratigraphic simulator. The model response is demonstrated in a Neogene section from the Baltimore Canyon, offshore New Jersey (Plate 1, Figs. 1a-11). The methodology for quantitatively analyzing a stratigraphic cross-section with the aid of computer simulation is also outlined.

Computer simulation of stratigraphy permits the geologist to generate detailed geologic cross-sections that approximate the stratal geometries and lithofacies distribution as determined from seismic sections, well logs, lithologic and paleontologic control. PHIL (Process- and History- Integrated Layers) comprises sedimentation and deformation algorithms that model:

* tectonic history (subsidence and uplift)
* eustatic sea-level history
* crustal response due to flexure loading of sediment and water loads
* compaction of sediment
* traction fill of siliciclastic sediment in fluvial and coastal settings
* dispersion of suspension load in marine settings
* gravity-flow sedimentation
* production and redistribution of carbonate sediment
* erosion

This simulation is sufficiently flexible to allow one to test most stratigraphic hypotheses by adjusting the many geologically important variables. Comparisons with other
published simulators show that PHIL is a comprehensive geologic simulation program (Table 1).

Modeling a stratigraphic cross-section integrates all that is known about a stratigraphic interval and reproduces it with a computer simulation that incorporates a depositional model with tectonic, climatic, eustatic, and sediment-supply histories. Modeling stratigraphic cross-sections allows the geologist to quantitatively define the variables controlling stratal patterns and lithofacies distribution within the strata. Modeling involves six stages: interpretation of stratigraphic data, integration of data, digitizing data, calibration of model parameters, simulation, and comparison (Fig. 2). Stratigraphic analysis includes sedimentological analysis, transgressive and regressive facies-cycle wedge analysis (White, 1980), and sequence stratigraphic analysis and must be assembled in a stratigraphic cross-section in depth for comparison. Calibration of parameters comprises backstripping to approximate the tectonic history and restoration of the initial bathymetric profile, measurement of gradients on clinoforms and stream profiles, selection of rates for carbonate production, erosion and calculating siliciclastic sediment influx rates to locally appropriate values for the model. This process continues until a satisfactory match is achieved.

Simulations of basin filling help to quantitatively model the timing, magnitude, and frequency of depositional, eustatic, and tectonic processes. Simulations of sediment transport and depositional processes are helpful in testing hypotheses related to the development of stratal patterns and the distribution of depositional systems and their associated lithofacies tracts. Interactive modeling increases the number of potential models, yet produces a detailed simulation of the stratigraphic history, providing criteria for characterizing the response of each process. The simulation results allow the geologist to improve both the geologic interpretation and the database. A simulation can also be used to test the viability of an exploration concept. In addition, the model results
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>This Model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lawrence, Doyle, and Algarr, 1990</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>P</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kendall et al 1991</td>
<td></td>
<td></td>
<td>T</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tetzlaff and Harbaugh 1989</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Biter and Harbaugh 1987</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jevry 1998</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lessenger 1993</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Thorpe 1985</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Matthews and Frolich, 1991</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bischoff 1988</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Demico and Spencer, 1989</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Read et al 1990</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kaufman et al 1990</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Goldhammer and Dunn 1991</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pitman and Golovchenko, 1984 - 1991</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Beamont and Quintan 1981, 1984</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cloetingh 1990</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reynolds, Cookley, and Stocker, 1991</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jordan and Flemings, 1991</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes:
- P - function of pore pressure
- LS - function of slope and lithology
- D - function of water depth
- I - Intervals
- 1D - vertical shear
- 2D - two dimensional
- 3D - Three dimensional
- E - Erosion
- T - Substrate
<table>
<thead>
<tr>
<th>Erosion</th>
<th>Techniques</th>
<th>Eustasy</th>
<th>Displays</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediment</td>
<td>Erosional/Depositional</td>
<td>Sea Level, Climate</td>
<td>Sedimentology, Lithology</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Time Steps</th>
<th>Depositional Systems</th>
<th>Chronostratigraphy</th>
<th>Relative Sea Level</th>
<th>Sediment Traps</th>
<th>Stratigraphy</th>
<th>Paleontology</th>
<th>Paleohydrology</th>
<th>Sediment Model</th>
<th>Revision</th>
<th>Morphing</th>
<th>Computer</th>
</tr>
</thead>
<tbody>
<tr>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
<td>1,2,3</td>
</tr>
</tbody>
</table>

**Conclusion**

- Ocean to Outcrop Scale Models
- Realistic Modeling of Geologic Cross-sections
- U.W. Cannot Determine Factors
- U.W. Can simulate modern transport processes
- Can simulate Siliclastic Deposition
- Relative Sea Level Change
- Cause of Outcrop
- Can simulate Siliclastic Deposition
- You may have bypassing in transgressive
- Sequences built with 4th and 5th Order Cycles
- Carbonate banks can be modeled
- Carbonate banks can be modeled as tidal bars
- Diffusion works for carbonates
- Hierarchy of stratigraphic forcing in carbonates
- M. Long-term changes in sea level create packages
- M. Changes in load distribution create packages
- M. Changes in stress state create packages
- U.W. Features and composition are important
- M. Features important, diffusion works okay

---

- T. "Triangle method"  
- M. Mainframe  
- U.W. - Unix workstation  
- D. MS DOS  
- V. Viscelastic  
- Mac - Macintosh  

**Bowman** 3.10.01.07
Figure 2. Stratigraphic cross-section modeling procedure.
help determine to the timing of hydrocarbon generation and migration when integrated with maturation and fluid-flow studies.

The simulator is capable of reproducing the stratal patterns observed in seismic data and geologic cross-sections from basin margins. Stratal termination patterns reproduced by coastal sedimentation include onlap, downlap, toplap, erosional truncation and apparent truncation (Vail et al., 1977a, 1991). Given an adequate sediment supply, the time lines reproduce forestepping stratal-patterns during decreasing-upward aggradation and associated offlap, toplap during a relative stillstand, offlap during a lowstand, as well as backstepping stratal-patterns during increasing-upward aggradation. Lowstand surfaces of erosion (Weimer, 1984) and ravinement surfaces (Stamp, 1921; Swift, 1968; Demarest and Kraft, 1987; Nummedal and Swift, 1987) form in documented stratigraphic settings. Downstepping stratal surfaces are produced during a relative fall of sea-level without gravity-flow sedimentation. Stratal termination patterns produced by gravity-flow deposits, including onlap and bi-directional downlap, are also displayed by the simulation.

**Modeling Philosophy**

Building a stratigraphic simulator poses two problems. First, it requires correctly defining the long- and short-term processes that result in the formation and preservation of stratal patterns. This requires an approach that integrates many tectonic and sedimentologic concepts based on studies of ancient rocks and modern process-oriented studies. Many process-oriented studies concentrate on local events with a short time-frame and are not concerned about their associated regional or long-term results (Komar, 1983; Aubrey, 1988; Lakan and Trenhaile, 1989; Seymour, 1989). Second, developing a stratigraphic simulator requires defining a sedimentation model in quantitative terms that responds realistically to a wide range of environmental conditions. An acceptable sedimentation model must be independent of, and correctly respond to all reasonable
tectonic, eustatic and sediment-supply conditions. This sedimentation model properly responds to a wide range of sediment supply, current and bathymetric settings.

The simulator attempts to treat each process as a discrete event. Many functions have been empirically derived to match observations. The events that determine the physical state within a cell are determined by the conditions existing within the surrounding cells as well as the available sediment supply. Unlike approaches using finite-difference solutions that typically converge on a solution that is averaged throughout the model, (such as the diffusion models of Flemings and Jordan, 1989, Jordan and Flemings, 1991; and Kaufman et al., 1990), the influences of many processes do not extend beyond neighboring cells. One of the most important assumptions is that many processes within each cell are only dependent upon the conditions in the closest cells. For instance, before traction load is deposited in one cell, excess sediment must be available from a proximal cell in the direction of the source and the cell surface must be less than base-level. As another example, the erosion of a channel only effects the cells containing the channel and its margins. Although some processes occur independently, other processes are linked. For example, changes in accommodation within the coastal and fluvial plain will influence the character and volume of sediment delivered to the basin as suspension load.

Comparison with Other Computer Models

Many quantitative approaches have been taken to improve our understanding of the stratigraphic record. Table 1 shows that greatest difference between the models is probably the completeness of the work. Only three models attempt to model actual geologic sections or seismic records (this model; Lawrence et al., 1990, and Kendall et al., 1991). PHIL is one of the most complete to date, as it includes flexure loading, compaction, erosion, and a unique depositional system that deposits sediments under gravity-flow conditions.
Computer simulations emphasizing tectonic processes on passive margins and foreland basins demonstrate the importance of flexural response to sediment loading and compaction (Thorne, J., 1985; Steckler et al., 1988; Flemings and Jordan, 1989; Reynolds et al., 1991). The models emphasizing tectonics have shown that in-plane stress may be important in creating local base-level changes (Cloetingh, 1989); viscoelastic rebound can create discontinuities (Quinlan and Beaumont, 1984). Earlier models include those for subsidence and heat flow by Steckler (1981), and modeling coastal onlap in response to lithospheric flexure and long-term sea-level changes by Watts and Thorne (1984). These models simplify the depositional aspects by either introducing measured layers or employing a simple clinoform or a diffusion algorithm that cannot be directly compared to realistic processes.

Models of carbonate sedimentation recognize the highly sensitive nature of the carbonate factory to bathymetric and climatic changes (Kendall and Schlager, 1981). Rocks record frequencies ranging from 19 - 400 ky and are correlated with orbital cycles within the solar system (Matthews and Frohlich, 1991). Most carbonate models produce prograding stratal patterns of a typical carbonate bank (Bice, 1988; Demicco and Spencer, 1989; Read et al., 1990; Goldhammer et al., 1991). All carbonate models employ a depth-dependent production function. Most models employ this function uniformly across a margin and determine lithologies from water-depth. PHIL determines lithologies based on each position within the depositional setting and the transport mechanism.

Models of siliciclastic sedimentation have attempted to predict the distribution of sand (Bridge and Leeder, 1975) and model the role that gradients created by tectonic uplift has on grain-size delivered to the basin (Paola, 1989). Syvitski (1985, 1988) has built a simulation that models deposition and slumping of prodelta sediments within a fjord. Harbaugh (1966), and Tetzlaff and Harbaugh (1989) have built a 3-dimensional dynamic model of near-shore sediment transport that solves the Navier-Stokes equation for fluid transport to predict the motion of particles. The resulting stratal patterns are
approximately similar to a prograding delta. This technique may provide a hydrodynamic basis for calculating the position of various grain-sizes in a depositional profile. This computation-intensive model works at very short time steps, and does not adjust for tectonism or flexural loading or sea-level. BITZER and Harbaugh (1987) have built a 2-dimensional simulator that deposits clastic sediment as a function of water-depth. They have adjusted sea-level and subsidence through time. As with the Navier-Stokes approach, this model predicted deposition of coarse-grained sediment at and below the shelf-slope break. With a 'straight line' depositional interface and infinite sediment supply Pitman (1978) and Pitman and Golovchenko (1983) have demonstrated how long-term changes in sea-level can produce the onlap pattern recorded on simple hinged margins. Collier et al. (1990) have employed a similar 'straight-line' depositional model with infinite sediment-supply and introduced a high-frequency sea-level curve, producing a transgressive and regressive depositional pattern without onlap or downlap.

The gradient difference between the non-marine and marine environment is often two to three orders of magnitude (from 0.065 in the marine environment to 0.0001 or less in the non-marine environment). This contrast alone is capable of producing the coastal onlap associated with base-level changes. The 'triangle method' of Strobel et al. (1989) predicts a sand/shale ratio at each cell, but does not model energy or facies and thus cannot predict the concentration of sands (along the shoreface as an example). Output from their model is not accurate enough to determine the contributions of eustatic changes versus tectonic changes. Strobel et al. (1989) do not calculate a flexural response to sea-level or sediment loading.

The work of Lawrence et al. (1990) is a comprehensive program; it includes a mixed carbonate-clastic depositional system, surface erosion, flexure loading and compaction. The model has successfully simulated seismic profiles from passive margins and carbonate buildups (Aigner et al., 1989). They do not include channel erosion,
shoreface erosion, or a unique depositional system that defines the gravity-flow sedimentation.

**BASIN FILL MODEL**

The following section describes the algorithms that deform the stratigraphic cross-section and the sedimentation algorithms that erode the depositional profile and fill accommodation space. The relation of sedimentation processes to base-level is discussed along with a description of how to model tectonic and climate effects on base-level. A list of the user-defined variables (marked •) is provided with ranges of parameter values and averages that we have measured or extracted from published data where referenced. All variables are set at the initiation of a model and can be varied during the development of the model.

The depositional algorithms model both coastal and gravity-flow sedimentation along a basin margin for both siliciclastic and carbonate sediment. The model generates each stratal or time line through a series of eight steps. The first step adjusts the profile for tectonic and eustatic specifications (Fig. 3A) and locates the intersection of sea-level with the profile. Second, the flexural response of the crust to a change in water column is calculated for the profile (Fig. 3B). Third, erosion may modify the profile where it lies above the base-level of erosion by four mechanisms including: 1) shoreface erosion produced by wave-energy, 2) non-marine channel-incision, 3) surface beveling, and 4) bottom-current transport. The eroded siliciclastic sediment is added to the siliciclastic sediment-supply, eroded carbonate sediment is added to the excess carbonate sediment-supply for redistribution. Fourth, a depositional profile of linked depositional systems defining a clinoform, progrades basinward until the siliciclastic sediment supply is exhausted (Fig. 3C). The siliciclastic deposition algorithm models traction processes with the prograding clinoform, and deposition of suspension load by injection and mixing within the marine water column. During a fifth step, a carbonate factory produces
Figure 3. Creation and fill of accommodation space with a traction load mechanism, A) Initial configuration of depositional setting, B) Adjust sea level for the level at the given time and shift the depositional surface according to the subsidence rates, C) Prograde the depositional interface parallel to sea level in the direction of the basin until the space beneath the clinoform is equal to the sediment supply for the time period.

sediment as a function of the bathymetry, position on the margin and distribution of siliciclastic sediment. The simulator deposits pelagic sediment as a function of water depth in marine settings. The excess productivity is suspension removed and redistributed first in the lagoon and then in the basin below a prograding clinoform. Suspended carbonate sediment is transported offshore in the water column and deposited by settling. During the sixth step, newly deposited sediment with slopes greater than a defined limit of stability is removed above a fault scarp, and transported basinward by slumping. If the rate of relative fall is less than the relative fall limit (relative fall is negative), then the newly deposited sediment that is proximal to the open basin is removed and added to the turbidite sediment-supply. If base-level is above a slope-fan threshold, basin-floor fan deposition occurs in the lowest position of the basin. If base-level falls below a "slope-fan threshold," submarine canyons form, and the model generates channel-overbank fans on the slope and basin floor. After deposition, the
seventh step adjusts the layers for the flexural response of loading or unloading due to sedimentation or erosion at each cell. Eighth, the simulator adjusts the depths of the horizons for compaction as a function of depth.

Base-level is a theoretical surface (Fig. 3C) that determines whether a surface is undergoing sedimentation or erosion (Powell, 1895; Barrell, 1917; Twenhofel, 1939; Wheeler, 1958, 1964). Deposition can occur when the depositional interface is below base-level and accommodation space (space available for sediments to be deposited) is available. Likewise, erosion can occur when the depositional interface is above base-level, and accommodation space is not available. Within the simulator, the position of base-level is linked to the intersection of sea-level and the depositional interface. In the non-marine setting, base-level is defined by the coastal plain and stream profile for siliciclastic sediment, and the upper tidal range for carbonate sedimentation. In the marine setting, base-level lies above fair-weather wave-base for traction-load and suspended-load deposition, although base-level probably extends to storm-weather wavebase with decreasing intensity. The basinward extent of base-level lies at the offlap break. The offlap break is the increase in gradient at the top of the clinoform rollover (Fig. 3C). When sediment fills all the space below base-level, the depositional profile migrates basinward one cell spacing.

Deposition of sedimentation is strongly influenced by the distribution of accommodation space for coastal processes and bathymetric slopes for gravity-flow sedimentation. In the simulator, relative changes of sea-level are controlled by two factors, eustasy and tectonism. Eustasy is equal throughout the cross section and has a predictable variation through time. Tectonism is unequal across the cross section and varies through time. The sedimentation system responds to movement of the base-level surface through time. Motion of base-level is a function of relative sea-level and sediment supply. Vertically, base-level is controlled by the sum of total subsidence and eustasy.
Horizontally, base-level is controlled by the rate of sediment supply -- the higher the rate, the more basinward base-level migrates (Fig. 3C).

**Substrata**

The bathymetry and base of the underlying compacting interval at the initiation of the model are defined by two horizons. The substrata will compact as sediment is deposited above. In younger sections with a thick substrate, this will account for a large portion of accommodation space.

**Tectonism**

We specify tectonic subsidence rates spatially and temporally that produce the observed changes in paleobathymetry. Subsidence is positive and uplift is negative. This information is initially calculated from backstripping analysis or acquired from a dynamic deformation model (Sawyer and Harry, 1991). We add the flexural response of sediment to produce the total subsidence response. Subsidence rates vary laterally and temporally. A vertical fault is simulated by specifying a sharp contrast in subsidence rates between two closely-spaced hinge points. A historical data file modifies the spatial data over time by multiplying the subsidence rate history with the spatial subsidence rates. Changing subsidence rates over time can reproduce patterns such as exponentially declining rates for a cooling passive margin (McKenzie, 1978).

Mature passive margins have low tectonic subsidence rates. Here, total subsidence rates often correlate with the sediment-supply history because most of the subsidence is caused by sediment loading. In settings with a moderate sediment supply, total subsidence rates range from 20 to 90 mm/ky. With a large sediment supply, compaction of underlying sediment and flexural loading will produce very high rates of subsidence. In the Gulf of Mexico, near the mouth of the Mississippi River, subsidence rates are as high as 2000 mm/ky. Tectonically active basins (such as trans-tensional basins) have tectonic subsidence rates that ranges from 40 mm/ky to as high as 250 mm/ky and total subsidence ranges from 100 mm/ky to as high as 1000 mm/ky.
Flexural Loading

The flexural response of the crust to loading or unloading by deposition and erosion of sediment and eustasy is calculated using a homogeneous elastic plate over a liquid half space solution (Turcotte and Schubert, 1982). The response is defined by three variables:

<table>
<thead>
<tr>
<th>Range</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flexural wavelength</td>
<td>1-65 km</td>
</tr>
<tr>
<td>mantle density</td>
<td>3340 Kg/m³</td>
</tr>
<tr>
<td>taper limit</td>
<td>approximately three times the flexural wavelength</td>
</tr>
</tbody>
</table>

During each time increment, the change in bathymetry due to flexural loading is calculated by adding the sediment and subtracting the eroded sediment load, and then adjusting for changes in water load due to eustatic changes. If the surface of the cell is above sea level, unloading by compaction-driven dewatering is removed.

The deflection in each cell is calculated by convolving the load in each cell with the deflection function \( w \). The deflection across the basin, \( w \), is defined by the following equation:

\[
w(x) = \frac{V_0 \alpha^3}{8D} e^{-x/\alpha} \left( \cos \frac{x}{\alpha} + \frac{x}{\alpha} \sin \frac{x}{\alpha} \right) \quad (1)\]

\( \alpha \) is the distance from the load, and \( V_0 \) is the line load applied at \( x = 0 \) and \( D \) is the flexural rigidity of the bending plate. We specify the flexural parameter, \( \alpha \), and use Equation 2 to calculate the flexural rigidity, \( D \):

\[
D = \frac{\alpha^4 g (\rho_m - \rho_w)}{4} \quad (2)
\]

\( \rho_m \) is the mantle density, \( \rho_w \) is the density of water, and \( g \) is the gravitational constant.

We hold Young's modulus, \( E \), constant at 70 GPa, and Poisson's ratio, \( \nu \), constant at 0.25 and use Equation 3 to define the effective thickness of the lithosphere, \( h \):

\[
h = \sqrt[3]{\frac{12D(1-\nu^2)}{E}} \quad (3)
\]
The flexural load at the margins are projected to the taper limit or to the limit of the flexural wavelength, whichever is less, and their influence is calculated within the cross section.

The principle assumption within the flexural calculation is that all sedimentation is distributed equally, perpendicular to the cross section, within the flexural wavelength of the model. This assumption is a good approximation for a passive margin. When modeling a carbonate atoll, the algorithm may overestimate the flexural response of a load. Another assumption is that the flexural response occurs within the time increment of the model. The rate of rebound of the crust in response to ice cap melting, indicates most of the rebound may occur within 1 ky (Peltier, 1990; Forman, 1990). Strand-plain records in regions of glaciation demonstrate that 90-95% of the response occurs within 1 ky. Therefore, for time increments larger than 1 ky, this may be a good assumption.

Compaction

The simulator incorporates sediment compaction as a function of burial depth. Ten siliciclastic and eight carbonate lithologies are used (Table 2) with constants applied to Equation 4 (from Dickinson et al., 1987).

\[
\phi(z) = \frac{\phi_0}{1 + z \cdot r_c}
\]

\(\phi\) is the porosity, \(z\) is depth in meters, \(\phi_0\) is the initial porosity, and \(r_c\) is the compaction coefficient.

After deposition, the simulator calculates from top to bottom, the mean depth for each interval and determines the porosities at that depth for the siliciclastic and carbonate component. The thickness of each interval is decreased as required for the decrease in porosity. The horizon depths are adjusted from bottom to top. The weight loss due to dewatering is calculated for each non-marine cell. If the interval is above sea-level, the water load is subtracted from the sediment load, whereas if the interval is below sea-level,
it is assumed that sea water replaces the water volume lost by compaction, and the load is not changed.

Table 2. List of the lithologies and typical compaction parameters

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Density (kg/m$^3$)</th>
<th>Initial Porosity ($\phi_0$)</th>
<th>Compaction Rate ($\tau_c$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz Boulders</td>
<td>2650</td>
<td>0.4</td>
<td>0.0001</td>
</tr>
<tr>
<td>Quartz Silt Size</td>
<td>2650</td>
<td>0.3</td>
<td>0.001</td>
</tr>
<tr>
<td>Quartz Silt/Clay</td>
<td>2650</td>
<td>0.4</td>
<td>0.002</td>
</tr>
<tr>
<td>Quartz Sand/Clay</td>
<td>2650</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Quartz/Silt</td>
<td>2650</td>
<td>0.45</td>
<td>0.0005</td>
</tr>
<tr>
<td>Silt/Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.002</td>
</tr>
<tr>
<td>Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Silt/Coal</td>
<td>2450</td>
<td>0.6</td>
<td>0.008</td>
</tr>
<tr>
<td>Clay/Coal</td>
<td>2300</td>
<td>0.85</td>
<td>0.009</td>
</tr>
<tr>
<td>Coal</td>
<td>2000</td>
<td>0.92</td>
<td>0.01</td>
</tr>
<tr>
<td>Cemented Carbonate</td>
<td>2800</td>
<td>0.45</td>
<td>0.0001</td>
</tr>
<tr>
<td>Carbonate Fine grainstone</td>
<td>2800</td>
<td>0.6</td>
<td>0.001</td>
</tr>
<tr>
<td>Carbonate Boundstone</td>
<td>2800</td>
<td>0.6</td>
<td>0.001</td>
</tr>
<tr>
<td>Carbonate Coarse grainstone</td>
<td>2800</td>
<td>0.7</td>
<td>0.002</td>
</tr>
<tr>
<td>Micrite</td>
<td>2800</td>
<td>0.7</td>
<td>0.004</td>
</tr>
<tr>
<td>Algal Laminates</td>
<td>2800</td>
<td>0.6</td>
<td>0.0005</td>
</tr>
<tr>
<td>Dolomite</td>
<td>2900</td>
<td>0.4</td>
<td>0.0001</td>
</tr>
<tr>
<td>Gypsum</td>
<td>2330</td>
<td>0.1</td>
<td>0.00001</td>
</tr>
</tbody>
</table>

In stratigraphic sections under lithostatic pressure, compaction responses occur simultaneously with deposition and do not effect the depositional pattern of the sediments, only their thickness and extent of progradation. PHIL assumes the sediment maintains lithostatic pressures. However, in basins with hydrostatic pressures or over-pressured sediments, compaction is delayed. Here the section will remain thicker and less dense. When modeling a hydrostatic setting, PHIL would predict more sediment, lower porosities and higher densities than observed.

**Eustasy**

Eustasy raises and lowers sea level equally throughout the model. We consider two forms of defining sea level including:

- Sea level cycle period, amplitude, phase, and character (sinusoidal or saw-tooth)
- age, sea level (Ma, meters above present sea level).

This information is specified as a sum of sinusoidal or saw-tooth sea-level functions or a digitized sea-level history (Fig. 4). Up to six different cycles can be superimposed and
adjusted by their period, phase, and amplitude. These cycles can be added to the digital sea-level file. This sea-level history is measure in meters with negative values above sea-level, and positive values below sea-level.

**Siliciclastic Deposition**

Siliciclastic deposition transports sediment by traction and suspension processes, slumping, and gravity-flow sedimentation in marine settings. Siliciclastic sediment is derived from erosional sources, as well as specified as either a constant or cyclic input, or as a digital sediment influx history. The siliciclastic sediment-supply is defined by three variables:

<table>
<thead>
<tr>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>age, supply rate pairs</td>
<td>0-2.0E6 m2/ky</td>
<td>1.0E4</td>
</tr>
<tr>
<td>% transported by Traction</td>
<td>0-100%</td>
<td>80%</td>
</tr>
<tr>
<td>% sand size</td>
<td>0-100%</td>
<td>20%</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>age, supply rate pairs</td>
<td>0-2.0E6 m2/ky</td>
<td>1.0E4</td>
</tr>
<tr>
<td>% transported by Traction</td>
<td>0-100%</td>
<td>80%</td>
</tr>
<tr>
<td>% sand size</td>
<td>0-100%</td>
<td>20%</td>
</tr>
</tbody>
</table>

**Figure 4.** Up to six eustatic cycles can be superimposed. Each cycle is defined by the period, amplitude, phase and character of the eustatic cycle (sinusoidal or saw-tooth function). The peak or trough of the saw-toothed function is defined by the percent of the cycle composed of the first segment. The sign of the amplitude changes the direction of the first leg.

The traction algorithm constructs the siliciclastic depositional profile with the following variables (Fig. 5):
Gradients are a unit independent ratio (meter/meter or feet/feet etc.) that measures elevation loss per distance traveled.

![Diagram of a siliciclastic profile shape established by traction processes.](image)

**Traction**

Traction processes move sediment from the margin of the model into the basin along the sediment surface. This profile is tied to sea level at the shoreface. A linked set of fluvial, coastal and shore face depositional interfaces form segments on a continuous depositional profile and defines base-level (Jervey, 1988). The boundary between the fluvial and coastal plain is the bay line. The boundary between the coastal plain and shoreface is the shoreline. The boundary between the upper shoreface and lower shoreface is the offlap break. The offlap break is the top of the depositional front. When a sediment packet reaches a location that lies below base-level, the algorithm removes the required volume of sediment from the sediment supply and fills the location to base-level. When all the space below base-level is filled the depositional interfaces prograde one cell spacing basinward, parallel to sea level (Fig. 3C). The profile progrades into the
available space until the volume of sediment transported by traction processes for the time increment is exhausted.

Fluvial gradients of aggrading stream profiles are difficult to document because the most recent global sea-level rise has left many streams under-filled. Many recorded fluvial gradients are in regions experiencing uplift and are controlled by the erosion rates of the substrata (Leopold and Wolman, 1957). However, in most cases, the last transgression has forced streams landward and filled their previously incised and flushed valleys; the Mississippi River Valley is an example (Fisk, 1944). Here, fluvial gradients are very low (0.0005 - 0.00001). During times of rapid sea-level fall, the stream undergoes downcutting. Basinward of the knickpoint the fluvial gradient increases (Begin et al., 1980).

The geometry of the coastal plain is defined by the coastal-plain width and gradient. The coastal plain is typically flat and therefore has a gradient close to zero. Values for coastal-plain width are poorly documented. Competing marine and fluvial processes interacting with coastline irregularities control the coastal-plain width. In settings where wave activity dominates, the coastal plain is very narrow (less than 1 km). In contrast, settings with a strong fluvial influence, such as the Mississippi delta, have a very broad coastal plain (greater than 200 km). However, as the rate of relative sea-level change approaches zero (stillstand), the coastal plain fills with sediment and the fluvial system progrades over earlier-deposited coastal-plain sediments. A more precise definition of the position of fluvial progradation requires knowledge of the sediment supply and the hydrodynamics of the fluvial and marine environments. This is a 3-dimensional problem that is beyond the present scope of this simulator. It is assumed that deltaic processes are capable of prograding without gravity-flow sedimentation of the coarser fraction (except for special bathymetric settings) when the sediment influx rate does not produce over pressured conditions. In nature, where over pressured conditions occur with high sediment flux rates, growth-faults form and carries sediment basinward.
The simulation does not form growth faults but modeling brackets the sediment influx above which over pressuring occurs between 20,000 and 60,000 m$^2$/ky.

If the barrier island height is greater than zero, a barrier island will form. The island may also be removed by ravinement processes during subsequent relative sea level rises.

The depth of the offlap break typically corresponds to the fair-weather wave-base. In regions with mild winds, as in the Gulf of Mexico, the offlap break is 5-10 meters (Coleman and Wright, 1975; Coleman, 1982). In regions with heavy winds, as Southwest Australia, the offlap break extends to 20 meters. The transition between the shoreface and depositional front is defined by the rollover width.

**Suspension**

Suspended sediment is injected into the marine environment and settles as a function of the mixing volume within a hypopycnal inflow (Bates, 1953). This dispersion process is defined by two variables:

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
<th>This Report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Width of the mixing layer</td>
<td>1-200 meters</td>
<td>20 meters</td>
<td></td>
</tr>
<tr>
<td>Dispersion distance</td>
<td>5 - 100 km</td>
<td>50 km</td>
<td></td>
</tr>
</tbody>
</table>

Sediment settles from an injected stream as it mixes with sea water with each incremental distance from the fluvial-marine contact (Fig. 6). We empirically derived a function that simulates deposition of suspension sediments is defined by:

$$ T(x) = \frac{V_w V_s}{50000 C_d} \quad (5) $$

$T$ is the thickness of sediment deposited in a cell, and $V_w$ is the volume of water that the sediment has passed to reach that cell. This volume is described by $V_w = V_p + V_x$, where $V_p$ is the total volume of water encountered in previous cells. $V_x$ is the volume of water in the mixing cell, $V_x = M C$, where $M$ is the depth of the mixing column (meters), $C$ is the cell spacing (meters). $V_s$ is the volume of sediment that reaches each cell. $C_d$ is a dispersion distance constant and represents the distance in kilometers that the injected
stream extends into the marine setting. As the suspension load is carried further into the basin, a fraction of the remaining sediment supply is removed that is proportional to the thickness of the mixing layer. No deposition occurs on a bathymetric surface that is above fairweather wavebase.

**Peat Formation**

Peat formation is simulated within the coastal plain as a function of accommodation rates, the transitions between different ratios of peat and silt are defined by the following variables.

<table>
<thead>
<tr>
<th>Category</th>
<th>Range (mm/ky)</th>
<th>Average (mm/ky)</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Silt</td>
<td>1 - 20</td>
<td>5</td>
<td>&quot;</td>
</tr>
<tr>
<td>Silty 75% - Clay 25%</td>
<td>20 - 100</td>
<td>50</td>
<td>&quot;</td>
</tr>
<tr>
<td>Clay 50% - Peat 50%</td>
<td>100 - 1000</td>
<td>150</td>
<td>&quot;</td>
</tr>
<tr>
<td>Peat 75% - Clay 25%</td>
<td>1000 - 2000</td>
<td>1500</td>
<td>&quot;</td>
</tr>
<tr>
<td>Peat</td>
<td>1500 - 2500</td>
<td>2000</td>
<td>&quot;</td>
</tr>
<tr>
<td>Lacustrine</td>
<td>&gt;2500</td>
<td>&gt;2500</td>
<td>&quot;</td>
</tr>
</tbody>
</table>

In conditions with accommodation rates higher than the peat rate, lacustrine sediments are deposited.

Often the model dimensions are not big enough to contain all the depositional activity. If the bathymetry on the supply-side margin approaches 250 meters, it is
assumed that deposition is occurring at a distance too great to reach the cross section and the sediment supply is reduced to zero. Likewise, if there is no space in the cross section below base-level, deposition will not occur.

**Carbonate Deposition**

Carbonate deposition is unique because the biota, bathymetry, and climate, control the volume and type of sediment produced. The depositional profile configuration is a result of carbonate production and subsequent redistribution, processes that occur simultaneously in nature (Wilson, 1975; Lerche et al., 1987). The thickness at each cell, therefore, is a record of sediment accumulation rates, not production rates. PHIL represents the carbonate production system with four water-depth-dependent functions including. Coarse-grained restricted traction-load production that is composed of an unrestricted factory, a shelf-margin factory. Fine-grained suspension-load production is composed of an unrestricted fine-grained factory, and a pelagic factory. Sediments are redistributed by traction processes from regions of high productivity and low preservation potential, such as lagoons, to regions with high preservation potential, such as tidal flats.

![Diagram](image)

**Figure 7.** Carbonate productivity versus depth. The function is defined by the width of the production function, depth of maximum production and maximum production rate. The upper limit is defined by the top of the tidal range. Productivity is linearly interpolated between the top and bottom of the tidal range.
Fine-grained sediment is generated in the water column on the platform and transported by suspension to the basin proximal to the platform (Boardman and Neumann, 1984).

The factories produce carbonate sediment as a function of bathymetry and turbidity (Huston, 1985). The productivity functions are normal distribution curves with a specified width and maximum production at a specified bathymetry. Production is linearly interpolated between the calculated value at low water and 0 mm/ky at high water (Fig. 7). Unrestricted traction and fine-grained production will occur on any interface below high water. Shelf-margin production is centered about an optimal location for production with respect to the open basin and reduced exponentially with distance from that location by a specified factor. The carbonate production is controlled by the following variables:

"Unrestricted"

<table>
<thead>
<tr>
<th>Variables</th>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum growth rate</td>
<td>50 - 500 mm/ky</td>
<td>100 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Depth of maximum growth</td>
<td>2 - 25 meters</td>
<td>8 meters</td>
<td></td>
</tr>
<tr>
<td>Width of the depth function</td>
<td>5 - 40 meters</td>
<td>20 meters</td>
<td></td>
</tr>
</tbody>
</table>

"Shelf-margin"

<table>
<thead>
<tr>
<th>Variables</th>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum growth rate</td>
<td>10 - 500 mm/ky</td>
<td>140 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Depth of maximum growth</td>
<td>1 - 20 meters</td>
<td>12 meters</td>
<td></td>
</tr>
<tr>
<td>Width of depth function</td>
<td>5 - 40 meters</td>
<td>20 meters</td>
<td></td>
</tr>
<tr>
<td>Width of distance function</td>
<td>0.1 - 5 km</td>
<td>1.5 km</td>
<td></td>
</tr>
</tbody>
</table>

"Fine-grained"

<table>
<thead>
<tr>
<th>Variables</th>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum growth rate</td>
<td>10-100 mm/ky</td>
<td>20 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Depth of maximum growth</td>
<td>2 - 25 meters</td>
<td>8 meters</td>
<td></td>
</tr>
<tr>
<td>Suspension distance</td>
<td>1-50 km</td>
<td>15 km</td>
<td></td>
</tr>
<tr>
<td>Maximum suspension depth</td>
<td>3-40 m</td>
<td>15 meters</td>
<td></td>
</tr>
</tbody>
</table>

"Pelagic"

<table>
<thead>
<tr>
<th>Variables</th>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum production rate</td>
<td>1 - 40 mm/ky</td>
<td>10 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Calcite Compensation Depth</td>
<td>3 - 4.5 km</td>
<td>4 km</td>
<td></td>
</tr>
<tr>
<td>Dissolution width</td>
<td>500-2000 km</td>
<td>500 km</td>
<td></td>
</tr>
<tr>
<td>Siliciclastic damping limit</td>
<td>1 - 150 mm/ky</td>
<td>100 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Production time increment</td>
<td>1 - 25 ky</td>
<td>3 ky</td>
<td></td>
</tr>
</tbody>
</table>

These variables define the productivity by the following normal distribution function centered at the bathymetry of maximum production:

\[ P_{\text{depth}} = M \cdot T \cdot R \cdot \exp \left(-\frac{(B - D_{\text{mp}})^2}{W^2}\right) \]  

(6)
$P_{\text{depth}}$ is the production for a cell during the time increment, $M$ is the maximum production rate (mm/ky), $T$ is time (ky), $R$ is the siliciclastic reduction factor (described below), $B$ is the bathymetry in meters, $D_{\text{mp}}$ is the maximum production bathymetry (meters), and $W$ is the width of the productivity function (meters). Changing the bathymetry of maximum production shifts the function up or down. The width of the productivity function specifies the width of the exponential function.

The bathymetry is analyzed for shelf-margin production (Eq. 8) by convolving a depth-dependent function (Eq. 6) and a damping function (Eq. 7) that reduces the production as function of distance from optimal location for production (shelf-margin crest). The damping function is a normal distribution centered about the shelf-margin crest:

$$P_{\text{distance}} = \exp\left(-\frac{x^2}{D^2}\right) \quad (7)$$

$P_{\text{distance}}$ is the distance factor for each cell, $X$ is the distance from the shelf-margin in kilometers and $D$ is the distance factor in kilometers. The shelf-margin carbonate production is defined by:

$$P_{\text{shelf-margin}} = P_{\text{depth}} \times P_{\text{distance}} \quad (8)$$

$P_{\text{shelf-margin}}$ is the shelf-margin carbonate production at each cell. Studies of recent coral growth show much higher rates, 5000 to 10000 mm/ky (Schlager, 1981; Hubbard and Scaturo, 1985; Huston, 1985), than we have needed to simulate sections. For carbonate sedimentation, the simulator will iterate through a smaller time increment than the time increment designated for the recording of stratigraphic surfaces within the model.

Preliminary studies suggest that siliciclastic sediment will reduce the productivity of the carbonate system (Mount, 1984; Cortes and Risk, 1985). We have empirically derived the following relationship for carbonate productivity reduction:

$$R = 1 - \sqrt{\frac{S}{C}} \quad (9)$$
R is the reduction factor, C is the rate limit of siliciclastic sedimentation (mm/ky) above which no carbonate production will occur, S is the siliciclastic sedimentation rate and suspended sediment flux for the cell (mm/ky).

Organically derived pelagic sediment is produced as a function of bathymetry. The pelagic productivity function matches observations that production is 0 mm/ky with no water depth and exponentially increases to a maximum at 100 meters of paleobathymetry and is defined by:

$$P = M \cdot T \left(1 - \frac{1}{\exp(0.1 \cdot B)}\right)$$  (10)

P is the production for a cell during the time increment (mm/ky), M is the maximum pelagic production rate (mm/ky), T is time (ky), and B is the bathymetry (meters). Pelagic sediment is deposited on all localities that are greater than the bathymetry of the offlap break. The ODP Site 625 from the Gulf of Mexico contains both pelagic and suspended siliciclastic sediments (Joyce et al., 1990). The minimum sedimentation rates average 30-40 mm/ky and may indicate the pure pelagic background sedimentation (Fig. 8) but this is high compared to 10 mm/ky in other ODP wells. Pelagic sedimentation is reduced below a calcite compensation depth (CCD) to simulate dissolution. Modern CCD's range between 3 and 4.5 km, earlier values are shallower (Opdyke and Wilkinson, 1988).

An erosion-redistribution function disperses excess carbonate sediment from the cells that produce more sediment than can be stored to cells with space below baselevel. Excess sediment is located by superimposing the carbonate depositional profile on the margin, aligned at the shelf-margin crest. The algorithm strips off the sediment deposited above the profile. Using the stripped sediment-supply, the fill algorithm initially fills the lagoon behind the shelf-margin. If the lagoon is full, the algorithm transports the sediment into the basin, and the profile progrades. The fill algorithm is based on observations from many atolls that show carbonate buildups that rim the margin during sea-level rise and lagoons filling during stillstands (Kendall and Schlager, 1981).
On extensive carbonate platforms where the lagoon is filled above sea level, the shelf margin will prograde basinward.

Five depositional interfaces are represented by sabkha, tidal flat, back reef, shelf-margin crest, and fore-slope gradients (Fig. 9). The offlap break is defined as the shelf-margin crest.

<table>
<thead>
<tr>
<th></th>
<th>Range</th>
<th>Average</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sabkha gradient</td>
<td>0.0 - 0.00001</td>
<td>0.0</td>
<td>This report</td>
</tr>
<tr>
<td>Tidal flat gradient</td>
<td>0.0001 - 0.0006</td>
<td>0.0003</td>
<td>Ginsburg and Hardie, 1975</td>
</tr>
<tr>
<td>Tidal range</td>
<td>0 - 15 meters</td>
<td>0.5 meters</td>
<td>Reading, 1982</td>
</tr>
<tr>
<td>Backreef gradients</td>
<td>0.1 - 0.001</td>
<td>0.02</td>
<td>Wilson, 1975</td>
</tr>
<tr>
<td>Shelf-margin crest</td>
<td>1.0 - 20.0 meters</td>
<td>8 meters</td>
<td>This report</td>
</tr>
<tr>
<td>Fore-slope gradients</td>
<td>1.1 and 0.01</td>
<td>0.25</td>
<td>&quot;</td>
</tr>
<tr>
<td>Rollover width</td>
<td>0.5 - 3 km</td>
<td>0.8 km</td>
<td>&quot;</td>
</tr>
</tbody>
</table>
The tidal-flat width is determined by projecting the tidal-flat gradient above sea-level until the surface reaches an altitude equal to half of the tidal range. Sediment accumulation landward of the tidal flat is classified as a sabkha setting. With a low back-reef gradient, the productive surface area is increased and the total carbonate production will be higher for the platform as a whole. Coarse grain-sizes and early cementation are associated with steeper gradients of the fore-slope. The rollover width defines the distance over which the transition from backreef to fore-slope gradients occurs.

**Gravity-flow sedimentation**

Gravity-flow sedimentation is defined by five variables:

<table>
<thead>
<tr>
<th>Variable</th>
<th>Range</th>
<th>Average</th>
<th>This report</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum bathymetric relief</td>
<td>100-400 meters</td>
<td>200 meters</td>
<td></td>
</tr>
<tr>
<td>Slope-fan threshold depth</td>
<td>0-30 meters</td>
<td>20 meters</td>
<td></td>
</tr>
<tr>
<td>Basin-floor fan gradient</td>
<td>0.001 - 0.0001</td>
<td>0.001</td>
<td></td>
</tr>
<tr>
<td>Slope-fan gradient</td>
<td>0.01 - 0.001</td>
<td>0.003</td>
<td></td>
</tr>
<tr>
<td>relative-sea-level trigger-factor</td>
<td>0 to -250 mm/ky</td>
<td>-30 mm/ky</td>
<td></td>
</tr>
<tr>
<td>Turbidite volume factor</td>
<td>0 - ?</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

Within the simulation, gravity-flow sedimentation to deep water occurs when either 1) space above a more stable substrate is exhausted, or 2) during a relative fall of sea-level at the offlap break and coastal onlap falls below the offlap break of the previous highstand on a margin with greater than 200 meters relief (Fig. 10). This sediment is deposited as basin floor or channel-overbank fans in the distal portions of the basin.
PHIL continuously removes sediment on unstable slopes and deposits it as slumps. Cells containing surface gradients greater than a stable gradient (depositional front gradient for siliciclastic or fore-slope gradient for carbonate sediment) will collapse by gravity-flow transport. Unstable material is removed above a slump scarp and carried downslope as long as momentum is greater than zero. The simulator uses empirically derived functions that monitor the momentum of a sediment packet as it moves over the sediment surface. If material passes over a gradient that is less than the stable gradient the momentum is reduced. When momentum drops below a threshold, a fraction of the sediment packet is deposited in each cell as slumps or turbidites. The packet migrates basinward until all the sediment is deposited.

Figure 10. Two conditions that produce gravity-flow sedimentation. The basin must be deeper than the minimum bathymetric contrast (100 - 200 meters). Condition 1 will produce basin-floor fans and slope-fan complexes from the volume of interval 2. Condition 2 will produce slumps.
Gravity-flow sedimentation will produce detached basin-floor fans in the early stage of a relative fall, and attached channel-overbank fans in the late stage. The transition is determined by comparing the magnitude of the fall with the slope-fan threshold depth. Basin-floor fans are deposited in bathymetric lows in the lowest position of the basin with the sand fraction of the sediment supply. Basin-floor fan surfaces are smooth and flat when compared to surfaces of channel-overbank fans at the time of deposition (Vail et al., 1991). The program locates the axis of a channel-overbank fan where the momentum of the sediment packet migrating down the slope reaches zero; the deposition pattern radiates from this point. Surfaces of channel-overbank fans are steeper than the surfaces of basin-floor fans.

The turbidite volume factor specifies the relative volume of sediments that are deposited in gravity-flow systems. This factor changes the sediment supply delivered to the deep basin. It is helpful if turbidite and shelf sources are different. If the turbidite volume factor is 1, then all the sediment volume introduced to the shelf during each time increment is deposited as turbidites. A turbidite volume factor of 0.1 produces a turbidite with a volume that is one-tenth the sediment supply.

The relative change of sea-level trigger-factor (R) specifies the limiting rate of fall that relative sea-level must reach before gravity-flow sedimentation will occur. If the rate of relative sea-level change is greater than R, then the clinoforms are stable. If the rate of relative sea-level change is less than R, the clinoforms are unstable and sediments bypass into the deep basin and deposit as turbidites. This factor controls the fraction of time within a eustatic cycle that is represented by gravity-flow sedimentation, as well as the amount clinoforms will downstep during a relative fall. A relative change of sea-level trigger-factor of -30 mm/ky produces gravity-flow of coarse-grained material during roughly 80% of a relative fall when total subsidence is zero. A smaller relative change of sea-level trigger factor produces less time for gravity-flow sedimentation and more downstepping clinoforms.
The user can vary the amount of time and volume represented in gravity-flow sediments. If the amount of time represented is too great, then the relative sea-level trigger factor should be decreased. If the volume of sediment in the channel-overbank fan or basin-floor fan is too small the turbidite volume factor should be increased.

**Definition of lithologies**

Siliciclastic lithologies are based upon the prescribed sediment supply and the location of each cell within the depositional system (Fig. 11). The fluvial plain contains coarse sand and mud. The boundary between the coastal plain and the shoreface occurs at the intersection of the depositional profile and sea-level. The coastal plain contains sand, silt and mud when deposited under conditions of slow relative rise. In humid settings, an increasing percentage of coal is deposited with faster rate of relative rise. Coal formation and lakes are common within the coastal plain during rapid increases of accommodation potential. Increases in accommodation potential create conditions of low

![Diagram](image-url)
siliciclastic influx and extensive flooding within the coastal plain (0.1 - 1 m). If the climate is humid, marshes will form if the plant growth can keep up -- or lakes if plant growth cannot. Shoreface deposits are the first units to contain sand. The boundary between the depositional front and suspended sediment is at the clinoform base. The depositional front environment is characterized by silt and mud. Suspension deposits beyond the depositional front are fine-grained mud. Basin-floor fans are thought to be sourced from sand-dominated fluvial and shallow marine sediment and are defined as pure sand. The channel-overbank fans form as valleys incise and muddier sediments are eroded in addition to the fluvial source. The channel-overbank channels fill with sand. 

The sediment grain-size decreases with distance from the channel.

![Figure 12. Depositional systems within the carbonate environment.](image)

Carbonate lithologies are based upon the prescribed sediment supply and the location of each cell within the depositional systems (Fig. 12) (Wilson, 1975, p. 64-69). The carbonate depositional systems are controlled by the interaction of the carbonate "factory" with the antecedent bathymetry. The carbonate buildup is initiated at the first cell from the open basin on the profile that is optimal for production. Sediment basinward of this cell is fore-slope fill and designated as grainstone. Sediment between the shelf-margin and the crest of the platform is defined as reef and designated as boundstone. If a lagoon has formed, the sediment from the crest of the buildup to the base of the lagoon is defined as backreef and designated as grainstone. Sediment from the crest of the buildup to the base of the tidal flat is designated as wackestone. The tidal flat extends from the crest of the buildup when there is no lagoon, or the base of the lagoon when one exists, to the landward extent of the tidal range and is designated as
mudstone. Deposition landward of the tidal flat is within the sabkha and is designated as a mixture of dolomite and evaporite. Carbonate sediment that builds up on gradients greater than the fore-slope is shed into the nearby basin as slump blocks and calciturbidites; and designated as grainstone. Pelagic sediment is designated as mudstone.

**Erosion**

Erosion is modeled in four different modes; surface beveling and channel incision in the non-marine environment, and shoreface erosion and current reworking in the marine environment (Fig. 13). All eroded sediment is added to the clastic sediment supply and redeposited. Variables that specify the erosion process include:

<table>
<thead>
<tr>
<th><strong>Range</strong></th>
<th><strong>Average</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>• Shoreface erosion rate</td>
<td>0 - 150 sq. km/ky 0.1 sq. km/ky This report</td>
</tr>
<tr>
<td>• Surface erosion rate</td>
<td>0 - 5000 mm/ky 1000 mm/ky &quot;&quot;</td>
</tr>
<tr>
<td>• Channel erosion rate</td>
<td>0 - 100000 mm/ky 4000 mm/ky &quot;&quot;</td>
</tr>
<tr>
<td>• Channel depth</td>
<td>0 - 1000 meters 4 meters &quot;&quot;</td>
</tr>
<tr>
<td>• Channel margin gradient</td>
<td>0.01 - 1 0.1 &quot;&quot;</td>
</tr>
<tr>
<td>• Channel spacing</td>
<td>1 - 100 km 10 km &quot;&quot;</td>
</tr>
<tr>
<td>• Marine current smoothing width</td>
<td>0 - 2x cell spacing &gt; cell spacing &quot;&quot;</td>
</tr>
</tbody>
</table>

Shoreface erosion is calculated by progressively stepping landward and removing sediment below an erosive profile until a defined volume of sediment is removed. The base of the erosive profile is defined by the storm-weather wave-base (two times the depth of the offlap break). The shoreface erosion is limited by the shoreface erosion rate.

After completion of the shoreface-erosion process, fluvial incision occurs in the non-marine portion of the margin (Fig. 13). A base-level of erosion is defined by superimposing the siliciclastic depositional interface, referenced from the coastline, on the margin. The region of potential erosion is defined by sediment that is above base-level. Channel incision occurs in this region in random locations within segments that have a width equal to the channel spacing. The potential depth of the channel is determined by subtracting the channel depth from the base-level at the cell. The channel will erode the magnitude defined by the erosion rate and the time increment. Erosion will not cut below
the potential depth of the channel. The shape of the channel is cut according to the channel margin gradient.

Surface beveling erodes the resulting surface (Fig. 13). The rate of surface beveling is a function of the average surface gradient on either side of the cell. We developed Eq. 11 to define the amount of sediment eroded at each cell above base-level:
\[ S = 0.001 \ E \ T \frac{R_r+R_l}{2 \ C} \]  

(11)

S is the amount of sediment eroded (m), 0.001 converts centimeters to meters, E is the erosion rate (mm/ky), T is the time increment (ky), R_r and R_l are the differences in relief (meters) between the cell of interest and the cells on the right and left, respectively, and C is the cell spacing (meters).

After deposition the new sediment layer is analyzed in the marine setting for slope stability. Sediment deposited on unstable slopes is moved down-slope. Unstable gradients are greater than 1.25 times the stable gradients of the depositional front for siliciclastic sediment and the fore-slope for carbonate sediment. If a surface slope is greater than the unstable gradient, a slump scarp dipping the gradient of the stable gradient is projected upward through the sediment until the scarp reaches the surface. The material lying above this scarp is removed and transported down-slope until momentum reaches zero. All slumped material is deposited in this region as a slump block.

Finally, the program averages the marine sediment surface over a specified smoothing width. This produces an effect similar to erosion of topographic highs and redeposition in topographic lows by marine currents. The smoothing width is greater than the cell spacing of the model, or no averaging will occur.

**Time and Space**

Time and space are two variables that regulate processes controlling the geometry of strata. Most sediment is deposited or eroded by ephemeral processes that occur over periods of a few days to hours during and after a storm. Their effect must be averaged when selecting an algorithm to represent their observable results over long time-intervals (100-1,000,000 years) used in computer simulations of basin formation. This is accomplished by emulating the resulting form of a given process such as a prograding clinoform for traction deposition. Many processes iterate through small increments until
the sediment supply is completely distributed. Many process rates are also a function of the character of the surrounding region. For example, cliff-forming strata will experience slow rates of erosion where the exposure has low gradients, and high rates of erosion where the exposure has steep gradients, such as at the cliff face.

The simulator uses equal time increments to analyze the basin conditions and record stratal lines every 1 year to 2 My. The user specifies a starting and ending time. In the sedimentation algorithm, all processes are calibrated with time to study the effects of different time increments on the expression of processes. In many cases, the size of the time increment will not matter. The time increment is least important when modeling siliciclastic deposition. However, some processes such as carbonate deposition will produce dramatic differences between models with short time-increments versus those with long time-increments. For instance, with a long time-increment, a major relative sea-level rise may occur between time steps, drowning a platform that normally may have kept up in a simulation based on a short time-increment. For this reason, carbonate deposition has an independent time-increment that is usually smaller than the model time-increment.

It is important to set the time increment short enough to resolve the details of the process of interest (Fig. 14). When modeling the interaction of sedimentation and faulting (continuous or intermittent movements with periodicity of 50–200 years between earthquakes), set the time increment to 1/2 or 1/4 of the shortest periodicity involved. If you are modeling a 1500 ky eustatic cycle, a time increment of 100 ky would represent it very well with fifteen time slices.

We have built models that range from 20 km for outcrop and reservoir studies to as wide as 800 km for basin studies. The simulator records the stratigraphy at the cell spacing. We have used a cell spacing that ranges from 50 to 2000 meters. The cell spacing controls the lateral resolution of the model. Some processes, such as
Figure 14. A 3.5 My model containing two 1.5 My eustatic cycles modeled with 400, 100 and 41 ky timesteps. A 100 ky time step best represents the details of the eustatic cycle. A 400 ky time step does not record some prograding and backstepping parasequences. A 41 ky time step uses more layers than it needs to record the 1.5 cycles but adds details like slumps at the end of the first lowstand prograding systems tract.
progradation, step from cell to cell. Therefore, the precision of the shoreline position is within one cell spacing. All spatial and sea level measurements use the convention that depths are positive. A relative fall is negative.

Graphic Presentations

One objective of stratigraphic modeling is to generate a reasonable likeness to seismic data (Fig. 1a and b). Comparison of depth converted correlated horizons leads the geologist to suggestions for improving the model, or interpretation (Fig. 1c). The stratal age figure (Fig. 1d) plots the age of time-lines recorded at the end of each time step in the cross section. These time-lines are parallel to primary reflectors on seismic data, and form stratal surfaces and associated discontinuities (Vail et al., 1977b; Greenlee and Moore, 1988). The relative sea-level figure (Fig. 1e) plots the given eustatic and tectonic histories, with the resulting total subsidence, and relative change of sea-level history (sum of total subsidence and eustasy), at the position of the offlap break through time. The vertical lines represent the geologic time of stratal lines and are coded for systems tract boundaries. The chronostratigraphic diagram (Fig. 1f) plots the thickness of each horizontal time-line according to its position in the cross section. The porosity model shows the distribution of porosities in the cross section (Fig. 1g). The lithofacies figure (Fig. 1h) plots the distribution of lithofacies in the cross section. The depositional systems figure (Fig. 1i) plots the distribution of siliciclastic and carbonate depositional systems in the cross section. The thickness of the interval is represented by the height of the interval at each cell. The sediment is colored according to the paleobathymetry at the time of deposition. Values of time-variant input such as tectonic history, siliciclastic sediment-supply, or eustasy are plotted on the left margin. The paleobathymetry figure (Fig. 1j) plots the bathymetry in which the sediments were deposited. The sequence stratigraphic figure (Fig. 1k) plots the systems tracts in the cross section. Stratigraphic columns (Fig. 1l) can be drawn for each cell for comparison with wells or sections. The intervals may be filled either with the lithofacies, systems tract, or paleobathymetry of the
interval. The column contains the stratal surface type, resistance to erosion profile, and geologic age of the interval. Missing time is represented by an arrow; its width is proportional to the amount of missing time.

ANALYSIS STRATIGRAPHIC CROSS-SECTIONS

Given a complex geologic data-set containing tectonic, sediment-supply, and eustatic variations, it is possible to separate the history of each variable by analysis of its stratigraphic signature. The procedure for separating the factors controlling the geometry of stratal patterns and their associated lithofacies (Fig. 2) involves a six stage procedural loop of: (1) stratigraphic analysis, (2) integrating outcrop, well-log, and seismic interpretation with biostratigraphy and a global cycle chart, (3) digitizing key observations, (4) calibration of input parameters, (5) simulation of the geologic cross-section, and (6) comparison of results with observations. The signatures of tectonic subsidence, eustasy and sediment-supply variations are distinguished by their frequency, magnitude and expression in stratal patterns.

Stratigraphic Analysis

Stratigraphic analysis involves transgressive and regressive facies-cycle wedge analysis, interpretation of sequences and systems tracts, and correlation with a global chronostratigraphic framework (Vail et al., 1991). This analysis can provide the detailed framework necessary to correlate through complicated facies transitions and structures, as well as a means to estimate the lithofacies distribution. Time lines are provided by physical stratigraphy defined by correlative shifts in lithofacies and stratal termination patterns. A correlation between the well log character and the lithologies in the region must be established. It is also very helpful to establish an age and paleobathymetric framework with biostratigraphic analyses of local wells.

Transgressive and regressive facies-cycle wedge analysis defines the long-term position of the shoreline or offlap break and documents episodes of folding or faulting,
major flooding surfaces and unconformities (White, 1980). The long-term position of the shoreline is compared with long-term sea-level history and a subsidence history (tectono-eustasy) analysis to determine the causes of the features. Tectono-eustasy is a long-term variation that cannot be separated from tectonic effects without comparing sections from tectonically different basins and correlating with stable platforms. Estimation of tectono-eustasy using approaches such as those of Kominz (1984) and Sahagian (1990) work well to provide the long-term sea-level history for much of the Mesozoic and Cenozoic. Much of the long-term curve of Haq et al., 1987 was based on this work.

Sequences and systems tracts are interpreted on seismic data, well-logs, and outcrops to define the geologic control. Seismic data is interpreted for sequences and systems tracts by identifying the stratal terminations and grouping in terms of onlap, downlap, and toplap patterns. To interpret sequences and systems tracts on well-log data requires defining (1) maximum flooding surfaces, (2) fining and coarsening upward trends, and (3) surfaces of onlap or omission. Each systems tract has a unique accommodation pattern such as lowstand prograding (increasing-upward aggradational-progradation), transgressive (increasing-upward aggradational-backstepping), and highstand (decreasing-upward aggradational-progradation).

Global correlation studies document the magnitude and timing of both long-term and third-order eustatic changes. For a eustatic change to be considered global, it must be correlated between basins on different continents with diverse tectonic settings, using bio-, geochrono-, and magnetostratigraphy (Kominz, 1984; Haq et al., 1987; Sahagian, 1987; Bartek et al., 1991). This is critical for evaluating local versus global effects on stratigraphy.

**Calibration of Input Parameters**

The purpose of the calibration step is to define the sedimentation system and the size of the model. It is necessary to adjust the character of the sediment supply, gradients of stability, erosion rates, compaction rates, crustal rigidities, and model dimensions to
reflect the conditions observed in each area. Many of the values for the sedimentation system do not vary by more than 20-50% and do not require adjustment from the default values. Calibration of input parameters involves:

1. digitizing in depth or time: (1) the present sediment surface, (2) key sequence boundaries between the present sediment surface and bottom horizon, (3) the bottom horizon of interest, and (4) the basement. These values must be converted to depth.

2. restore the paleobathymetry of an initial time surface to the configuration at the start of the model,

3. estimate a sediment influx history by calculating the cross-sectional area of intervals and noting the ratios of carbonate to siliciclastic sediment,

4. adjust the depositional model to match stratal geometries, carbonate volumes, erosion features, etc.,

5. calculate differential tectonic subsidence rates across the profile,

6. backstrip sections to obtain a tectonic subsidence history, and

7. select or adjust a eustatic sea-level history.

Depending on the objective, it may not be necessary to determine all of these parameters for each section. In some cases, it may be sufficient to assume that values or histories determined from other regions are valid; eustatic sea-level history or rate of carbonate production as an example. Of course, subsidence distribution and history, sediment-supply history, and the initial bathymetric-profile are unique to each region.

The siliciclastic traction-load and suspension-load ratios are adjusted to reflect the volume of distal mud draping the gravity-flow sediments versus the volume deposited as clinoforms. The dispersion distance for the suspension-load in marine environment is adjusted to reflect the distance suspended sediments are found from the delta mouth. The stable gradients for the depositional front are measured from clinoform geometries on seismic or geologic cross-sections. Input gradients are always greater than measured
because compaction reduces the inclination with increased burial. Fluvial gradients are measured from local fluvial gradients. The coastal-plain gradient is usually very close to zero. The erosion rates are adjusted to reflect surface erosion and shoreface erosion rates by comparing the results with the stratigraphic cross-section. Channel depths and margins are adjusted to reflect channel geometries. Standard initial porosities and compaction rates are assigned for each lithology specified. The initial porosity and compaction rates are not modified during a simulation. The effective elastic thickness of the lithosphere is adjusted to reflect the width of the flexural response of sediment loading. This is measured by comparing the extent and slope of onlapping wedges in the results with that observed in the data. If the pinchout of strata is landward of the observed pinchout, the effective thickness of the lithosphere may be too great. The cell spacing defines the distance between cells. A model with a shorter cell spacing has a higher horizontal resolution. A longer cell spacing can be used with a larger sediment supply. However, with a short cell spacing, increased resolution is gained at the expense of information on the model boundaries. The best resolution is produced by a model that has spacing of 50-800 meters (traces on standard seismic data have a spacing of 25 meters). A rough model can be run with a 1000-2000 meter spacing.

A backstripped subsidence-analysis is performed to obtain a first approximation of the tectonic subsidence history within the region. This history will require adjustments to account for differences between one- and two-dimensional data.

**Simulation**

The simulation procedure involves adjusting the fundamental geologic controls of tectonic, sediment supply, and eustatic history until the model successively converges toward a cross section that matches the observations. This is done with the aid of a spreadsheet or by automated routines that use digitized measurements. The most important criteria to match are the total subsidence through time at each locality and
migration history of the offlap break. The following variables are adjusted to match the corresponding observations (Fig. 15):

<table>
<thead>
<tr>
<th>Variable</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tectonic Subsidence Rates</td>
<td>Total Depth</td>
</tr>
<tr>
<td>Sediment supply</td>
<td>Horizontal Migration History</td>
</tr>
<tr>
<td>Carbonate production rates</td>
<td>Volume of carbonate rock</td>
</tr>
<tr>
<td>Eustasy</td>
<td>Onlap, truncation, accommodation, cyclicity</td>
</tr>
</tbody>
</table>

The sediment supply influences the total subsidence history through flexure loading. If the sediment supply is decreased, the total subsidence will be less. Therefore, if the total subsidence was correct, tectonic subsidence will have to be increased to compensate for the loss of subsidence due to the decreased sediment-load.

**Initial Conditions**

To define the initial bathymetric conditions, use a time-line restored according to the best estimate of paleobathymetry, or a typical topographic and bathymetric profile from the region of study. It is important to match the initial onlap, or base-level, on the initial bathymetry. The position is especially sensitive when modeling carbonate rocks, as a decrease in the initial bathymetry will allow the carbonate factory to produce more rock material, as well as require less space to fill. All spatial data are referenced from the

![Figure 15. Four main variables controlling stratal patterns. Sediment supply controls the rate of progradation. Tectonism creates or destroys accommodation space. Climate controls the lithology. Eustasy controls the stratigraphic packaging.](image-url)
present-day geoid, or present mean sea-level. Therefore, if long-term sea-level is
different, it is necessary to raise or lower the initial bathymetric profile to place the initial
onlap in the proper location.

With minimal deformation, the initial bathymetric profile (Fig. 16) is determined
by projecting a reference horizon (the top of a clinoform) basinward to approximate sea-
level. The "compacted" paleobathymetry is approximately the difference between the
bottom horizon and the reference horizon. Backstripping an interval at this depth will
provide a decompacted thickness for a true paleobathymetry.

If the bottom horizon cannot be reasonably restored, it is best to choose a typical
bathymetric profile for the initial bathymetry. Shift the typical bathymetric profile
vertically to align the coastline parallel with the sea-level for the initial time setting. Shift
the profile horizontally within the model reference frame so the coastline lines up with a
known coastline position.

![Calculation of Paleobathymetry with a Reference Horizon](image)

![Calculation of Paleobathymetry from a typical bathymetric profile](image)

Figure 16. Determination of paleobathymetry for the initial time setting.
**Tectonic history**

Subsidence rates are used within the model to represent several sources of subsidence, including flexural loading from loads outside the model, thermal subsidence or uplift, compaction of sediments below the modeled layers, faulting, and diapirs. These rates should represent the motion of the basal surface that is not accounted for by flexural loading of sediment. Backstripping analysis can provide a first approximation of the tectonic subsidence or uplift history, sediment accumulation history and compaction state of a locality.

PHIL iteratively arrives at the subsidence history of a region by comparing the depths of horizons with those produced by the model and adjusts the tectonic subsidence rates to match the observations.

**Eustatic history**

The eustatic sea-level history is responsible for generating most of the detailed variations in the distribution of lithofacies within stratigraphic cross-sections. We have found through simulation of many stable margins that a sea-level-curve modified from Haq et al., (1987) correctly reproduces the "third-order" stratigraphic record. We have adjusted the long-term component as well as added new cycles to the short-term component of the curve. Overall, magnitude and timing of sea-level falls on the eustatic curve produces geometries within the limit of measurement. On the basis of expanded sections in the Pliocene and Pleistocene, additional cycles are required. During some intervals, the long-term component varies more radically than the model predicts. Continued refinement of all parts of the curve is possible with the help of the simulation tool. In different settings with very detailed stratal information, the third-order sea-level curve is adjusted to reflect the magnitude and shape of stratal geometries. Varying eustasy reproduces the regionally correlatable cyclic changes in base-level.
Sediment-supply history

Siliciclastic sediment-supply is best derived by measuring the volume of sediment deposited between sequence boundaries. If this level of detail is not available, it is necessary to match the position of the offlap break through time within the simulation.

The carbonate sediment-supply is adjusted with the shelf and shelf-margin carbonate production rates. We prefer not to change these constants through time unless data suggests the paleogeographic setting has changed. If siliciclastic sediment invades the region, the carbonate factory will automatically reduce its output. Drowning of carbonate banks will occur if the productivity of a locality cannot keep up with a sea-level rise (Schlager, 1989). In the model, drowning occurs when the rate of sea-level rise is greater than the combined shelf and shelf-margin production rates at the platform margin.

DISCUSSION

In this section, simulation results are compared with the stratigraphic record, the sequence stratigraphic paradigm, and other stratigraphic models to highlight some of the knowledge we have gained from the model. We consider the uniqueness of each solution as well as discuss some problems and limitations of the 2-dimensional model.

Comparison with the Qualitative Model of Posamentier and Vail (1988)

Depositional sequences and systems tracts proposed by Posamentier and Vail (1988), including lowstand, transgressive, highstand and shelf-margin systems tracts and their respective stratigraphic boundaries develop directly from the model. However, a problem remains within the lowstand systems tract in defining the mechanisms that separate gravity-flow sedimentation into two distinct units of basin-floor fan and slope fan (channel-overbank) as defined. We have found the concept of a stream equilibrium-gradient applicable to define fluvial base-level. However, it does not operate as a fixed concave-upward surface that migrates with the equilibrium point (position on the basin margin that bounds the region of relative rise from relative fall), as Posamentier and Vail
(1988) illustrated. The equilibrium-gradient is a dynamic surface that changes according to the bathymetry and sediment supply. As the shoreline migrates basinward, the gradient of the fluvial plain decreases because it is limited landward by the first substrate that is resistant to erosion. It cannot build up above that surface unless there is a surface upstream that is higher than a straight line between the two points.

The equilibrium point is not an important feature controlling stratal patterns. In complicated deforming regions, there may be several equilibrium points, with each moving in different directions. The concept that the bayline migrates with the equilibrium point is not viable. The bayline is the intersection between sea-level and the depositional interface. During a sea-level rise, the bayline is a function of antecedent topography and coastal sedimentation processes. In a sea-level rise, the bayline may be located at the head of a canyon cut by the previous fall. If the coastline is not near a filling canyon, the coastline may erode the antecedent topography, so the bayline and coastline are equivalent. During a stillstand, the bayline may lie inward of the coastline if fluvial processes dominate over marine processes; otherwise they may be equal.

Posamentier and Vail (1988) state that sequence boundaries will correlate to the point in time where the rate of eustatic fall exceeds the rate of tectonic subsidence, before the inflection point. Modeling Pleistocene sedimentation, we have observed that a Type 1 sequence boundary will form when relative sea-level is falling faster than approximately -30 mm/ky and base-level is below the offlap break of the maximum highstand-surface. If the bathymetric contrast is great enough in the delta front, this geometry will force deposition of the delta-mouth bar sands on unstable pro-delta silt and mud. This setting will cause slumping and gravity-flow transport into the deep basin. This relationship will frequently occur before a sinusoidal sea level curve reaches the inflection point. With slower subsidence rates, the sequence boundary is likely to occur earlier in the eustatic cycle.
It is observed that during relative falls of sea-level sediment bypasses the shelf and slope, and is deposited as basin-floor fans or as channel-overbank fans in deep water (Damuth et al., 1988; Kolla and Macurda, 1988; Manley and Flood, 1988; Bouma et al., 1989; Weimer, 1990). Earlier work (Jervey, 1988) defined the time of gravity-flow sedimentation when the relative sea-level is negative (when the rate of sea-level fall is greater than the rate of subsidence). If realistic rates of tectonic subsidence are used, this condition would not produce as much toplap as has been observed. In addition, some deltas on passive margins and foreland basins exhibit downstepping progradation (Tesson et al., 1990). Downstepping progradation requires coastal sedimentation during a relative fall. In nature, gravity-flow of coarse-grained material may be controlled by the stability conditions at the locus of deposition. Gravity-flow of coarse-grained material will occur when the delta-mouth bar is built on top of unstable, porous shales (perhaps over pressured) perched at their angle of repose. These conditions produce deposition of denser sand material on top of less stable shales in the depositional front of the previous sequence. This geometry is likely to produce slumping of coarse-grained sediments and formation of bottomset turbidites (shingled turbidites) at the base of depositional front when topographic relief is shallow (approximately less than 200 meters).

A relative fall exists under different eustatic and tectonic conditions. The rate of fall is dependent on the total subsidence rate, and the period and amplitude of the eustatic cycle. We have found that a model allowing sediments to prograde under a small rate of relative-sea-level-fall (less than -30 mm/ky) reproduces the development of forced regressions (called perched lowstands in Posamentier and Vail, 1988), and the observed toplap in the lowstand prograding complex. The rate of fall of relative sea-level is not dependent upon the time increment used, but produces an appropriate response in many varied settings.

The nomogram in Figure 17 illustrates the required subsidence rates and the period and amplitude of sinusoidal eustatic cycles necessary to induce a rate of fall of
relative sea-level that is less than -30 mm/ky. The amplitude also must be great enough to produce a fall below the offlap break of the previous highstand. Geologic settings with subsidence rates below the line representing a given periodicity will require eustatic cycle amplitudes greater than the amplitude determined by the intersection of the line with local subsidence rates. Therefore, gravity-flow sedimentation is more likely to occur with a cycle of shorter period, because the rate of fall is faster on a shorter period cycle with the same amplitude. Likewise, long period cycles require higher amplitudes in order to induce incision.

![Figure 17](chart.png)

Figure 17. Chart showing the relationship between magnitude and period of a eustatic cycle, and subsidence rates that are necessary to cause gravity-flow sedimentation with a relative fall threshold rate -30 mm/ky. Each line represents the minimum condition necessary for a eustatic cycle of a given period before bypassing will occur during the relative fall. Settings with subsidence rates and eustatic amplitudes below each line will produce bypassing conditions. Likewise, conditions above the each line will produce stable progradation.

In general, the amplitudes of eustatic cycles increase with period or duration of the cycle (Fig. 18). Higher frequency cycles need less amplitude to induce gravity-flow
sedimentation. A short-period cycle may fall at a high rate, yet it may not fall long enough for the coastal plain to reach the previous offlap break. Likewise, cyclicity with periods of 6 million years or longer is less likely to produce gravity-flow sedimentation because subsidence rates typically are faster than the associated rate of relative fall. Sea-level falls with amplitudes greater than 100 meters are common in the Pleistocene, Pennsylvanian and Permian, yet are rare throughout most of the geologic record.

![Figure 18. Spectral analysis of the Mesozoic-Cenozoic sea level history modified from the Haq et al. (1987) curve.](image)

Although the sea-level cycle is a composite of cycles of different period, one period typically dominates during each geological period (Fig. 19). The dominant-cycle periodicities range from third-order (0.5 - 3.0 million years) to fourth-order cycles (0.1 - 0.5 million years). When third-order cyclicity dominates, fourth-order cycles are usually present, yet are typically an order of magnitude lower in amplitude. This contrast may be due to periods of marine-based versus land-based glaciers (Bartek et al., 1991). Marine-based glaciers respond quickly to a warming event and sea-level rise. Therefore, if conditions are optimal for marine-based glaciers to develop, the eustatic changes may respond to higher-frequency, lower-magnitude temperature variations. When fourth-order cycles dominate, the amplitudes of eustatic cycles are higher as well.

The question remains how gravity-flow sedimentation naturally divides itself into two types -- basin-floor fan and slope-fan. Perhaps in the early phase of a relative fall the supply is sand-dominant as shoreface sands flow down the depositional front and the
fine-grained sediments are separated in the turbulent flow. Later, the supply may be mud-dominant due to additional mud removed from the incising submarine canyon.

Figure 19. The dominant period of "third-order" eustatic cycles through time. Determined by Fourier analysis of the Haq et al. (1987) curve over 5 My windows with periodicities less than 1.5 My and visual inspection for longer periodicities.

Posamentier and Vail state maximum flooding surface will correlate to a point after the inflection point of the eustatic rise, due to the slope of the onlap surface. A maximum flooding surface will form when the rate of relative sea-level rise decreases to a rate where space is created at the same rate as the sediment supply can fill the space. With a comparatively small sediment supply or large eustatic variation, this may occur considerably after a sinusoidal cycle has passed the inflection point. In this case, there may be regions that are not receiving sediment at the rate that rising sea-level is creating accommodation space. These localities will continue to record greater water depths above the maximum flooding surface defined by coastal onlap or an inflection point. In a setting with a high rate of sediment input, sea-level may never rise fast enough to create more space than the supply can fill. In this case, the stratal components will continue to
prograde into the basin. However, the surface of maximum rate of accommodation change can still be determined from the stratal patterns.

The bimodal nature of gravity-flow sedimentation remains unresolved. PHIL splits the turbidite system into two different systems that occur independently: 1) basin floor fan and associated slumps and 2) slope fan. Formation of submarine canyons, as suggested by Dingus and Galloway (1988), during a sea-level rise and flushing of fluvial valleys by during deglaciation could not be reproduced with the model. This condition would produce a package of gravity-flow sediments in the deep basin that is correlatable with large volumes of sediment on the shelf. This package would in turn be bound by condensed sections or maximum flooding surfaces. I was unable to establish a mechanism that could produce gravity-flow sedimentation concomitant with major increases of accommodation and sediment supply on the shelf nor find evidence of such processes. If gravity-flow sediments were associated with high sediment volume, it would be expected that high-flux deltas such as the Mississippi or Niger delta would continuously form turbidites.

Uniqueness of Solution

The exercise of building a simulator is also instructive in the uniqueness of each response. Subtle changes in the input parameters or an algorithm are usually apparent in the results. For a simple margin without differential subsidence (this may be an unusual condition), Kendall et al. (1989) has argued that it is possible to produce the accommodation changes in two ways. First, by holding sea-level constant, calculating the rate of sea-level change, and adding it to the tectonic-subsidence rates. However, the simplest way to simulate these stratal patterns is with (1) a slowly changing tectonic-subsidence history, (2) flexure loading, (3) a high-frequency eustatic record, and (4) a sediment-supply history. For a complex margin with rotating blocks, it is extremely difficult to model a subsidence and eustatic history combined as subsidence. In addition, the rates and variations of tectonic subsidence necessary to reproduce the relative sea-level
changes are difficult to explain solely with a tectonic mechanism. For instance, third-order variations in sea level are observed in stratigraphy deposited on epicontinental seaways during the first-order highstand in the Cretaceous. It is unreasonable to expect that stable interior continental crust should experience "third-order" tectonic changes only when it is buried by a water column. Documentation of sequences and systems tracts with global biostratigraphic correlation and similarity of their duration, shape, and total magnitude will test the eustatic nature of the regional base-level changes.

The ultimate foundation of all stratigraphic modeling is the geologic time scale. Comparing the time scales (Haq et al., 1987; Harland et al., 1989) indicates that interval duration's may differ by as much as 50-200 percent. These variations do not effect the stratal geometries but require that all time-dependent variables are adjusted equally.

PHIL generates stratigraphic packages of equal duration, whereas the geologic record is a continuum of correlatable surfaces forming at time increments that range from hundreds, to hundreds of thousands of years. The time increments represented in the stratal patterns in seismic data or well logs are dependent on the data resolution. Their resolution in turn depends on the compaction state of the sediments, sedimentation rates at the time of deposition, and seismic data resolution. Five to one-hundred Hertz seismic data used by the oil industry can resolve reflections from third- (0.8-1.5 My) and fourth-order (100-400 ky) stratigraphic time lines (Vail et al., 1977b). Shallow high-resolution seismic data, acquired to study recent deposition, resolve time lines that may be as short as 10 years. Well logs contain events that range from 1 My to 20 ky in most data, but as short as 1 year in regions with high sediment accumulation rates (Van Wagoner et al., 1990).

CONCLUSIONS

PHIL simulates a stratigraphic cross-section with an erosional and depositional model that is sensitive to the character of the bathymetric profile. The erosional system is represented by independent algorithms that simulate shoreface erosion, channel incision,
surface beveling and redistribution by marine currents. The depositional system is simulated by a prograding clinoform, filled by externally-sourced siliciclastic sediments and water-depth-dependent carbonate production that is damped by the siliciclastic sediments. The siliciclastic system is represented by traction transport and mixing of suspended sediment within the marine water column. Sediments slump and flow into deeper water if the sediment is deposited on a gradient that is unstable for its lithology.

Gravity-flow sedimentation is important during relative sea-level falls and in settings with deposition on resistant substrate dipping more than the stable gradient for each lithology. Gravity-flow sediments migrate downslope and are deposited as turbidites. The turbidites will deposit as channel-levee complexes and basin-floor fans in deep-basinal positions. These positions are controlled by the surface gradients over which the sediment is transported. Flat gradients decrease the momentum of the turbidite mass and cause sediment to deposit. We divide the gravity-flow sedimentation processes into two types, basin-floor fan and slope fan. The basin-floor fan is a massive sand-rich unit that is deposited during the early part of a Type 1 relative sea-level fall. This unit may be sourced by coarse-grained sediment that is flushed from the fluvial valleys and coastlines during the first part of the fall. The slope fan is a mud-rich unit that is deposited as the Type 1 relative fall causes deeper cutting into the older muddy units and the coeval prograding delta collapses.

By simulating dip-aligned sections, one can produce a comprehensive, quantitatively defined, reproducible model of the eustatic, sedimentation, and tectonic history within stratigraphic records along the margins of basins. When a good representation of a geologic cross-section has been achieved, the quantitative input provides a data-set from an integrated model that includes a depositional model and the effects of compaction, erosion, stability conditions, crustal response to sediment and water loading.
Chapter 3.
Response of Stratal Patterns and Lithofacies Distribution to
Sediment Supply, Tectonic, and Eustatic Variations

ABSTRACT

Eustatic cycles, tectonic events, sediment supply pulses, and their subsequent equilibration responses are the dominant factors controlling stratal geometries. The eustatic cycles, tectonic event and sediment supply pulses operate over duration's greater than ten thousand years with magnitudes and extent that form features that are measured and correlated in stratigraphic data. These events and cycles form a hierarchy of packages that compose the stratigraphic record and differ in onlapping relationships, stacking patterns and internal character. The hierarchy is as follows (Fig. 1):

1. continental-encroachment megasequence;
2. transgressive-regressive facies wedge and supersequence;
3. sequence -- Type I, Type II and incomplete;
4. systems tract -- lower and upper lowstand, shelf margin, transgressive, and highstand;
5. component group -- parasequence, slope fan lobe, etc.;
6. and stratal component.

Each package is composed of the subordinate packages.

Sediment supply, tectonism, and eustatic fluctuations produce these packages by changing and filling shelf accommodation space and basinal bathymetry. Each of these mechanisms operates with characteristic rates and patterns. The interactions of sediment supply with changes in accommodation space control the distribution of lithofacies and geometry of stratal surfaces. The change in accommodation space near the offlap break is the dominant control on the geometry of stratal surfaces and distribution of lithofacies within the deeper marine environment. In addition, the geometry of the substrate and
Figure 1. Hierarchy of stratigraphic packages produced by eustatic, tectonic and sediment supply variations.
bathymetric changes due to compaction and flexural loading of sediment have a strong
influence on stratal geometry.

In this paper we present results of computer simulations designed to
independently model the response of these variables to identify their unique stratal
signatures. Computer simulations of different settings (passive margin, growth faults,
mixed carbonate and siliciclastic sediment supplies, carbonate ramp, steep carbonate
platform margin, and others) demonstrate how siliciclastic and carbonate depositional
systems interact with the bathymetric conditions produced by the variables. This
computer simulation includes algorithms that model flexure loading (compensating for
sediment and water loads), compaction, deposition of traction load, dispersion of
suspended load, gravity flow sedimentation, carbonate production and redistribution,
and erosion. This method is sufficiently flexible to allow tests of most stratigraphic and
tectonic settings, as well as simulation of most dip-aligned stratigraphic cross-sections.

Stratal patterns record the interaction of sediment supply with relative changes of
sea level. The sum of tectonic subsidence (thermal subsidence or flexure loading of
thrusts) and the flexural response of the water column and sediment loading minus
compaction equals total subsidence. Relative change of sea level is the sum of total
subsidence and eustasy. Sediment supply, relative changes of sea level, and local
sedimentation conditions control base-level for any locality.

Sediment supply controls the amount of space that is filled and the lithology of
the fill. A change in sediment supply can produce a transgression and regression of
facies, or perhaps a change in the lithofacies delivered to a location. However, changes
in sediment supply cannot produce a downward shift of onlap. This type of
transgression and regression is characterized by minimal subsidence rates during the
transgression (due to minimum loading) and maximum subsidence rates during the
regression (due to maximum loading).
Tectonic and eustatic changes are recorded as changes in stratal thickness, erosion patterns and paleo-water depth. Both factors can produce a downward shift of onlap. Changes in tectonic subsidence are often simultaneous throughout a region but vary both in direction and magnitude. Most tectonic changes are irreversible, whereas eustasy is cyclic. For example, when compressional deformation reactivates extensional structures, the deformation inverts the layers within the down thrown blocks rather than reversing the subsidence at the boundary of the block. The writers have not observed short term cyclic changes (less than 2 My) that are "tectonic" in character.

INTRODUCTION

In this paper, we present computer simulations from a computer program (PHIL) that models stratigraphic variations due to changes in initial bathymetry, sediment supply, tectonism, and eustasy to demonstrate how these factors influence the stratigraphic record. The results demonstrate that the sequence geometries and lithofacies distribution change dramatically depending upon tectonic and sediment supply conditions.

Every geologic variable has a unique signature commonly recognizable in a stratigraphic section. Unfortunately, the scale of the system is much larger than outcrops or more details are necessary than are available from seismic data to construct a complete record. The simulation produces controlled experiments that isolate single variables, and allow us to compare their impact with isolated geological observations. This has allowed us to present a comprehensive organization of the stratigraphic record. This paper documents the relative importance and effect of each variable by independently changing it to observe the response in the stratal patterns. The effects of these variables and the values used in the sedimentation model influence changes in the distribution of lithofacies and shelfal accommodation space, geometry of stratal surfaces (lines in two-dimensional space) and stratal terminations (an appendix contains a
glossary of terms and their usage). Ultimately, analyses of models resulting from combinations of these effects demonstrate how to separate the individual signals.

Tectonic and climatic activity creates cyclicity in the stratigraphic record (Vail et al., 1977; Perlmutter and Matthews, 1989; Cloetingh, 1988). Tectonic activity is the result of stresses generated by the interaction of lithospheric plates or density-driven diapirs and faults. Variations in the intensity of solar energy reaching the earth's surface influence its climate, e.g. its temperature and precipitation distribution. These factors affect the volume of water stored as ice, and erosion and flow rates of fluvial systems.

An objective sequence-stratigraphic analysis requires separating the interpretation into three fundamental elements: 1) a description of the stratal geometries and lithofacies distribution, 2) establishment of a chronostratigraphic framework, and 3) a model of the depositional processes that are active in each component. For instance, a stratigraphic package is defined as a sequence if it is bounded by surfaces characterized by starvation, sediment influx, coastal onlap or erosion, or surfaces that correlate with documented sequence boundaries. This sequence identification is independent of the chronostratigraphy. Ultimately, a cause of stratigraphic variations can be interpreted after completion of a subsidence analysis and simulation. From simulation of stratigraphic information, we interpret the character of the depositional system, sediment supply history, subsidence rates, eustasy, paleobathymetry, slopes of stability, and erosion rates. The relative importance and specific effects of these factors are not known.

The model shows that high subsidence rates found in active foreland basins and high-flux depocenters along passive margins containing growth faults reduce the magnitude of the relative fall of sea level. High subsidence rate may eliminate a relative fall of sea level altogether. Times of relative fall and gravity flow sedimentation on a passive margin may correlate in a foreland basin with times of relative fall accompanied by an increased rate of progradation. Where bathymetric contrasts are small
(approximately less than 200 meters), clinoform deposition is stable and preserves
down-stepping progradation, forming the perched lowstands of Posamentier and Vail
(1988).

SEDIMENTATION ALGORITHM

The sedimentation model includes siliciclastic and carbonate deposition. The
model is two dimensional and accounts for compaction and flexural loading. Bowman
and Vail, (in review), have described the algorithms used by the model and the typical
values for geologic variables based on modeling siliciclastic and carbonate settings for
passive margins, foreland basins, and rifts. The present paper focuses on the
stratigraphic response to variations in these geologic variables, principally changes in
sediment supply, eustasy, and subsidence.

Computer simulations of depositional processes generate detailed stratigraphic
cross sections that closely approximate geologic time-lines and lithofacies distributions
as determined from exploration data (seismic sections, well logs, lithologic and
paleontologic control). PHIL (Bowman and Vail, 1994a) is sufficiently flexible to
simulate most stratigraphic cross sections and test many stratigraphic hypotheses. The
simulation includes linked depositional algorithms that model transport of siliciclastic
traction load, dispersion of suspended load, gravity flow transport and sedimentation,
production and redistribution of carbonate sediment, and erosion. These algorithms
respond to bathymetric conditions. Variations within different basin types or tectonic
settings result from the distribution of faulting or subsidence history with respect to the
sediment source. The user defines the tectonic history, perhaps with the aid of
subsidence analysis. The simulator calculates the flexural response to sediment and
water loading, and sediment compaction. The simulator works in passive margin, rift,
foreland basin, and strike slip basin settings.
STRATIGRAPHIC SEQUENCES AND THEIR RELATIONSHIP TO VARIATIONS IN SEDIMENT SUPPLY, TECTONISM, AND EUSTASY

A hierarchy of stratal components, sequences and groups of sequences forms the stratigraphic record. Tectonic, eustatic, and sediment supply cycles generate this hierarchy of packages that can be identified by their onlapping relationships, stacking patterns, and internal character. The hierarchy is as follows:

(1) continental-encroachment megasequence;
(2) transgressive-regressive facies wedge and supersequence;
(3) sequence -- Type I, Type II and incomplete;
(4) systems tracts -- lower and upper lowstand, shelf margin, transgressive, highstand;
(5) component group -- parasequence, genetic sequence, etc.; and
(6) stratal components -- backstepping, upstepping, forestepping, downstepping, and gravity flow components.

Each package is composed of the subordinate packages.

In this hierarchy each cycle has a characteristic duration, pattern and scale (Vail et al., 1991). The resultant depositional patterns are dependent upon local subsidence rates and sediment supply conditions. Cycles of eustasy and sediment supply driven by orbital forcing (10-500 ky) are the building blocks of sequence cycles. Sequence cycles (100-3000 ky) are the building blocks of transgressive-regressive facies-cycles. Transgressive-regressive facies-cycles (3-70 My) are the building blocks of continental encroachment cycles.

Very long term changes (periods greater than 50 My) in sea level (tectono-eustasy) induce continental encroachment megasequences. Changes in sediment supply and tectonic activity induce transgressive-regressive facies-cycle wedges (White, 1980). All frequencies of eustatic cycles build independently but in combination with transgressive-regressive facies-cycles.
The stratigraphic signatures of changes in the rate and type of sediment supply and long term (>3 My) transgressions and regressions are apparent in the stratal patterns and lithofacies. The stratigraphic signature of tectonic activity is tilting and rapidly changing subsidence rates often associated with local movement around faults.

The stratigraphic signature of eustasy is a cyclic pattern of very rapid changes in accommodation, up to 25000 mm/ky (Fairbanks, 1989), often triggering erosion and gravity flow sedimentation in deep water (Damuth et al., 1988; Kolla and Macurda, 1988; Manley and Flood, 1988; Bouma et al., 1989; Weimer, 1990).

Continental encroachment cycles are driven by variations of spreading rates and the total length of spreading ridges (Vail et al., 1991). Continental encroachment megasequences are characterized by continental exposure and restricted deposition on the margins of the continents, followed by flooding of the continents. They are not necessarily tied to the long term position of the coast line because onlap may be accompanied by progradation of the coast line. Examples include long-term eustatic rises associated with the breakup of Pangea in the Triassic and the breakup of Pan Africa in the late Proterozoic, and the associated long term sea level maximums in the Turonian and early Ordovician.

Transgressive-regressive facies-superequences are defined by the long term position of the shoreline (Table 1). They may be caused by either tectono-eustatic sea level changes or basin-scale tectonic events. Typical transgressive-regressive facies-cycles last 3-70 million years. Second-order eustatic supersequences are composed of a set of sequences. These sequences form a series that prograde, backstep, and prograde. Major tectonic events, such as rifting or thrusting, often initiate basin-scale transgressive-regressive facies-superequences. On continental margins, strata commonly transgress over a hiatus above an angular unconformity. Transgression consists of backstepping sequences followed by major aggradational stacking of lithofacies, landward translation of the coastline, and deepening paleobathymetry.
Table 1. Summary of transgressive and regressive signatures caused by tectonic events, sediment supply and eustatic cycles.

<table>
<thead>
<tr>
<th>Regression</th>
<th>Sediment Supply Cycle</th>
<th>Tectonic Event</th>
<th>Type 2 Eustatic Cycle</th>
<th>Type 1 Eustatic Cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sigmoidal Offlap</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Aggradational Offlap</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Oblique offlap/Toplap</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Downlap</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Base-level fall with erosion</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Lowstand deposits</td>
<td>Yes</td>
<td>Minor</td>
<td>Yes</td>
<td></td>
</tr>
<tr>
<td>Slumping</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Gravity-flow transport</td>
<td>Minor</td>
<td>Yes</td>
<td>Yes</td>
<td></td>
</tr>
<tr>
<td>Uplift</td>
<td>Major</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Angular discontinuities</td>
<td>Major</td>
<td></td>
<td>Minor</td>
<td></td>
</tr>
<tr>
<td>Transgression</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ravinement</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Apparent Truncation</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Stratal patterns</td>
<td>Divergent</td>
<td>Divergent</td>
<td>Parallel</td>
<td>Parallel</td>
</tr>
<tr>
<td>Onlap</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Globally correlative</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnitude and timing</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Marine condensed section</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>correlates with MFS</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Non-marine condensed</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>section correlates with MFS</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum accommodation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>rate correlates with MFS</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
</tbody>
</table>

Backstripping analyses show a rapid increase in the rate of tectonic subsidence, lasting 3-5 My during a major aggradational phase. The increased rates of tectonic subsidence may decline over a period of 50-70 My above a thermally altered crust. The associated regression may progress gradually in an extensional basin or initiate rapidly in a foreland basin as thrusting ceases.
Comparison with a eustatic cycle chart and a backstripping analysis may indicate whether the transgressive-regressive facies-supersequence is caused by:

1) long term tectono-eustasy,

2) local sediment supply changes, or

3) basin-scale tectonism.

Long-term tectono-eustatic changes are globally equivalent in magnitude and timing and typically lower in magnitude than relative changes produced by tectonic events. Tectonic events and changes in sediment supply occur at different times globally and have highly variable magnitudes -- even within the same basin.

A Type I eustatic sequence cycle (periodicity of 0.5 - 3 Ma) produces a sequence that contains four distinct packages defined as the lower lowstand, upper lowstand, transgressive, and highstand systems tracts. Third order Type II eustatic sequences contain shelf margin, transgressive, and highstand systems tracts (Posamentier and Vail, 1988). Each systems tract forms in response to local changes in accommodation and sediment supply. Each systems tract contains unique lithofacies distributions and locations of sub-aerial erosion and ravinement surfaces. In many tectonic and sediment supply conditions, a sequence that records a full suite of systems tracts elsewhere forms an incomplete sequence with one or more of the systems tracts missing.

**Model Parameters**

The model used throughout this analysis applies corrections for the crustal response to flexural loading of water and sediment, compaction, and erosion. In the model, the effective elastic thickness of the crust is set to 11 km. The siliciclastic sediment supply is 2000 m²/ky, which is low compared with many margins. The tectonic subsidence rates are constant at each point through time. They increase from 0 mm/ky on the landward margin, to 40 mm/ky on the seaward margin. Each model records time lines (stratal geometries) every 100 ky. Each time step is preceded by a lag period followed by a pulse of sediment. Between each depositional pulse, the program
compacts sediment, adjusts sea level and tectonic subsidence, calculates the subsidence
due to flexure loading of sediment, and records the resulting paleo-water depths at the
beginning of the stratal component. All models have a constant pelagic sediment supply
of 10 mm/ky. Drape of pelagic and suspended sediment produce continuous, parallel
stratal surfaces that mimic the antecedent topography.

Models isolating the three controlling factors show that tectonic and eustatic
variations change base-level, whereas sediment supply controls the amount of space
filled or the rate of progradation (Table 1). A transgression due to an increase in
subsidence rates correlates with an increase in accommodation rate that is often unequal
across the coastal and fluvial plains. This variation across the basin is recorded by
divergent stratal patterns. A transgression due to a decrease in sediment supply
correlates with a decrease in accommodation rates in the coastal plain. A transgression
due to eustatic rise correlates with an equal base-level change that produces parallel
stratal patterns within the coastal and fluvial plain. Equal base-level changes enhance the
preservation of onlap.

The progradation-aggradation ratio (PA) determines gross basin fill geometry;

\[ PA = \frac{\text{rate of accommodation space creation}}{\text{rate of sediment supply}} \]  

(1)

where;
PA > 1: rate of space creation > rate of sediment supply = transgression
PA = 1: rate of space creation = rate of sediment supply = vertical aggradation
PA < 1: rate of space creation < rate of sediment supply = regression

Space is created and destroyed by tectonism, flexural loading, and eustasy. Compaction
creates space. Changes in rates of sediment supply are a result of tectonic uplifts or
climatic variations that create proximal sources or a major diversion of streams into or
out of the region.
Sediment Supply Variations

Sediment supply, the rate of siliciclastic influx or in-situ biogenic carbonate productivity, controls the rate of progradation and indirectly the rate of aggradation due to changes in flexure loading of sediment. A 10 My model with a 7.5 My sediment supply cycle that varies from 2000 to 0 m²/ky, produced the stratal patterns in Figure 2. In this model tectonic subsidence rates are constant and eustasy does not change. All relative rise is due to tectonic subsidence and flexural loading of the sediment. We assume the depositional processes within deltas under high rates of sediment influx are similar to those under low rates. This is a valid assumption if we compare modern deltas and note that deltas are capable of systematically prograding without causing gravity flow sedimentation of the coarser fraction (except for special bathymetric settings). However, perhaps a rapid change in sediment flux may cause instabilities at the offlap break. Above 30,000 m²/ky growth faulting may be active due to overpressuring. A divergent stratal pattern characterizes a sediment supply cycle. Landward merging of the components cause a gradual thinning resulting in a non-marine condensed section. The non-marine condensed section is due to a decrease in accommodation rates. The onlap is confined to the region near the hingeline by landward decrease of tectonic subsidence and associated landward decrease of flexural response to sediment loading. Erosional unconformities do not develop because base-level does not fall. Sediment starvation during low rates of sediment supply creates a marine condensed section in the basin. The regressions terminate basinward by downlap. Toplap does not form during any part of the cycle.

During a transgression induced by a decrease in sediment supply, aggradation rates decrease as a result of a decrease in the flexural response to sediment loading (Fig. 2). The transgression is characterized by thin strata and a condensed section in the basin. Stratal components condense into a thin non-marine interval landward of the offlap break. Stratal components terminate basinward by apparent truncation.
Figure 2. Stratal pattern produced by a sediment supply cycle. Maximum regression occurs during the maximum rate of sediment flux and is associated with maximum rates of creation of accommodation space in the coastal plain. Peak transgression occurs during the minimum rate of sediment flux and is associated with minimum rates of creation of accommodation space in the coastal plain and a condensed section in the basin. Stratal patterns are parallel or downlap onto condensed section. Onlap does not form except initially while the depositional profile is equilibrating with the initial bathymetric profile.

Shoreface erosion produces a ravinement surface with each rise but does not cause a significant amount of sediment to collect in a region that can slump into deep water. Storm currents carry coarse-grained material shoreward and carry fine-grained material further offshore. Slumps are much more prevalent during periods of high rates of sediment supply and conditions with large bathymetric contrasts.

Galloway (1990) documented direct correspondence between subsidence rates and sediment supply in the northwestern portion of the Gulf of Mexico basin. He shows that episodes of high rates of sediment supply correspond with high rates of subsidence and accommodation on the shelf. All subsidence rates increase basinward and reach a maximum beyond the extent of his data. This indicates most sedimentation occurs in the deep basin most likely by gravity flow transport. The flexural response to sediment loading in the deep basin is the dominant source of subsidence on the shelf.
The transgressive-regressive facies-cycle wedges described by Galloway are controlled by sediment supply due to tectonic uplift in the interior of the North America.

**Tectonic Variations**

Tectonism includes both active processes and passive responses. Active processes, such as faulting and volcanism, tend to be short term (< 5 My), are regional in scale, and are an important cause of rapid changes in subsidence or uplift (Bowman, 1985). Active processes are commonly associated with plate-tectonic reorganization or collisions. Active processes in extensional settings include uplift or subsidence associated with fault block rotation and isostatic adjustments of the crust and mantle. Active processes in compressional settings include major subsidence in the basin due to thrust loading, perhaps minor uplift associated with a flexural bulge, and major uplift above compressional structures. Passive responses occur as the crust returns to an equilibrium state after a tectonic disturbance and tend to be long-term (5-50 My) and of regional extent. Passive responses include subsidence due to crustal cooling after stretching, and flexure loading and unloading of sediment associated with crustal thickening or thinning.

Figures 3 and 4 illustrate active mechanisms and their associated passive responses that produce a transgressive-regressive facies-cycle wedge. The two tectonic mechanisms are: (1) exponentially decreasing subsidence rates on a rifted margin that occurs over 160 million years (Fig. 3A), and (2) a 1.5 million year tectonic event that interrupts a stable basin margin followed by a passive equilibration response to the event (Fig. 4).

A tectonic event produces a transgressive-regressive facies-cycle wedge that exhibits maximum subsidence rates during the transgression. Exponentially decreasing subsidence rates on a rifted margin may produce a transgressive-regressive supersequence when the initial uplift enhances the regression. The subsequent rapid subsidence rates cause a transgression (Fig. 3A). As subsidence rates decrease, the
sediment supply surpasses the rate of creation of accommodation space and the stratal path begins to prograde. If a long-term 250 meter eustatic cycle (Fig. 2B), such as has occurred in the past 200 million years (Haq et al., 1988), is added to the rifted margin, the transgressive-regressive megasequence is enhanced (Fig. 3C). This is similar to continental flooding megasequences observed on Atlantic margins (Vail et al., 1991) and to many second-order transgressive-regressive facies-supersequences.

The tectonic subsidence history of this simulated tectonic event has a period of uplift of 20 meters lasting 1 million years, during which tectonic uplift rates reach 200 mm/ky (Figs. 4 and 5), followed by a 1.5 million year period of rapid subsidence that exponentially decreases to zero. Unconformities form due to erosion of the coastal plain. Toplap at the locus of deposition records a regression associated with an uplift or a period of stability. If the uplift rate is greater than 30 mm/ky, the shelf margin collapses forming gravity flow deposits in the deep basin.

An increased rate of relative sea level rise during the subsidence phase causes a transgression of the shoreline and forms a condensed section within the deeper portion of the basin. The shoreline stacks vertically when the sediment supply is in equilibrium with the creation of new space. Onlap is well developed landward and apparent truncation basinward of the shoreline. Stratal lines within the coastal and fluvial plain are divergent if the rates of tectonic subsidence increase in a basinward direction. Regression follows the period of rapid tectonic subsidence and forms when the sediment supply is greater than the rate of space creation. Broad belts of meandering channels, and minor onlap characterize the fluvial depositional system during a regression.

As long as sediment and space below sea level exist, a basin subsides due to flexure loading of the sediment. After the crust stabilizes following a tectonic event, subsidence continues to occur in response to sediment loading as deposition fills the remaining water column. A location filled with sediment to base-level, therefore
continues to subside in response to loads deposited farther basinward within two or three times the flexural wavelength of the crust.

Figure 3. Transgressive-regressive patterns produced with the combination of a constant siliciclastic sediment supply of 2000 m2/ky, a tectonic event and eustatic cycle. A) Stratigraphic pattern of the short uplift followed by exponentially declining subsidence rates without the eustatic cycle does not produce continental flooding. B) Stratigraphic pattern of the 160 My 250 meter eustatic cycle illustrates 300-500 meters of continental flooding without the exponentially declining subsidence rates does not produce significant transgression. C) Stratigraphic pattern of a 160 My 250 meter eustatic cycle and exponentially declining subsidence rates produces transgression and a transgressive/regressive cycle.
Figure 4. Stratigraphic pattern produced by a 1.2 million year tectonic event. Maximum regression occurs during the minimum rate of subsidence or uplift (active phase) and is associated with minimum rates of accommodation in the coastal plain or erosion. The stratigraphic surfaces often terminate along a tectonically enhanced unconformity. Peak transgression occurs during the maximum rate of subsidence (passive response phase) and is associated with maximum rates of accommodation in the coastal plain and a condensed section in the basin.

Figure 5. Subsidence history for the model with a 1.2 million year active phase with an uplift of 20 meters, during which tectonic uplift rates reach 20 mm/ky. This is followed by 2 million years during which tectonic subsidence rates start at 75 mm/ky, then exponentially decline to zero.

Regional base-level changes caused by tectonism may be separated from eustatic changes by backstripping stratigraphic sections. Backstripping documents the timing
and magnitude of base-level changes. Second-order eustatic changes are synchronous with globally equal magnitudes, whereas changes in tectonic subsidence rates often form a step function with higher magnitudes that vary throughout a region. Comparing the timing of transgressions and regressions with a global sea level record (Haq et al., 1988) will help differentiate local from global changes. Large-scale tectonic activity such as plate reorganizations have a dominant but slowly changing effect on eustasy by changing the volume of the oceans (tectono-eustasy).

**Eustatic Variations**

An example in a siliciclastic shelf-margin setting (Fig. 6) illustrates the response of our sedimentation algorithm to short-term changes in eustasy. This 3.5 million-year-long model uses 100 ky time-steps. A constant, relatively low sediment flux (2000 m²/ky) feeds the depositional system where traction processes transport 70% of the sediment and suspension processes transport 30%. The fairweather wavebase or offlap break is 10 meters. The maximum gradient of stability is 0.065 (3 degrees) and the distance of injection of suspended sediment is 15 kilometers. The initial bathymetric profile (surface 0) represents a relatively shallow (maximum depth 300 meters) shelf margin that equilibrated with the sedimentation model during an earlier 1.5 My eustatic cycle of 80 meters. Tectonic subsidence rates are linearly interpolated from 0 mm/ky, on the landward side of the model, to 40 mm/ky on the basinward side of the model. Tectonic subsidence rates are constant through time at each location.

The two eustatic cycles in this example are similar to cycles documented in the stratigraphic record. These cycles have a periodicity of 1.7 and 1.4 million years and amplitudes of 80 and 25 meters, respectively. These cycles produce two complete sequences (surfaces 4-21, and 21-35). The first sequence forms in response to a rapid relative fall of sea level, and forms a Type I sequence boundary (surface 4) (Vail et al., 1977). The second sequence forms in response to a slow relative fall of sea level, and forms a Type II sequence boundary (surface 21).
Figure 6. The basic input to a siliciclastic model with two eustatic cycles of 80 meters and 25 meters respectively and 1.7 and 1.4 My in duration. Sediment supply is constant at 2000 sq. m./ky. Subsidence is constant through time but varies across the margin.
The interaction of eustasy and flexure loading of water on a basin margin produces subsidence that adds to the variation due to eustasy (Christie-Blick et al., 1987) by an amount that decreases away from the space that is available to be filled (Fig. 7). The magnitude of the subsidence depends on the rigidity of the crust and slopes on the bathymetric profile. The flexural response above crust with an elastic thickness of 10 km amplifies the relative change of sea level by approximately 10%.

![Figure 7](image)

Figure 7. Comparison of the flexural response to a highstand versus lowstand in sea level. Flexure loading and unloading of the crust due to the weight of the water column adds to the change due to eustasy by approximately 20% depending on the location on the margin.

A eustatically induced transgression is characterized by a progressive increase in stratal thickness, parallel stratal surfaces, and onlap throughout the rise. Transgressive strata terminate landward by onlap and basinward by apparent truncation. A relative change of sea level that is equal across the section produces parallel stratal surfaces. A condensed section develops in the basin during the transgressive and early highstand systems tracts. Increasing rates of relative rise of sea level, continued sediment loading, and an increase in water load due to the rise of sea level produces a progressive increase in stratal thickness during the transgression.
Type I Eustatic Cycle

Eustatic cycles that produce relative falls of sea level that fall below the offlap break of the previous highstand (Fig. 6) produce stratal configurations that form Type I sequences (Vail, 1987; Posamentier and Vail, 1988). Type I sequences occur in response to short-term cycles with high magnitudes (Bowman and Vail, 1994a). Type I eustatic changes produce moderate rates of progradation during stillstands and high rates of backstepping during the rise.

Type II Eustatic Cycle

Eustatic cycles that produce relative falls that remain above the offlap break of the previous highstand produce configurations that form Type II sequences (Vail, 1987; Posamentier and Vail, 1988). Type II sequences can occur over a range of conditions ranging from response to long-term cycles (>5 My) with higher magnitudes or short-term cycles (.02 to 5 My) with lower magnitudes (Bowman and Vail, 1994a). Eustatic cycles that are greater than 50 million years in duration form megasequences and those that are 10-50 million years form supersequences (Haq et al., 1988; Vail et al., 1991). Global changes in base-level characterize a eustatically induced regression, transgression, and regression. Type II eustatic changes produce high rates of progradation during stillstands, and minimal rates of backstepping during the rise.

Three Type II sequences are shown: (1) 160 My sequence produced by a 160 My cycle with a 250 meter amplitude sampled at 5000 ky time steps (Fig. 3B); (2) 10 My sequence produced by a 7.5 My cycle with an 80 meter amplitude (Fig. 8); and a (3) 1.5 My sequence produced by a 1.5 My cycle with a 25 meter amplitude (Fig. 6). An 80-meter sea level cycle with a period less than 6 My produces a Type I sequence with associated turbidite deposition (Fig. 9). However, a cycle with a 7.5 My period produces a relative stillstand because the rate of relative sea level fall is never fast enough for base-level to fall below the offlap break of the previous highstand (Fig. 8). Relative stillstand or small relative fall produces rapidly prograding packages bounded landward
Figure 8. Type II sequence produced by a 80 meter eustatic cycle with a 7.5 Million year period. Type II sequence boundaries form with an associated shelf-margin wedge. Stratal lines within the coastal plain are parallel, show onlap and maximum rates of accommodation at the maximum flooding surface. The relative change of sea level chart shows the timing of sequence boundaries and maximum flooding surfaces. rcs1 - relative change of sea level, e - eustasy, tect - tectonic subsidence, tot - total subsidence.
Figure 9. Type I sequence produced by a 80-meter eustatic cycle with a 1.5 Million year period. Type I sequence boundaries form with an associated basin-floor fan, slope fan, and lowstand prograding complex. Stratigraphic lines within the coastal plain are parallel, and show onlap and maximum rates of accommodation at the maximum flooding surface. The relative change of sea level chart shows the timing of sequence boundaries and maximum flooding surfaces. rcsL - relative change of sea level, e - eustasy, tect - tectonic subsidence, tot - total subsidence.
long-term (frequency greater than 50 My) or tectono-eustatic changes (Rona, 1973; Pitman, 1978; Kominz, 1984). The average age of oceanic crust determines the volume of ocean basins. The average age of the crust becomes younger during times of extensive spreading ridges and high rates of spreading. Younger crust is more buoyant and maintains a profile that is higher than old, dense, and cool crust, resulting in less oceanic basin volume and higher eustatic levels. These tectono-eustatic changes create continental encroachment cycles that change very slowly as the exponentially subsiding crust cools -- a process that can take 70 - 100 My.

SEQUENCE STRATIGRAPHIC MODEL

In this section we present a sequence stratigraphic model that is consistent with real data modeled to date. We demonstrate how packages of strata (beds, bed-sets) form stratal components bounded by deprivation and influx surfaces. Components occur in pairs to form component groups. These component groups are the building blocks of systems tracts. A sequence forms when a complete series of lowstand, transgressive and highstand systems tract forms. When several sequences occur in a transgressive-regressive pattern, they group into a supersequence. Likewise, when several supersequences form a transgressive-regressive pattern they group into a megasequence.

Stratal surfaces terminate with characteristic patterns that aid the identification of systems tract boundaries. These boundaries can be interpreted on seismic, well-log, and outcrop data, enabling a complete integration of geologic data. Comparison of systems tracts, generated by the simulator with different relative change of sea level histories, serves as an analog to understand the relationship between relative changes of sea level, changes in accommodation, and the development of system tract boundaries in sedimentary basins. These boundaries, tied to biostratigraphy and absolute ages, provide a chronostratigraphic framework for the basins.

Physical surfaces marked by sediment deprivation or influx and bypassing events bind stratal components, component groups, systems tracts, and sequences.
They form important surfaces for correlation (Figs. 10, based on parameters of Fig. 11 and 12). Rapid changes of base-level form these surfaces (Fig. 12). Deprivation surfaces bind component groups within aggradational coastal settings. Deprivation surfaces are relatively easy to identified in well-logs and form regional surfaces for correlation (Van Wagoner et al., 1991). Deprivation surfaces are often sites of ravinement as storms lower the wavebase and scour the sea bed, removing fine grained material from the surface.

As fluvial depositional systems prograde into a basin, small scale changes in base-level create episodes of deprivation or influx. These events control vertical lithologic changes and erosional contacts that define stratal surfaces within the rock record. Seismic reflection wavelets derived from multiple stratal surfaces are convolved into composite reflections at seismic resolutions. Stratal surface terminations form when strata thin or pinch out (Fig. 13). This situation often forms where base-level intersects the sediment interface or marine surfaces change from bypass to depositional conditions. At these intersections, boundaries form between deposition and non-deposition or erosion. Stratal termination patterns define discontinuities in the stratigraphic record (Fig. 12). Seismic configuration of the stratal patterns (parallel, chaotic, etc.) aid prediction of depositional environments and lithofacies between discontinuities.

**Stratal Components**

In this paper, we define stratal components as the smallest package of strata that are correlatable along an outcrop section, between wells, or within seismic data. Stratal components are a relatively conformable succession of genetically related beds or bedsets bounded by a correlative surface that contains a single depositional episode or a series of similar episodes without a significant change or break. Examples include coarsening or fining upward packages. These packages form parts of forstepping,
Figure 10. Stratal components, component groups, and relative changes of sea level. Sedimentary model with additional 4 meter 100 ky cycle recorded with 50 ky time steps - stratal vector, stratal components, component groups and relative changes of sea level. Component groups are bounded at the base of the component with the most downward directed stratal vector.
Systems Tract Boundaries

- Sequence Boundaries
- Top Trangressive
- Top Lowstand
- Top Slope Fan
- Top Shelf Margin Wedge

Highstand Systems Tract: 0-4, 18-21 & 33-35
Transgressive Systems Tract: 13-18 & 27-33
Upper Lowstand Systems Tract: 9-13
Lower Lowstand Systems Tract
Slope Fan Complex: 6-9
Basin Floor Fan Complex: 4-6
Shelf Margin Systems Tract: 21-27

Figure 11. Siliciclastic model - systems tract nomenclature. Systems tracts are based on onlap and downlap terminations and forestepping and backstepping stacking patterns.
Figure 12. Siliciclastic model - stratal termination patterns. Downlap forms when steeply dipping traction processes intersect the depositional interface in a distal direction. Coastal onlap occurs when base level intersects with sediment interface. Underwater onlap occurs when traction load sedimentation intersects with the sediment interface. Apparent truncation occurs when the sediment supply is distally starved. Toplap forms when sea level is at a relative stillstand.
Figure 13. Siliciclastic model - sequence stratigraphic surface nomenclature. The sequence boundaries are the first surface with onlap above. The top lowstand and top shelf margin surface is the surface with the maximum progradation within a sequence. The maximum flooding surface is the surface with the largest extend below sea level. The top basin floor fan surface is the last surface above a basin floor fan. The top slope fan surface is the last surface above a slope fan.
downstepping, backstepping, and upstepping packages, slumps, individual lobes of basin floor fans or leveed-channels, or difficult to correlate packages of shingled turbidites, channel fill, point bar migration (Fig. 14).

**Stratal Component Groups**

A component group is a pair or series of stratal components that alternate or repeat through an interval. Component groups include parasequences (Van Wagoner et al., 1991), shallowing upward cycles, punctuated aggradation cycle (PAC's of Anderson and Goodwin, 1991), prograding clinoform and basin floor fan pairs, and prograding clinoform and slope fan pairs. Many stratal component groups form in response to sediment supply changes or small-amplitude high-frequency changes (component group cycles) of sea level superimposed on large-amplitude low-frequency sea level changes (Fig. 15). These cyclic changes have the frequency of climatic fluctuation indicated by rhythmic bedding described by Fischer (1991). Commonly coarsening-upward intervals define component groups within a shallow marine setting (Fig. 14). Fining-upward intervals define component groups in a deep marine setting. They may correlate with wet-dry cycles and fining-upward intervals due to changes in flow regime in the non-marine setting. These stratal packages can also be episodic or auto cyclic. These groups or pairs may or may not have all the surfaces of a complete sequence, including an unconformity, ravinement surface, and deprivation surface.

Each component may or may not be associated with a stage of relative rise or fall in sea level and associated sediment flux. Separation of allo- from auto cyclic components are rather difficult. If a component group correlates regionally and forms part of a continuous progression of changes, then it is more likely a response to a sediment supply cycle or a relative change of sea level. An episodic component group caused by auto cyclic processes tends to be shorter in duration, local in extent, and variable in size. Climate cycles that control sediment flux or production may be partially responsible for forming component groups. Contemporaneity of cycles is difficult to
Figure 14. Components form in response to relative changes in sea level. Backstepping, upstepping, and forestepping form in response to a relative rise in sea level. Downstepping components form during a stillstand or small relative rise but are deposited between two components formed during relative falls. Levee-channel lobes, slumps, and basin floor fan lobes form in response to a relative fall in sea level.

document because time measurement resolution is commonly greater than the cycle period. Eustatic cycles producing component groups do not have amplitudes great enough to produce full suites of systems tracts under typical conditions.

Systems Tracts

The stratal components and component groups stack into one of five packages called the lower lowstand, upper lowstand, transgressive, highstand, and shelf-margin systems tract (Brown and Fisher, 1977; Posamentier and Vail, 1988). Each systems tract has characteristic stratal stacking and termination patterns and contains characteristic
Figure 15. Relative changes of sea level produced by the summation of third-order eustatic cycles with higher frequency cycles to create stratal components, component groups and systems tracts. Components are: b - backstepping, u - upstepping, f - forstepping, d - downstepping, s - slump, t - basin floor fan, c - channel levee lobe.
lithofacies relationships. Systems tracts group into larger units called depositional sequences bounded by sequence boundaries. An unconformity or its correlative conformity characterizes a sequence boundary.

**Lowstand Systems Tracts**

On the basis of the character of the discontinuity and stratatal components deposited during the lowstand, two types of sequences form: (1) a Type I, sequence containing lower and upper lowstand systems tracts, forms when base-level falls below the offlap break of the previous highstand; and (2) a Type II sequence containing a shelf-margin systems tract, forms when base-level fall remains above the offlap break of the previous highstand.

When sea level persistently drops below the offlap break, lower and upper lowstand systems tracts forms above a Type I sequence boundary. The lower lowstand systems tract consists of basin floor fan and slope fan lobes. The upper lowstand systems tract consists of a prograding complex. These units include stratatal patterns that onlap below a previously deposited offlap break (Fig. 10, 11, surfaces 4-13). If the bathymetric relief proximal to the delta, is great enough (usually greater than 100 meters), coarse grained sediment is removed by gravity flow sedimentation from the basinward edge of the delta and is transported into the deep basin, forming turbidite fans.

For hydrodynamic reasons not yet fully understood, deposition of two types of turbidite fans occur; initially the basin-floor fan followed by the slope-fan complex (Damuth, 1988). During a relative fall of sea level, a network of incised valleys is established on the shelf and shelf-margin, and submarine canyons are cut on the slope. This channel network feeds the basin-floor fan or slope-fan complex. The slope-fan is mud-rich compared with the basin-floor fan. Perhaps the initial incision removes the sand-rich shoreface. As base-level falls, the incised valleys are remove an increasing percentage of fine-grained material that was deposited on the delta front. Headward
erosion of sub-marine canyons also contributes a high percentage of fine-grained material.

**Basin-floor fan**

The basin floor fan lies above a Type I sequence boundary and below the top basin-floor fan surface. The vertical transition from coarse grained sediment to a condensed section defines the top basin-floor fan surface. The basin-floor fan is commonly a sand-rich, mounded unit that lies near the base of the slope at a transition to a flat basin floor (Sangree et al., 1988). The basin-floor fan commonly has a non-erosive base and terminates by downlap on all margins (Fig. 10, surfaces 4-6). A channel network feeding the basin-floor fan may not be obvious. The top basin-floor fan surface may be erosive in cases of bottom current reworking. In some cases, submarine currents rework the basin-floor fan, and form climbing toplap mounds (contourites of Mutti, 1989). Slumps and slides of the slope margin are often coeval the basin floor fan. They form large blocks that slide down the slope.

**Slope fan**

Slopes fans are a system of channel levees and lobes and associated facies such as slumps, mass flow units, and sub parallel beds that accumulate at the base of the depositional front. Slope fans lie above either a Type I sequence boundary or the top basin-floor fan surface and below a top slope-fan surface. Stratal patterns terminate landward by marine onlap and basinward by marine downlap or thinning. The top slope-fan surface is commonly a condensed section and formed by downlap of the overlying lowstand prograding complex, commonly in uppermost bathyal to outermost neritic water depths. The slope-fan complex is a series of channel overbank deposits and attached turbidite lobes at the base of slope (surfaces 6-9).

**Upper Lowstand Systems Tract**

The upper lowstand systems tract is characterized by coastal or deltaic deposition that begins below the offlap break of an older highstand (lowstand prograding complex).
Stratal stacking patterns exhibit a trend from oblique to aggradational offlap (Fig. 10, 16, surfaces 9-13) produced by initial progradation followed by increasing rates of aggradation (Posamentier and Vail, 1988). Stratal surfaces terminate landward by coastal onlap and basinward by downlap. Progradation continues as long as sediment supply can fill the space created by subsidence and rising sea level. Commonly, the maximum extent of progradation within the entire sequence occurs during the deposition of the lowstand prograding complex. In some cases, especially during a long term relative fall of sea level or uplift, the highstand may prograde past the lowstand prograding complex. Ravinement surfaces develop best during rapid rates of base-level rise and are common near the top of lowstand prograding complex within marine sediments. In sand rich systems, shingled turbidites deposit near the toe of slope.

The upper lowstand system tract lies above either the top slope-fan surface, basin-floor fan surface, or a sequence boundary. The top lowstand surface is a deprivation surface at the top of the lowstand prograding complex above the component containing the maximum progradation. In the shallow portion of a clinoform, reworked coarse-grained sediment commonly concentrates along a ravinement surface during the deprivation phase of the next component.

**Shelf-Margin Systems Tract**

The shelf-margin systems tract forms during the relative lowstand and early relative rise of sea level (Fig. 10 and 12). In most cases a Type II sequence boundary lies below a shelf-margin systems tract (surface 21). Stratal stacking patterns exhibit a trend from oblique to aggradational offlap (Fig. 10, 16, surfaces 21-28) produced by initial progradation followed by increasing rates of aggradation (Posamentier and Vail, 1988). Near the top of the shelf-margin systems tract, where rates of aggradation are high, ravinement surfaces form at the base of components in marine settings above storm wavebase. A deprivation surface above the maximum regression forms the top of the shelf-margin systems tract. This surface often onlaps above the offlap break of the
Figure 16. Siliciclastic model - stratal patterns produced by relative changes of sea level.
previous highstand. Stratal surfaces terminate landward by coastal onlap and basinward by downlap. In the shallow portion of a clinoform, reworked coarse grained sediment often concentrates along a ravinement surface during the deprivation phase of the next component. Gravity flow transport of coarse grained sediment before deposition of a shelf-margin systems tract is minimal.

In settings with a high sediment flux (Fig. 17), a Type I sequence with lowstand systems tracts may transform into a Type II sequence with a shelf-margin systems tract,
as a flexural response to loading depresses the margin, preventing base-level from dropping below the previous highstand.

**Transgressive Systems Tract**

The transgressive systems tract is characterized by onlap and backstepping component groups, increasing rates of aggradation (Fig. 16, surfaces 13-18, 28-33) or sediment starvation within the basin (Posamentier and Vail, 1988). At the base of the transgressive systems tract lies a top lowstand surface (surface 13), top shelf-margin systems tract surface (surface 28), or a sequence boundary (surfaces 4, 22). A deprivation surface containing the maximum landward shift of facies, or maximum flooding surface, lies at the top of the transgressive systems tract (surfaces 18, 33). During deposition of the transgressive systems tract, coastal or deltaic deposition dominates, and stratal patterns terminate landward by coastal onlap and basinward by apparent truncation. Stratal patterns in the transgressive systems tracts are parallel to sub-parallel. In the transgressive systems tract, ravinement surfaces are plentiful at the base of stratal components in shallow marine settings. The maximum flooding surface commonly consists of a condensed section within the basin, and of sediment-starved equivalents of coal deposits and lakes within the coastal plain and fluvial environment. Downlap of the overlying highstand systems tract is present above the maximum flooding surface.

**Highstand Systems Tract**

Early upward-decreasing aggradation followed by progradation (Fig. 16, surfaces 0-4, 18-21, 33-35) defines the highstand systems tract. At the base of the highstand systems tracts is a maximum flooding surface or a sequence boundary. In shallow settings, the top of the highstand is marked by erosion or exposure. Relative falls of sea level at the component group scale produce high-frequency unconformity-bound units similar to sequence boundaries in the late highstand systems tract. Ravinement surfaces are not commonly present within the late highstand systems tract.
In deep settings the top of the highstand is marked by an increase in grain size in overlying beds. During deposition of a highstand systems tract, coastal or deltaic deposition is characteristic, and stratal patterns terminate landward by coastal onlap. Stratal patterns in proximal portions of the highstand systems tract show thinning basinward, producing a downlap pattern onto the maximum flooding surface.

**Sequences**

Sequences, by definition, are the smallest stratigraphic packages that contain full suites of lowstand or shelf-margin, transgressive, and highstand systems tracts. Sequences form in response to cyclic changes in eustasy that range from 20 to 200 meters in magnitude, with periods that range from 20 ky to 3 My. Over the course of a eustatic cycle, in a shelf-slope-basin setting in a Type I sequence, deposition shifts from gravity flow sedimentation in deep marine settings during a relative fall of sea level, to coastal deposition in fluvial, coastal plain, and proximal marine settings during a relative rise of sea level.

Sequence cycles produce offlapping and backstepping stratal stacking patterns during coastal or deltaic deposition (Fig. 18). In settings with adequate siliciclastic sediment influx (greater than 1000 m$^3$/ky) and subsidence rates, correlation between seismic data and well-logs show stratal stacking patterns group into sets with similar aggradational and progradational patterns. These aggradational sets form a regular succession (Fig. 18) from aggradational offlap (prograding complex), backstepping (transgressive systems tract), sigmoidal offlap (early highstand systems tract), to oblique offlap (late highstand systems tract) (Vail et al., 1977). In the lowstand prograding complex, an increasing rate of aggradation associated with a decreasing rate of progradation characterizes aggradational offlap. In the transgressive systems tract, an increasing rate of aggradation associated with a landward retreat of each component or component group characterizes backstepping. In the early highstand systems tract, a decreasing rate of aggradation associated with an increasing rate of progradation.
Figure 18. Siliciclastic model - stratal stacking patterns, stratal termination patterns, and relative changes of sea level.
characterizes sigmoidal offlap. In the late highstand systems tract, a minimal increase or a
decrease in aggradation associated with the highest rate of progradation characterize
oblique offlap. The interaction of sediment supply and creation of accommodation space
(equation 1) by eustasy and subsidence (Jervey, 1988) produces this succession. The
succession is commonly resolvable on seismic and well-log data.

Incomplete Sequences

When a group of stratal components bounded by two sequence boundaries is
missing a systems tract, it forms an incomplete sequence. Under various tectonic and
sediment supply conditions, a sequence that recorded a full suite of systems tracts at one
location, forms an incomplete sequence with one or more of the systems tracts missing in
another. Commonly, three variations on the typical sequence occur 1) highstand is small
or missing, 2) transgressive systems tract is missing, 3) slope fan and/or basin floor fan
is missing.

![Graph of Siliciclastic model with an uplift of 10 mm/ky where highstand is reduced or
eliminated forming incomplete sequences. Transgressive systems tract is minimized or
eliminated during the first Type I sequence. The second Type II sequence boundary is
converted to a Type I sequence boundary. However the transgressive and highstand
systems tract are eliminated.](image)

Figure 19. Siliciclastic model with an uplift of 10 mm/ky where highstand is reduced or
eliminated forming incomplete sequences. Transgressive systems tract is minimized or
eliminated during the first Type I sequence. The second Type II sequence boundary is
converted to a Type I sequence boundary. However the transgressive and highstand
systems tract are eliminated.

The highstand systems tract is commonly small or missing when the eustatic
high is minimized by low subsidence rates, or by a short stillstand at the highest stand in
sea level, or in sequences formed during a long-term relative fall (Fig. 19). The
transgressive systems tract is commonly missing when sediment supply is greater than
the amount of accommodation space created during the relative rise of sea level (Fig. 17). The depositional system continues to prograde throughout the development of the sequence. A maximum flooding surface may be identified that represents the maximum rate of relative rise. It often contains the finest grain size and may not exhibit downlap onto the top of the previous lowstand prograding complex. The lowstand slope fan and/or basin floor fan is often missing in a setting with a shallow bathymetric contrast between the shelf and basin or if the relative fall is minor.

Timing of Systems Tract Boundaries with Respect to Relative Changes of Sea Level

Stratal components can be grouped according to their stacking patterns into systems tracts. The resulting stratal termination patterns and stratal stacking patterns correlate with specific changes in accommodation and relative changes of sea level. Continuous intervals bounded by stratal surfaces change in thickness due to relative rise and fall of sea level.

Figure 15 depicts the factors involved in an analysis of the relative changes of sea level to produce systems tracts and sequence boundaries. Tectonic subsidence calculated at the offlap break, third- and fourth-order eustatic changes of sea level, relative changes of sea level (sum of eustatic changes of sea level and tectonic subsidence), and position of the basal layer or total subsidence (the sum of tectonic subsidence and flexure loaded subsidence) through time. Dashed (relative rise) or solid (relative fall) vertical lines mark the timing of each stratal component that was produced within the model (Fig. 15). In this example, variations in the rate of tectonic subsidence through time are due to migration of the offlap break through regions of varying subsidence rates. For this model, total subsidence is about 60% greater than tectonic subsidence. Total subsidence varies through time with tectonic subsidence plus any subsidence or rebound due to rise and fall of sea level, and changes in sediment influx rate. In this model, the maximum flooding surface occurs late in the eustatic rise, 200
ky before sea level reaches its relative maximum and 200 ky after the inflection point of the eustatic rise.

Comparison of systems tracts generated (Fig. 16) by the stratal model with the relative sea level curve (Fig. 15) demonstrates the relationship between relative changes in sea level, changes in accommodation, and the timing and development of system tract boundaries. Changing aggradation and oblique offlap patterns define the history of relative changes of sea level.

The relative fall of sea level in a dominant cycle emphasizes influx surfaces. An influx surface defines a surface of erosional truncation below, with later onlap above. In most geologic settings, erosion destroys the relative sea level history on the shelf during a fall because shallow marine sediments erode. Erosion and flushing of sediment from incised valleys transports sediment farther basinward. Therefore, in a siliciclastic setting the influx surface correlates in the basin to a surface across which a transition from fine-grained to coarser-grained or siliciclastic-rich sediment occurs. However, siliciclastic and carbonate sedimentation on ramp settings often record the fall as down-stepping components and minimize transport to the distal basin.

The relative rise of sea level in a dominant cycle emphasizes deprivation surfaces. Sediment starvation, decrease in grain size, and subsequent downlap characterize deprivation surfaces. Condensed sections form in deep water during periods of starvation long enough to concentrate fossil material (Loutit et al., 1988; Kidwell, 1991). A maximum flooding surface marks the time when the rate of sediment influx begins to overtake the rate of space creation on the shelf within a depositional sequence.

Stratal surface 4 is a Type I sequence boundary because it marks beginning of a fall in base-level below the offlap break. Our computer experiments have found that a rate of relative fall greater than approximately 30 mm/ky reproduces observed
occurrences of gravity flow transport. This situation can occur under many combinations of eustatic and subsidence variations (Fig 20).

Stratal surfaces 4 through 9 form the lower lowstand systems tract. Stratal surface 6 is the top basin-floor fan surface. This surface forms during the early part of the relative fall. Downward stepping components or component groups, erosion, and gravity flow transport of coarse grained sediment into deep water is associated with this interval. Slope fans form during a relative fall of sea level when the magnitude of the fall reaches below the sand-rich fluvial and beach deposits of the previous highstand (typically greater than 20 meters). This may occur when the relative rate of fall of sea

Figure 20. Chart showing the relationship between magnitude and period of a eustatic cycle, and subsidence rates that are necessary to cause gravity-flow sedimentation with a relative fall threshold rate -30 mm/ky. Each line represents the minimum condition necessary for a eustatic cycle of a given period before bypassing will occur during the relative fall. Settings with subsidence rates and eustatic amplitudes below each line will produce bypassing conditions. Likewise, conditions above the each line will produce stable progradation.
level is the highest. Stratal surface 9 is the top slope fan surface. This surface forms at
the base of the relative fall, just before the rise.

Surfaces 9 through 13 bind the upper lowstand systems tract (lowstand prograding complex). This package exhibits a trend from oblique to aggradational offlap produced during the transition from slow rates of relative fall, stillstand to slow relative rise. Stratal surface 13 is the top lowstand surface. This surface forms during the early part of the relative rise when the space created by subsidence and sea level rise surpasses the rate of sediment influx. The timing of the top of lowstand surface varies by as much as 15% of the cycle periodicity.

Stratal surface 13 through 18 bind the transgressive systems tract. Backstepping and upward-increasing aggradation between flooding surfaces defines the transgressive systems tract. This pattern occurs when the combination of subsidence and rising sea level creates space faster than sediment supply can fill it. Stratal surface 18 is the maximum flooding surface. This surface forms during the last part of the relative rise when the rate of space created by subsidence and sea level rise is equal to the rate of sediment influx.

Stratal surfaces 18 through 22 bind the highstand systems tract. Stratal stacking patterns exhibit a trend from aggradational to oblique offlap (Fig. 10, 16) produced by initial progradation followed by increasing rates of aggradation. This pattern is produced during the transition from high rates of relative rise to slow relative rise, a stillstand, or the initiation of a relative fall of sea level (Fig. 15). Under these conditions, the sediment supply exceeds the space created by the combination of subsidence and sea level fall.

Stratal surfaces 22 through 28 bind a shelf-margin systems tract. Stratal stacking patterns exhibit a trend from oblique offlap to aggradational (Fig. 10, 16) produced by initial stillstand followed by increasing rates of aggradation. As in the lowstand prograding complex, this pattern is produced during the transition from slow rates of
Figure 21. Siliciclastic model - paleobathymetry and the depositional profile.
relative fall, stillstand to slow relative rise (Figs. 10, 15, and 16) after a small relative fall (Fig. 10 and 15). Stratal surface 22 is a Type II sequence boundary. It forms at the inflection point of the rate of relative fall (when the rate is highest).

Stratal surface 28 is the top of the shelf-margin systems tract. This surface forms when the rate of space creation begins to overtake the rate of sediment influx.

Figure 22. Siliciclastic model - chronostratigraphy. The lateral distribution and relative thickness of each unit within the cross section plotted parallel to a time horizon. They are colored according to the paleobathymetry at the time of deposition. See Figure 31 for the paleobathymetry legend. The time variant variables are plotted on the left margin. See Figure 11 for systems tract boundary definition.

Paleobathymetry is a function of local base-level and sedimentation rate at the location (Fig. 21). During periods of gravity flow transport into the deep basin, the paleobathymetry for locations on the relict shelf decreases but may not be recorded because of bypass and non-deposition on the shelf. When coastal sedimentation re-establishes, it may record a dramatic decrease in paleobathymetry from that of the
previous highstand (basinward shift of facies). During periods of aggradation, paleobathymetry in coastal regions may not increase because sediment supply keeps up with the rate of space creation.

**Sequence Chronostratigraphy**

Building a sequence chronostratigraphic chart (Fig. 22) requires dating the systems tract and stratal surfaces and plotting the stratal packages parallel to their ages. This chart shows the distribution of sediment as a function of time along a shelf to basin profile (Wheeler, 1958; Vail, 1977). It may also include time-variant inputs such as tectonic subsidence rates, sediment supply and eustasy. In this figure, eustasy is plotted on the left-hand margin.

Standard chronostratigraphy binds stratigraphy between the last or first occurrence of species. This position is a function of preservation, changes in depositional environment and evolutionary rates. Sequence chronostratigraphy is based on physical surfaces that can be correlated and dated with biostratigraphy.

The chronostratigraphic chart (Fig. 22) demonstrates how sinusoidal eustatic sea level changes produce the saw-toothed shaped onlap curve (Jevrey, 1988). The border marking the most landward position of sediment (onlap) has a shape similar to the relative onlap curve of Vail et al. (1977). A slow rate of onlap interrupted by rapid falls characterizes the onlap curve. The shape of the onlap curve is a function of sea level, subsidence, and antecedent bathymetry during each component.

Sequence boundaries and maximum flooding surfaces recorded in different tectonic settings and locations are chronostratigraphically equivalent within one-fourth of a cycle period (Fig. 23 and 24). For example, holding all other variables constant, increasing subsidence rates will cause the position within the cycle where a relative fall occurs to shift later in time (Fig. 23). Depending on the periodicity of a cycle, it may be difficult to determine the absolute age of a sequence boundary beyond the resolution of
Figure 23. The timing of the systems tract boundaries changes as the subsidence rate changes. Eustasy and the sediment volume are constant for each setting. The arrows represent the change from the zero subsidence rate setting. The sequence boundaries form later in the cycle as subsidence increases.

sb - sequence boundary  
msf - maximum flooding surface  
tlpc - top lowstand prograding complex  
tsf - top slope fan surface  
tbfs - top basin floor fan surface  
tsl - Total Subsidence  
rl - Relative Changes of Sea Level  
sl - Sea Level

Figure 24. Changes in subsidence rates, eustatic cycle amplitude, sediment supply and bathymetric profile will affect the timing of systems tract boundaries. The black and gray arrows around the circles represent the change in phase angle in the timing of surfaces by changing variable (in bold face) from a minimum value (0) to a maximum value (max.).

contemporary dating techniques. However, their globally similar magnitude, pattern, and duration suggest that eustasy causes many base-level changes. The top basin floor fan, top slope fan, and top lowstand prograding complex do not form consistent time horizons because they change considerably with changes in sediment supply.
SEPARATING COMBINED EVENTS

Combining tectonic events with a eustatic and sediment supply history for a section illustrates the relative importance of each variable in defining stratal geometries (Fig. 25). This model combines the tectonic event presented in Figure 4 with a 1 My, 30 meter eustatic cycle. This eustatic cycle produces a Type I sequence in this setting. However, during the rapid subsidence phase of the tectonic event, the maximum rate of relative fall of sea level is less than 30 mm/ky and a Type II sequence forms.

Backstripping stratigraphic columns along the cross section separates the individual events. Comparing the results with rates used within two dimensional models (Fig. 25) illustrate that a one dimensional analysis (Fig. 26) over-emphasizes uplift rates, attributes some of the flexural response to sediment and sea level changes to

Figure 25. Combined tectonic event with a 30 meter, 1 million year eustatic cycle. A 30 meter cycle produces Type I sequences for all but one cycle in the middle of an episode of rapid subsidence.
tectonic subsidence (Bowman and Vail, 1994b). Significance of this difference is discussed in the section on subsidence controls on stratal geometry later in this paper.

**DEPOSITIONAL SYSTEMS**

In nature, a range of variations exists on siliciclastic shelf-margins due to differences in the bathymetric profile with respect to the sediment source, and the distribution of erosion and climate controlled variables such as siliciclastic sediment flux, and carbonate sedimentation rates. In the next section, we illustrate some of the stratal patterns created by these differences in siliciclastic and carbonate sediments.

**Siliciclastic Depositional Systems**

Siliciclastic depositional systems include fluvial, coastal plain, barrier island, shoreface, delta, pro-delta, hemipelagic, pelagic, contourites, slope fan, and basin floor fan (Brown and Fisher, 1977). The position on the depositional profile, water depth, and depositional process determines the distribution of depositional systems (Bowman and Vail, 1994a). Simulation examples demonstrate the stratal geometries characteristic
of traction dominated, suspension dominated, and mixed traction and suspension settings.

**Climate-Controlled Variations**

Climate-controlled variations are an important source of eustatic changes. Global temperature influences the total volume of water stored in the oceans versus the volume stored on land in ice caps as well as local precipitation patterns. Major climatic cycles have been documented with periodicities ranging from 19 ky to 800 ky (Fischer et al., 1985; Fischer, 1991).

There may be a phase relationship between climatic and eustatic cycles (Perlmutter and Matthews, 1989; 1992). The most recent eustatic rise accompanied a wet period in the presently arid western North America and northern Africa. This may be due to melting of the ice caps as well as increased evaporation from the oceans due to warmer temperatures or may be due to a change in atmospheric currents. Other regions changed from dry to wet climates. Some ancient rocks show a correlation between maximum humidity and maximum flooding. Examples in the Rotliegende include eolian deposits within the lowstand and early transgressive systems tracts, that are stabilized by a rising water table during transgressive systems tract, and capped by fluvial sandstone (Glennie, 1984).

Climate controls the average grain size and composition of the siliciclastic sediment supply. This influences the ratio of sediment transported as traction load versus suspension load. Traction load is often coarse grained, whereas suspended sediment is fine grained. The principle differences in stratal patterns between traction-dominated and suspension-dominated systems are the linked coastline and offlap break in the traction system versus the unlinked coastline with the suspension-dominated system (Fig. 27). In traction load settings the coastline is linked to the offlap break because sediment must move along the sediment interface to migrate basinward. In suspension-load settings the offlap break migrates independent of the coastline because sediment deposited above
Figure 27: Siliciclastic model - depositional systems with a traction dominated sediment supply. The sediment supply contains 80% traction load and 20% suspended sediments. Traction deposits tend to form angular discontinuities such as downlap and toplap. Stratigraphic patterns tend to be sub-parallel rather than parallel and form more discontinuities comparable to suspension deposits.
Figure 28. Siliciclastic model - depositional systems with a suspension dominated sediment supply. The sediment supply contains 30% traction load and 70% suspended sediments. Suspension deposits tend to aggrade as whereas traction deposits prograde. Stratal patterns tend to be parallel.
storm wavebase is re-suspended and transported in the water column and deposited further offshore in a position below storm-weather wavebase (Fig. 28).

In settings with a low flux of mostly fine-grained sediment and a slow uplift, offlapping patterns do not exhibit transgressive or highstand systems tracts (Fig. 29). These offlapping sequences stack as a series of lowstand prograding complexes. This setting occurs in many locations along the western margin of Africa. The previously deposited highstand and transgressive deposits are eroded during the relative lowstand in sea level. We model this setting with 1000 m$^2$/ky of sediment, 10% traction load and 90% suspension load. We use typical erosion rates of 50 mm/ky for surface beveling, 100 mm/ky for channel incision rates, and 0.01 sq. km./ky for shoreface erosion.

**Bathymetry-Controlled Variations**

Bathymetry controls the nature of the basin-filling process and the rate of progradation and transgression.

**Ramp Setting $<1^\circ$**

In many cases, a sea level fall does not bring the base-level in contact with steep slopes on continental margin. This occurs where the bathymetric relief never exceeds neritic depths and the turbidite cannot generate the high velocities necessary to move sediment long distances. This may be due to several conditions including: 1) the
sediment supply is available to fill the space created by tectonism, or 2) sea level is within a long-term rise, flooding the shelf or there was not a well-developed shelf-basin edge. These conditions will produce a broad shallow dipping (<1°) bathymetric surface (ramp setting of Vail et al., 1991). A ramp setting often forms in foreland basins, or epicontinental seaways.

The ramp setting reduces slumping and minimizes gravity flow transport (Fig. 30). Besides prograding or transgressing during aggradation, clinoforms may downstep, forming a forced regression (Posamentier and Vail, 1988; Posamentier et al., 1992). Progradation accompanied by degradation without the formation of turbidites produces a unique sequence containing down-stepping component groups, called the lower lowstand systems tract (Vail et al., 1991). Subsequent prograding during the lowstand is called the upper lowstand systems tract.

![Figure 30. Ramp setting - An upper lowstand systems tract develops in a setting with shallow bathymetry. See Figure 11 for system tract boundary legend.](image)

During the relative fall, downstepping component groups are deposited progressively lower along the depositional surface. As base-level falls, these groups and previous highstand groups are partially or completely eroded. The wedge produced before the lowstand of the fall is called the lower lowstand systems tract. Subsequent prograding during the early rise of sea level is called the upper lowstand systems tract. The sequence boundary is marked by erosion and incised valleys in the area of the previous highstand. However, in the area of the lower lowstand systems tract, the time-equivalent surface of the sequence boundary passes under the onlap point of the
downstepping components and lies under the upper lowstand systems tract. The time equivalent surface of a sequence boundary occurs at the beginning of the relative fall. This physical expression in neritic water depths is a downlap surface at the base of an increase in silt or coarsening-upward trend. The coarsest sediment is not deposited until the lowstand of the relative fall. During down-stepping, portions of previous component groups lying above base-level may be eroded. This erosion may result in a local unconformity that is above the sequence boundary at the top of the lower lowstand systems tract.

**High Sediment Volume**

Sediment supply has an important impact on stratal geometries. Measured siliciclastic sediment fluxes range from 50 to over 1,000,000 m$^2$/ky. Comparison of the model response to a setting with a large (20,000 m$^2$/ky) and small (2000 m$^2$/ky) sediment supply demonstrates (Fig. 17) that as sediment supply increases, the thickness between stratal surfaces increases, rate of progradation increases, and the size of the transgressive system tract is proportionately less than that of the lowstand and highstand system tracts (components change from transgressive to progradational as sediment supply increases). In settings with very high sedimentation rates and low subsidence rates, or a fall in long-term sea level, may not create accommodation space fast enough to produce a transgression. A similar response occurs with a slow rise or a rise following a sea level fall of small magnitude. Large volumes of sediment flux cause increased flexural subsidence that may transform a Type I sequence into a Type II sequence.

**Carbonate Depositional Systems**

Three models of carbonate sedimentation document the response of the carbonate factory to the sea level curve. The three models are a shelf-margin setting with slopes that are less than 10°, a steep slope margin, and an oceanic atoll. The initial bathymetric
Figure 31. Paleobathymetry for a carbonate model in a shallow-dipping margin setting. Marine onlap is very common in the carbonate setting. With a large bathymetric contrast, the lowstand prograding complex is minimized because the paleobathymetry within the depth of high productivity is reduced the production depths is reduced. See Figure 11 for systems tract boundary legend.
profile and subsidence rates are equal to those used in the siliciclastic model.

The dominant controls on carbonate productivity are the amount of light (latitude, and water depth) and position on the margin with respect to nutrients from the basin and salinity (Lerche et al., 1987; Bosscher, 1992). The model has three bathymetrically-controlled productivity functions. They use maximum values that range from 50 mm/ky for geographically unrestricted productivity (shelf-margin), 80 mm/ky for unrestricted traction load, and 250 mm/ky for suspension load, that adds up to a total of 480 mm/ky in high productivity settings. Pelagic sedimentation rates are 10 mm/ky. The type of carbonate sediment varies by position on the carbonate margin with respect to the shelf-margin and tidal range. Models that define sediment type by paleobathymetry alone (Fig. 31), oversimplify depositional environments.

The dominant regional control on carbonate growth is paleogeography. Ziegler et al., (1984) have shown that tropical carbonate sedimentation rarely produces at high rates above 30 degrees paleolatitude because light is refracted off the water surface at angles greater than 22 degrees. The values for this model are rates typical of low latitudes. When modeling a real setting, we prefer to keep productivity rates constant through the entire model, unless the progradation direction significantly changes. A value is chosen that produces the volume of carbonate sediment observed or a value appropriate for the paleolatitude.

**Shallow-dipping (<3°) Carbonate Margin (Ramp Setting)**

A carbonate platform built on a shallow-dipping margin (Fig. 32), without shoreface erosion preserves component groups during all phases of a eustatic cycle. The lack of erosion is typical of a well-cemented bank. Down-stepping component groups form during a relative fall and are equivalent to the basin floor and slope fan in a siliciclastic shelf-margin setting. The highest volume of sediment production occurs during the transgressive and early highstand. This is due to the large surface area available within the water depths for optimum growth.
Figure 32. Carbonate model of a shallow-dipping margin setting. The slope fan and basin floor fan do not develop because there is minimal erosion. Total carbonate production is 450 mm/ky. See Figure 11 for system tract boundary legend.
Figure 33. Carbonate model - steep platform margin setting. Carbonate sedimentation is minimized during the lowstands when the area for carbonate production is reduced. During highstands carbonate production significantly increases. Much of this sediment is deposited as calciturbidites in the basin. See Figure 11 for systems tract boundary legend.

A carbonate platform built on a shallow-dipping margin (Fig. 32), with shoreface erosion active, removes most of the component groups deposited during a relative rise and remnants of downward stepping component groups deposited during a relative fall. The removed sediment is deposited as gravity flow sediments or the lowstand prograding complex. Shoreface erosion might be typical for a bank subject to intense erosive activity or poorly cemented. Similar to the siliciclastic setting, slumping processes and turbidity currents transport the unconsolidated and eroded material into deep water. Again, high volumes of sediment production due to a large surface area available
within water depths optimal for growth, characterize the transgressive and early highstand systems tracts. The surface equivalent to the sequence boundary is onlapped by the transgressive systems tract with truncation beneath, and is very similar in character to the siliciclastic setting.

Platform drowning occurs when sea level rises faster than sediments can accumulate (Schlager, 1988). If the relative rise of sea level is faster than the carbonate production rate, the platform will drown. Carbonate production rates range between 150 and 1000 mm/ky depending on the paleogeographic setting and contemporary carbonate producing organisms. These rates are significantly less than coral growth rates applied by some authors in comparison with sea level rate changes to infer when drowning might occur.

**Steeply-Dipping Carbonate Margin**

A steeply dipping margin, flanking a basin with upper bathyal or greater water depths (45° or greater), is similar to the modern Bahama platform (Fig. 33). This setting establishes a condition where deposition of carbonate sediment is strongly favored during the late transgressive and highstand systems tract (Droxler and Schlager, 1985). This occurs when productivity is minimized because the carbonate factory has very little substrate on which to establish itself during a sea level lowstand. However, when sea level floods the platform, a dramatic increase occurs in the surface area available for carbonate production, and the rate of sediment production increases accordingly.

The model in Figure 33 was built on a margin that dips 45° with the previously used sea level curve. The model produced ten times more sediment during the late transgressive and highstand than the lowstand with an 80 and 25 meter sea level fall. Sediment accumulated in the basin where the bathymetric gradient decreased, forming a dump bump (Wilber, 1990).
Figure 34. Carbonate model of an rimmed oceanic atoll. Sedimentation dominates during the transgressive systems tract. Highstand progradation is minimal.

Oceanic Atoll

A reef-rimmed lagoon characterizes carbonate sedimentation on an atoll (Fig. 34). Reefs grow most rapidly along the atoll margins because wave activity and tidal currents flush the location with sea water (Wilson, 1975). Locations proximal to the open seaway are more productive because upwelling provides extra nutrients. Rimmed reefs aggrade during relative sea level rises because this location has the highest productivity and can keep up with a relative rise. When sea level reaches a stillstand, the lagoon fills with sediment produced both within the lagoon and at the reef because the strongest and most prevalent currents originate from the ocean basin and transport the sediment from the reef rim into the protected lagoon. After the lagoon fills, sediment is carried offshore and is deposited as calciturbidites on the slope, or in the basin if the slopes are too steep to allow deposition.
Figure 35. In a model of a mixed siliciclastic and carbonate system, carbonates flourish during the transgressive systems tract. With the subsidence rates and bathymetry used in this model, progradation is diminished. In the rock record, pelagic sedimentation flourishes during the lowstand relative to the transgressive systems tract. This does not occur in the model because the carbonate production is not integrated basinwide. This would decrease basinal production during the transgression.
Climatic variations within carbonate depositional systems control the volume of carbonate sediment and mineralogy (Opdyke and Wilkinson, 1990). Evaporites, eolian sands, and pisolites record dry settings. Root borings or coal seams deposited in mangrove swamps record humid settings. These differences change only the lithofacies or mineralogy within the coastal plain, or tidal flat and do not change the stratal pattern geometries.

**Mixed Siliciclastic/Carbonate Depositional Systems**

Mixed siliciclastic and carbonate settings produce a unique pattern during a eustatic cycle. In an example built with low siliciclastic influx (1000 sq. m/ky) and a sensitive carbonate production factory (siliciclastic sedimentation limit of 50 mm/ky), carbonate production stops during lowstand and highstand prograding systems tracts. However, during the transgressive systems tract the topography of the buildups of the previous component group forms an ideal locality for continued carbonate growth because they are distal to siliciclastic and freshwater influx (Fig. 35). In the rock record, pelagic sedimentation flourishes during the lowstand prograding complex when compared with the transgressive systems tract perhaps because CO₂ is abundant because it is not being absorbed by shelf margin production. This does not occur in the model because the carbonate production is not integrated basinwide. Increased competition for CO₂ and nutrients from flourishing shelf margin production would decrease basinal pelagic production during the transgression.

**PHYSICAL CONTROLS ON STRATAL GEOMETRY**

In this section, we present a discussion of the importance of subsidence, compaction and discharge on stratal geometries. Modeling suggests that 1) relative changes of sea level should be calculated using total subsidence, 2) over-pressures delay the manifestation of compaction-derived subsidence, 3) change in discharge is an
important factor in controlling stratal geometries in the fluvial setting, and strongly influences the volume of sediment delivered to the basin margin.

Subsidence

The stratal path is the record of accommodation changes represented by the stratal surfaces of a stratigraphic cross-section (Fig. 15). The stratal vector is the change in position of the offlap break, or coastal plain per time step (Fig. 15) recorded in a stratigraphic cross section. It is best to base this measurement on the offlap break, because post-depositional alteration is less likely at this point during events leading to the next time-line. For example, ravinement processes may erode the coastal plain during a rapid rise, reducing the recorded position of the stratal line.

A question remains whether tectonic or total subsidence is recorded in the stratal path. Posamentier et al., (1988) and Vail et al., (1991) suggested the important subsidence is tectonic subsidence. Simulation results demonstrate that tectonic subsidence, compaction, and subsidence due to flexure loading of both sediment and water are important contributors to relative changes of sea level (Jordan and Flemings, 1991; Reynolds et al., 1991).

We observe that all subsidence that increases the thickness of a stratal component is recorded in the stratal path. Subsidence that is a response to localized instantaneous sediment or water loading, increases the thickness between two time-lines if sedimentation rates are higher than total subsidence. This subsidence may cause paleobathymetry to increase if sedimentation rates are less than total subsidence rates. The stratal path therefore records subsidence caused by flexure due to sediment and water loads. The stratal path records tectonic subsidence because it is independent of sedimentation and alters paleobathymetry. Subsidence due to compaction increases the thickness between stratal lines.

Conclusions derived from one-dimensional backstripping analyses suggest that the formula for calculating relative changes of sea level should be the sum of tectonic
subsidence and eustasy. Total subsidence is the sum of all subsidence, including tectonic subsidence, sediment and water loading, adjusted for the effects of compaction and paleobathymetry. Backstripping approximates tectonic subsidence by removing subsidence due to the sediment and water load within a section. In a backstripping analysis, tectonic subsidence accounts for roughly one third of the total subsidence. The analysis assumes that the crust instantaneously adjusts for the sediment load during sedimentation.

One-dimensional isostatic loading, as employed in backstripping analyses, implies that the crust has no strength, or sedimentation is equal everywhere. With these assumptions, one could argue that subsidence due to sediment loading will not affect the surface of deposition. As the water column fills, the base of the interval subsides faster than the base would without sediment, but never fast enough that paleobathymetry increases. It is possible to conclude, therefore, that sediment loading does not affect the depositional history even though it increases the thickness between time lines.

However, two- or three-dimensional flexure-loaded models suggest that the formula for relative changes of sea level should be the sum of total subsidence and eustasy. Sediments cannot differentiate between subsidence due to tectonics, compaction, and sediment and water loading. An example is a setting where components vary in thickness. One locality might be filling in, causing flexural subsidence in neighboring localities, while neighboring localities are sediment-starved. In sediment-starved localities, paleobathymetry increases due to flexure loading, which causes a relative rise of sea level. In summary, flexural loading causes two effects: (1) it delays the complete expression of the tectonic event, or spreads the duration of the subsidence event in the stratal record, and (2) it pulls the depocenter closer to the fluvial source.
Compaction

Reduced compaction rates due to lithostatic pressure (over-pressure) cause a relative fall of sea level by initially reducing the rate of space creation. The primary effect is to delay the manifestation of compaction-derived subsidence. Compaction occurring under hydrostatic pressure is a component of relative change of sea level because the associated subsidence occurs soon after the time of deposition. In regions experiencing high rates of sedimentation, subsidence due to compaction can become a very large component as well. An example was produced with initial porosities between 40-60% and compaction rates ($r_c$) between 0.001 and 0.005 (Fig. 36) using the following equation:

$$\phi = \frac{\phi_0}{1 + z \cdot r_c}$$ (2)

$z$ is depth in meters. Compaction subtly alters the geometry of the stratal patterns near sharp lateral contrasts in lithology. By removing volume, it also slightly reduces the rate of progradation. The largest vertical differences occur in the center of the stratigraphic

![Figure 36. The effect of compaction on stratal patterns. The volume is reduced which decreases the distance of progradation. Along the vertical axis the middle horizons are the most affected because the early horizons are above a thinner substrate, whereas the later horizons have been buried less.](image-url)
package because the lower layers have less compacting substrate and the upper layers have less burial.

**Discharge**

Climate-induced changes in water discharge within a drainage basin influence sediment volume and grain size as much as changing gradient due to a base-level fall (Table 2). Both sediment flux and grain size are strongly dependent on the stream velocity. In modern examples, such as the Mississippi River (Fisk, 1944), fluvial gradients decrease basinward, whereas stream velocity increases.

<table>
<thead>
<tr>
<th>Rainfall (m/y)</th>
<th>%Runoff</th>
<th>Gradient x 10^{-4}</th>
<th>Velocity (m/sec)</th>
<th>Grainsize (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.25</td>
<td>10</td>
<td>1.89</td>
<td>0.55</td>
<td>0.1-0.5</td>
</tr>
<tr>
<td>0.5</td>
<td>70</td>
<td>1.89</td>
<td>1.31</td>
<td>2.1</td>
</tr>
<tr>
<td>0.5</td>
<td>70</td>
<td>2.55</td>
<td>1.52</td>
<td>2.5</td>
</tr>
<tr>
<td>1</td>
<td>70</td>
<td>1.89</td>
<td>1.66</td>
<td>3</td>
</tr>
<tr>
<td>100</td>
<td>70</td>
<td>1.89</td>
<td>7.69</td>
<td>150</td>
</tr>
</tbody>
</table>

Empirical equations (King, 1939) for fluid flow in channels show a change in stream flux has a more significant impact on the maximum grain size transported than a change in gradient due to a typical fall in sea level. Table 2 illustrates the relative importance of flux and gradient with an estimate of the maximum grain size transported in a basin that has low rainfall (0.25 m/yr); average rainfall (0.5 m/yr with and without a 125 meter sea level fall); high rainfall (1 m/yr); and is very wet (100 m/yr, equivalent to a basin experiencing deglaciation).

In a 1600 by 300 km basin, with an average rainfall of 0.5 m/yr and 70% runoff, the flux is $2.4 \times 10^{11}$ m^3/yr; average flux, Q, would be $7.7 \times 10^3$ m/sec. Channel width, w, change in response to flux, according to:

$$w = 4 Q^{0.5}$$  \hspace{1cm} (3)

The channel depth d, is a function of flux and channel width:

$$d = \frac{w}{50}$$  \hspace{1cm} (4)

We use a channel roughness factor, n, of 0.033, and set the maximum height of the
Figure 37. Cross plot of average stream velocity versus average stream discharge demonstrating how discharge of a stream strongly controls the stream velocity. (data from Leopold and Wolman, 1957)

Figure 38. Cross plot of average stream velocity versus average stream gradient demonstrating how slope does not influence the stream velocity. (data from Leopold and Wolman, 1957)
basin to 1000 meters at the drainage divide. The fluvial gradient during a highstand is 0.000189 and a lowstand is 0.000255, assuming the knickpoint migrates to the far extreme of the basin. The Manning equation:

\[
V = \frac{kR^{2/3}G^{1/2}}{n}
\]

(5)

predicts velocity of \( V \) (m/sec), given constant \( k = 1.0 \text{ m}^{1/3} \), hydraulic radius \( R \) and gradient \( G \).

A 125 meter sea level fall produces an increase in the maximum grain size transported from 2.1 to 2.5 mm. This is equivalent to precipitation change from 0.5 m/yr to 1.0 m/yr. However, melting of temperate ice sheets may produce a 100 fold increase in water flux. This is capable of increasing the channel depth and moving small boulders.

Cross-plots of stream velocity versus discharge of a worldwide sample of natural channels (Leopold and Wolman, 1957) show a strong correlation (Fig. 37), whereas stream velocity and gradient show a weak correlation (Fig. 38). Glacial melt water, as well as changes in rainfall patterns can dramatically change the flux within a stream channel. The volume of fluid in a river system during a flood increases both sediment flux and maximum grain size transported, whereas fair-weather water flux lowers sediment flux and decreases grain size. The knickpoint (Begin et al., 1980) marks the upstream limit to changes in gradient due to a relative sea level fall. Erosion rates of the channel substrate and the magnitude and duration of a sea level fall determine the position of the knickpoint. Studies along the Texas coast of the Gulf of Mexico show the knickpoint rarely extends more than 125 km inland (Saucier, 1974; Blum, 1990). Therefore, it is reasonable to expect water flux changes have a stronger influence on sediment volume and average grain size than a change in gradient due to a sea level fall.

CONCLUSIONS

Tectonism and eustatic change are the dominant controls on relative changes of sea level. Spreading rates along mid-oceanic ridges control the average age of oceanic
crust, and, therefore, ocean-basin volume (Rona, 1973; Pitman, 1978; Kominz, 1984). Ocean basin volume influences tectono-eustasy, a major component of eustasy (as much as 200 meters over 90 My), and cause eustatic changes with rates less than 2.5 mm/ky (Kominz, 1984), a rate of change that is very low. Changes in tectonic subsidence vary throughout a region and differ both in timing and magnitude in each basin. Most tectonic changes are irreversible. For example, when compressional deformation reactivates extensional structures, the deformation folds the layers within the downthrown blocks rather than "un-rotating" the subsidence along the boundary of the blocks. Rebound is limited and often has a different distribution. The writers have not observed short-term cyclic changes (less than 2 My) that are "tectonic" in character. Maximum subsidence rates during the maximum transgression characterize a transgressive-regressive cycle associated with tectonic subsidence events. Sediment supply controls the amount of space that is filled and is indicated by the distance or rate of progradation. Minimal subsidence rates during the maximum transgression characterize a transgressive-regressive cycle associated with changes in sediment supply.

High-frequency (20-2000 ky) eustatic changes are the dominant control on the distribution of lithofacies. Dominant eustatic cycles form systems tracts composed of component groups and stratal components produced by higher-frequency subordinate cycles. Eustatic cycles commonly induce gravity-flow transport of coarse-grained sediment into the deep basin. Sequence stratigraphy explains both the character and distribution of bypassed sediments, and provides a chronostratigraphic framework to predict the timing of major gravity-flow events. The technique explains lithologic trends observed in well logs and the character of seismic reflection patterns.

Maximum flooding surfaces form the most consistent surfaces for biostratigraphic dating. They preserve concentrations of fossil material. They form by changes in tectonic subsidence rates, sediment supply, and eustasy. The most important
surfaces form when these three effects combine to create large increases in accommodation space.

Sequence boundaries are important for defining the physical expression of a sequence but are dated more often by the maximum flooding surfaces between them (Vail and Wornardt, 1990). Sequence boundaries are optimal for defining stratigraphic packages because they are contained in all stratigraphic sections, including those with non-deposition. The boundaries are based on physical surfaces that are more precise than intervals.

The greater the rate of relative rise, the more likely ravinement surfaces are to form. Ravinement is more likely to form during the lowstand prograding complex and transgressive systems tract.

Modeling of stratigraphy is important for testing stratigraphic and tectonic hypotheses. Stratigraphic modeling is useful for:

1. generating geologic cross-sections,
2. predicting distribution of lithology,
3. simulating geologic history,
4. quantifying tectonic subsidence, sediment supply and eustatic history,
5. comparing scales of different concepts,
6. constraining concepts with geologically reasonable values,
7. generating synthetic seismic sections, and
8. providing a constraint on lithology and burial history for thermal maturation or fluid flow models in regions of unknown lithology.
Chapter 4.

Subsidence Analysis of Stratigraphic Sections --
Comparison of One- and Two-Dimensional Data

ABSTRACT

Analysis of stratigraphic sections generated from a two-dimensional basin-fill simulator that compensates for crustal response to sediment and water loading suggests that one-dimensional backstripping analyses may have an error in the tectonic subsidence that is as much as 10-25% of the total subsidence. One-dimensional backstripping will introduce inaccuracies that arise from the principle assumption that depositional and erosional events recorded in a stratigraphic column are equal in magnitude within the flexural wavelength of the column. This varies as a function of initial bathymetry, flexural rigidity and position with respect to regional sedimentation and erosion.

Analysis of stratigraphic sections can provide a sediment accumulation and burial history, and a first approximation of the tectonic or total subsidence and uplift history. To determine the tectonic subsidence history of a location, the backstripping technique removes the effects of sediment- and water-loading from the total subsidence history. This approximates a theoretical depth to basement after removal of the sediment and water. We apply an air-loaded solution by removing the water layer. The air-loaded solution produces a succinct accounting for water loading due to changes in sea level and an easy visualization of the objective of the technique.

Sections were analyzed from a model that contained no sea level and tectonic changes. The results indicate three phases of pseudo-tectonic activity during infilling caused by: (1) early subsidence during infilling of distant locations and relative starvation; (2) uplift during infilling due to local sediment stacking and the flexural dispersal of the load throughout the flexural wavelength; and (3) subsidence during infilling of distant locations and filling above sea level. It is necessary to correct for these inaccuracies when
the results of the analysis constrain the subsidence history of a stratigraphic simulation. However, there is no systematic approach in one-dimension to account for the variations.

**INTRODUCTION**

One-dimensional backstripping techniques determine the total and tectonic subsidence history (Fig. 1), subsidence rates (Fig. 2), sediment accumulation and burial history of a stratigraphic section (Fig. 3). Long-term sea level curves have been calculated by adjusting the long-term sea level to match the calculated subsidence history with a theoretical subsidence history (Bond and Kominz, 1984). In this paper we test the feasibility of this approach by comparing results from a theoretical dataset generated by a two-dimensional stratigraphic simulation program (PHIL) that includes calculations for flexure loading and compaction (Bowman and Vail, 1993). Eustatic changes and tectonic subsidence rates are zero in this dataset. Paleobathymetry and the total subsidence (loaded and compacted stratigraphic column) are recorded by the simulator. The compaction algorithm used in the simulator is identical to the compaction algorithm in the subsidence analysis program. Therefore, the differences between the predicted tectonic subsidence of the backstripping algorithm and the input tectonic subsidence of the model illustrate the nature of the limitations of the backstripping technique.

Besides problems with paleobathymetry and age dating, errors arise in one-dimensional backstripping due to assuming local isostatic compensation. Inherent to the one-dimensional analysis is the assumption that the crust has no strength. When backstripping a section that filled a pre-existing water column, the technique calculates an uplift during the infilling phase that ranges in magnitude from 10 to 25% of the total subsidence (Fig. 4). This varies as a function of initial bathymetry, flexural rigidity and position with respect to regional sedimentation and erosion. The error is less when the paleo-water depth is closer to 0. The error is due to the flexural loading effect of sediment filling a pre-existing water column. The error increases with increasing effective elastic thickness and initial maximum paleobathymetry of the margin. A typical
elastic thickness of 10 km produces an error of 20% of the total subsidence. The expression of this error occurs when the sediment is delivered to the region.

"Tectonic subsidence" is caused by the interaction of several processes, including (1) flexural bending due to loads outside the location, cross-section, or region; (2) thermal contraction or uplift; (3) delayed compaction of sediments; and (4) faulting.
Figure 2. Tectonic and total subsidence rates between intervals of the Lamar Hunt McMahon well. Calculating the rates emphasizes the magnitude of the changes that occurred as well as the variation in the data.

Figure 3. Geohistory of the Lamar Hunt McMahon #1 well, Jasper County, Texas. This figure plots the depth of each layer adjusted for compaction, paleowater depth, and eustatic sea level. Shifts in subsidence rates show the timing of regional tectonism.
Tectonic subsidence might also include compaction of sediments below the modeled layers. Density-driven mass displacement around growth faults and diapirs contribute to

Figure 4. Calculated tectonic subsidence located at 62.5 km of model results with a 1, 2, 4, 8, 16, 32 km effective elastic thickness. The dashed lines are the sediment surfaces, the solid lines are the total subsidence histories, and the dotted lines are the calculated tectonic subsidence histories. This initial maximum paleobathymetry of the basin was 250 meters.
tectonic subsidence because it effects the total subsidence. The results of any one of these processes effect in the motion of the basal surface of a stratigraphic section.

**TECTONIC SUBSIDENCE ANALYSIS TECHNIQUE**

The objective of subsidence analysis is to separate the active tectonic processes from passive responses such as compaction or flexural loading of sediment within the stratigraphic column. Our program (BasinWorks) employs a forward-modeling algorithm. Initially the calculation approximates the present porosity state of each layer from the present burial depths of the input stratigraphic column. Next the program calculates the initial porosity, using an initial depth equal to the decompacted median thickness of each layer. The program continues to forward-model by adding each layer to the top of the stratigraphic column and adjusting for the compaction state of each subjacent layer at its mean burial depth. The algorithm calculates compaction as a function of burial depth (Equation 1). If at any time the burial depths exceed the present depths of each layer, due to erosion and subsequent uplift, a new approximation of the present porosity is assigned to each interval. This porosity should reflect the deepest burial, as sediment does not decompact when unburied. The initial conditions are recalculated if there is a change from the first approximation. The algorithm iterates through the forward model until there is an insignificant change between the approximation and results.

We use age-depth pairs based on either biostratigraphic tops or sequence boundaries in the sense of Vail et al., 1977. We found the use of sequence boundaries reduces the variance of the data, especially when the section was deposited in a shallow-water setting. A sequence boundary forms a good datum because this reflects a highstand in sea level that usually is tied to a long-term sea-level history. Sequence boundaries are dated for much of the Mesozoic and Cenozoic and offer comparatively high-resolution time control (Haq et al., 1987; Harland et al., 1989). A location in a dominantly shallow
marine setting is optimal for subsidence analysis because the error in determining paleobathymetry is minimized.

**COMPACCIÓN**

Compaction functions in BasinWorks are defined by two user-specified variables (porosity at the time of deposition, \( \phi_0 \), and compaction rate, \( r_c \)) and one of the following three equations (from Sclater and Christie, 1980; Falvey and Middleton, 1981; Dickinson et al., 1987; Royden and Keen (1980), McBride et al., 1991):

\[
\phi(z) = \frac{\phi_0}{1 + z \cdot r_c} \quad (1)
\]

\[
\phi(z) = \phi_0 - (z \cdot r_c) \quad (2)
\]

\[
\phi(z) = \phi_0 e^{-z \cdot r_c} \quad (3)
\]

\( \phi \) is the porosity at a given burial depth, and \( z \) is the depth. Each equation approximates the porosity as a function of burial depth, but with a different function (Fig. 5).

**TECTONIC SUBSIDENCE**

The tectonic subsidence algorithm within BasinWorks is adapted from Van Hinte (1978, 1981) and Steckler and Watts (1978) for approximating the total and tectonic-subside history of a region. We use an air-loaded solution for removing the subsidence due to local isostatic loading of a stratigraphic and water column. The solution is similar to the method of Steckler and Watts (1978) except that the water column is treated as a layer whose loading effect is removed from the total subsidence. This allows compensation for time-dependent water-depth changes (eustasy). If deposition at the base of a section begins in deep water, this algorithm will more accurately approximate tectonic subsidence.

Total subsidence is calculated by compacting the sediment column to the appropriate state for each depth and adjusting for eustasy and bathymetry. Tectonic subsidence is calculated by subtracting the thickness of mantle equivalent to the mass of
the sediment and water load from the total subsidence (Sawyer et al., 1982):

$$S_{\text{tect}} = S_{\text{tot}} - S_{\text{LL}} \quad (11)$$

$S_{\text{tect}}$ is the tectonic subsidence, $S_{\text{tot}}$ is the total subsidence, and $S_{\text{LL}}$ is the local loading subsidence.

Subsidence due to sediment- and water-loading at each step is calculated by using the following equation:

$$S_{\text{LL}} = \frac{T \rho_c + P \rho_w}{\rho_m} \quad (12)$$

$T$ is the column total thickness, $\rho_c$ is the column total density, $P$ is the paleobathymetry, $\rho_w$ is the density of water, and $\rho_m$ is the mantle density. A time series plot of $S_{\text{tect}}$ is an
air-loaded tectonic-subsidence history of the modeled section above an elastic crust with no strength.

STRATIGRAPHIC MODEL

It is difficult to collect a paleobathymetrically well-constrained dataset from the geologic record where the tectonic history of the location is known precisely. Therefore, to test the ability of the subsidence analysis to calculate the tectonic-subsidence history from a section, we analyzed the results from stratigraphic simulations that contained no tectonic subsidence or eustatic changes and a constant siliciclastic sediment influx to document the purely flexural effects in a cross-section (Fig. 6).

FLEXURAL SUBSIDENCE

The simulations account for flexural response to crustal loading of sediment fill and changes in water depth with the following equation for a homogeneous elastic plate (Turcotte and Schubert, 1982):

\[
D \frac{d^4 w}{dx^4} = q(x) - (\rho_m - \rho_w) g w \quad (4)
\]

D is the flexural rigidity of the bending plate, q is the downward or upward directed force per unit area applied to the plate, \( \rho_m \) is the mantle density, \( \rho_w \) is the density of water, g is the gravitational constant, and w is a horizontal force. Young's modulus, E, we hold constant at 70 GPa, and Poisson's ratio, n, is constant at 0.25. Equation 5 defines the flexural rigidity, D:

\[
D = \frac{Eh^3}{12 (1 - n^2)} \quad (5)
\]

Equation 6 defines the effective thickness of the lithosphere, h:

\[
h = \sqrt[3]{\frac{12 D (1 - n^2)}{E}} \quad (6)
\]

Therefore, E, n, and h are unknown variables that must be defined. The flexural rigidity,
Figure 6. The flexural response of the crust during migration of the depocenter. The cross-section represents 10 My with 500 ky timelines. In this example the crust has an effective elastic thickness of 1 km and the initial maximum paleobathymetry was 250 meters. Superimposed on the section is the deflection produced by a load at one location.
$D$, is also defined by the flexural parameter, $\alpha$, and the following equation:

$$D = \frac{\alpha^4 g \left( \rho_m - \rho_w \right)}{4} \quad (7)$$

$g$ is the gravitational constant. $\alpha$ defines the depression wavelength (Fig. 7) by:

$$x_0 = 2.35 \alpha \quad (8)$$

$x_0$ is the half-width of the depression created by the load. Also, $\alpha$ defines the distance to the top of the forebulge, $x_b$ (Fig. 7) by:

$$x_b = \pi \alpha \quad (9)$$

A value for $\alpha$ is selected to reflect the flexural deflection observed in the cross section.

![Diagram illustrating the relationship between Alpha and the shape of the deflection produced by the line load $V$.](image)

Figure 7. Diagram illustrating the relationship between Alpha and the shape of the deflection produced by the line load $V$.

Values for the effective thickness of the lithosphere have been summarized by Watts et al., (1982). Crustal wavelengths range from 0 kilometers in an active rift, to upwards of 200 kilometers for mature crust (>1 Ga) (Watts et al., 1982). An average elastic thickness is 10 kilometers, which corresponds to a flexural wavelength of 65 kilometers.

The deflection across the basin, $w$, is defined by the following equation:

$$w(x) = \frac{V_0 \, \alpha^3}{8 \, D} e^{-x/\alpha} \left( \cos \frac{x}{\alpha} + \sin \frac{x}{\alpha} \right) \quad (10)$$

$w$ is the depth at the distance $x$ from the load, and $V_0$ is the line load applied at $x = 0$. 
All subsidence in this study is due to flexure loading of sediment filling a pre-existing water column. Because the basin fill simulator is 2-dimensional, it assumes that events within the section occur equally in all section's perpendicular to the section. This is reasonable for a passive-margin setting over long time-steps (>100 ky), but less reasonable for a foreland basin or carbonate atoll setting. A 2-dimensional subsidence history specified within the simulator that closely reproduces the geometries in the stratigraphic cross-section, is closer to the real subsidence history that generated the stratigraphic record than to the backstripped subsidence-history.

ANALYSIS OF STRATIGRAPHIC SECTIONS

A one-dimensional subsidence analysis on two sections with BasinWorks (Figs. 5 and 8) demonstrates the effect of filling in 2-dimensional space. The location at 25 km was filled before the initiation of the model and demonstrates how subsidence occurs after the section fills due to loading within the flexural wavelength of the crust. The location at 62.5 km initially has 250 meters of water depth and demonstrates how two-dimensional flexural compensation is expressed in a one-dimensional analysis.

In an initially filled location, one-dimensional subsidence analysis calculates an uplift due to the migration of the bulge, followed by exponential decline after the boundary between the sloping fluvial wedge and flat coastal plain progrades past the location (Fig. 8). The magnitude increases with increasing paleobathymetry of the proximal basin (Fig. 9). The deviation depends on at least four factors: (1) the distribution of the sediment package, (2) distance from the depocenter, (3) crustal rigidity (Fig. 8), and (4) initial maximum paleobathymetry of the proximal basin (Fig. 10). Varying the elastic thickness from 1 to 32 kilometers demonstrates that the width of the flexural response increases from 10 to more than 100 kilometers (Fig. 8).
Figure 8. Calculated tectonic subsidence at an initially filled site located at 25 km of model results with a 1, 2, 4, 8, 16, 32 km effective elastic thickness. The dashed lines are the sediment surfaces, the solid lines are the total subsidence histories, and the dotted lines are the calculated tectonic subsidence histories. This initial maximum paleobathymetry of the basin is 250 meters.
Figure 9. Calculated tectonic subsidence from a initially filled site located at 25 km of model results with a 8 km effective elastic thickness. The dashed lines are the sediment surfaces, the solid lines are the total subsidence histories, and the dotted lines are the calculated tectonic subsidence histories. This initial maximum paleobathymetry of the basin is 250, 500, and 1000 meters.
Analysis of the results at 62.5 km (Fig. 5) indicates three phases of tectonic activity when the site was stable: (1) early subsidence during infilling of distant locations and relative starvation at the location; (2) uplift during infilling due to sediment stacking at the location; and (3) subsidence during infilling of distant locations and filling above sea level. When the calculated tectonic-subsidence curve is smoothed to approximate thermal contraction, an apparent sea level fluctuation is incorrectly inferred in three phases: (1) an early 25-meter rise; (2) 75-meter fall; and (3) 50-meter rise, when in fact are no tectonic or sea level changes. These results suggest a portion of the tectonic subsidence is recorded at the time of activity and another portion when the sediment supply is available to fill the water column.

After the basin fills to sea level, the total subsidence is a function of the initial paleobathymetry and is independent of elastic thickness, given a constant sediment density. With increasing paleobathymetry, the depocenter concentrates closer to the margin. This is demonstrated by the total subsidence of settings with an elastic thickness of 8 km and 250, 500, and 1000 meters of initial paleobathymetry (Figs. 11 and 12). When all the space fills to sea level within the flexural wavelength of the location, the column will no longer subside and the depocenter migrates basinward (Fig. 6). If sedimentation continues above sea level in a fluvial wedge, subsidence continues. When analyzing sections with a paleobathymetry of 0 meters, 1-D backstripping erroneously indicates that tectonic subsidence is still occurring, depending on the distance from the depocenter, paleobathymetry of the proximal basin (Fig. 11), and the elastic thickness (Fig. 10).

Subsidence events are dispersed overtime by the backstripping analysis. This occurs because a subsidence event is not completely recorded in the stratigraphic column until all the space is filled (created by the event) within a distance approximately twice the flexural wavelength of the crust at the locality (Fig. 7).
Figure 10. Distribution of total subsidence as a function of 1, 2, 4, 8, 16, 32 km effective elastic thicknesses.
Figure 11. Distribution of total subsidence for a crust with an effective elastic thickness of 8 km and initial maximum paleobathymetries of 250, 500, 1000 meters.
Figure 12. Calculated tectonic subsidence from a unfilled site located at 33 km of model results with a 8 km effective elastic thickness. The dashed line are the sediment surfaces, the solid lines are the total subsidence histories, and the dotted lines are the calculated tectonic subsidence histories. This initial maximum paleobathymetry of the basin is 250, 500, and 1000 meters.
COMPARISON OF ONE- AND TWO-DIMENSIONAL DATA

Problems with geohistory analysis make it difficult to assign too much importance to short-term variations (<5 million years). Along with problems with paleobathymetry and age dating, one-dimensional backstripping analysis contains errors because it compensates loads locally. Total compensation at the location of sediment and water loading assumes the crust has no strength and that events recorded within a stratigraphic section apply equally in all directions. In most cases, the sediment supply cannot fill all the space created during a subsidence event and a transgression occurs. When subsidence decreases or the region is rising, sediment rapidly fills the basin and a regression occurs.

One-dimensional subsidence analysis is over-sensitive to changes in water depth. When sediment replaces water, a local isostatic adjustment predicts the basement experiences subsidence due to the new load. In reality the crust does not subside the full amount predicted by a local isostatic adjustment unless the depositional event is regional in extent. More often, depositional events occur in shallow water depths (<100 meters), are short-term (<1.5 million years), and are distributed over a small area compared to the flexural wavelength of the crust. Instead, the flexural strength of the crust reduces or delays the subsidence associated with the load.

In some basins, the tectonic subsidence occurred long before the major sediment infilling phase. In the Gulf of Mexico, much of the space is due to thermal subsidence before 100 Ma, whereas most of the clastic sediment supply did not arrive until the last 30 My. A one-dimensional backstripping algorithm attributes most younger sediment loading to contemporaneous tectonic subsidence.

In the Lamar Hunt McMahon No. 1 well, Jasper County, Texas, calculated subsidence rates reflect the sediment-supply history as much as the tectonic-subsidence history (Fig. 1). The Lamar Hunt well shows an increase in subsidence rates at 94, 80, 55, and 45 million years ago. The event at 94 Ma correlates with peak subsidence rates in
the Western Interior of North America and the initiation of the East Texas Basin. The event at 80 Ma corresponds to local volcanism and the initiation of the Laramide orogeny. The event at 55 Ma corresponds to the collision of Cuba and North America. The event at 45 million years corresponds to the collision of India with Asia, which caused a change in Pacific-North America spreading rates (Engebretson, 1982) with documented uplift events in the Southern Rocky Mountain region and their associated pulses in sediment in the Gulf coast region (Galloway, 1990).

An example from a Cretaceous carbonate platform margin exposed in the Vercors region of southeastern France demonstrates over-estimating tectonic uplift (Fig. 13) (Jacquin et al., 1991). One-dimensional backstripping analysis subtracts the locally compensated deflection from the total subsidence, and indicates a tectonic uplift from 117 to 116 Ma. In this case, the uplift most likely occurred because total subsidence also decreased then. However, the uplift rates are between two and five times too great when the section is modeled in the basin fill simulator. Local loads in the plane of section are regionally compensated for in a simulation including flexural loading. Therefore, the tectonic-subsidence history generated by the backstripping method must be adjusted before the basin fill simulation can reproduce the observed stratal patterns.

CONCLUSIONS

Subsidence analysis provides a first approximation of the tectonic-subsidence history of a section. However, pitfalls to one-dimensional subsidence analysis must be considered when analyzing the results for possible eustatic variations or tectonic events. The complete expression of a subsidence event does not occur until the space created by the subsidence, within a distance approximately twice the flexural wavelength of the location, is filled with sediment. Likewise, complete expression of an uplift does not occur until all the rock placed above base level is removed. When changes in paleobathymetry occur, one-dimensional subsidence analysis over-emphasizes the
variations. Subsidence analysis of model results from a two-dimensional simulation (without subsidence or eustatic changes, where paleobathymetry is known), erroneously
calculates tectonic-activity with apparent subsidence during low rates of sedimentation, and apparent uplift during high rates of sedimentation.

Analyzing a stratigraphic section to determine a sea level curve may have an error in sea level or tectonic subsidence that is as much as ±25% of the paleobathymetry. The timing of the error is dependent upon the delivery of sediment to the basin. The results of this comparison between one and two dimensions indicate three phases of apparent tectonic activity during infilling when there are none: (1) early subsidence during infilling of distant locations and relative starvation; (2) uplift during infilling due to sediment stacking; and (3) subsidence during infilling of distant locations and filling above sea level. It is necessary to correct for these inaccuracies when the results of the analysis constrain the subsidence history of a stratigraphic simulation.
Chapter 5.

Early Guadalupian (Permian) Sequence Stratigraphy of the Last Chance Canyon, Guadalupe Mountains, New Mexico: Quantified with a Computer Simulation

ABSTRACT

Simulation of the Last Chance Canyon section demonstrates how stratal patterns are a product of the interaction of (1) a dynamic depositional system of carbonate production, (2) periodic influxes of siliciclastic sand, and (3) a relative sea level curve. The relative change of sea level curve is composed of slowly changing subsidence rates and a rapidly changing eustatic sea-level history that contains two approximately 1 million year and several higher periodicity (>=100) cycles. A regional tectonic subsidence rate of 10 mm/ky was calculated by backstripping a composite well from the region. Simulation results from PHIL (Bowman and Vail, 1993) reproduced the stratal patterns within 5% of the documented geometries.

This region offers a classic section for displaying prograding clinoforms and is important for illustrating the development of depositional components and systems tracts that build incomplete sequences within complete sequences with "third-order" duration. During deposition, the fall of the third-order eustatic cycle enhanced progradation and impedes backstepping. This fall was synchronous with an increase in the rate of siliciclastic sediment influx. Most of the siliciclastic sediment is delivered when the third-, fourth- and fifth-order contributions to eustasy are falling. Siliciclastic sediment supply is approximately a function of the rate of relative sea-level fall.

The initial bathymetric profile in this region is shallower than earlier or later shelf margins. This may be due to 1) a rapid transition from large (>100 m) to low (<50 m) magnitude eustatic cycles that occurred at the end of the Leonardian or 2) an earlier second-order eustatic rise. The drowned shelf and deep basin fill within 150-200 meters
of water depth during the early Guadalupian. Unlike the subsequent Capitan fore slope
that reached 600 meters in height, the shallow bathymetry increases the rate of
progradation and minimizes the potential for developing gravity-flow deposits.

We present field examples of a scheme for classifying sequences based upon the
completeness of the sequence. The basic units of sequences are stratal components.
Components consist of prograding aggrading or backstepping intervals, basin-floor
fans, slumps or channel-levee lobes. Components are the smallest correlatable units.
They resemble systems tracts because they have aggradational-progradational patterns.
However, they terminate at a significant change in accommodation. A component
forms during either the rising or falling half of a cycle. Depending on the position of
the depositional interface with respect to wavebase, the base of a flooding component
may be erosional (ravinement) or depositional. Stratal component groups contain a
series of repetitive components. Stratal component groups stack in a series to form a
systems tract. Systems tracts include the lowstand, transgressive, highstand and shelf
margin (Posamentier and Vail, 1988). Sequences and stratal component groups form
during a base-level cycle. A depositional sequence contains a complete suite of systems
tracts. A Type I sequence contains a lowstand, transgressive and highstand systems
tract. A Type II sequence contains a shelf-margin, transgressive and highstand systems
tract. An incomplete sequence is missing at least one systems tract. They form in
conditions such as a long-term relative fall that prevents the development of the
transgressive systems tract, or a rapid relative rise that prevents the formation of
lowstand deposits.

INTRODUCTION

In this paper we present the results of simulating a detailed stratigraphic cross
section through a mixed carbonate/siliciclastic basin margin. This allowed us to
calculate the absolute magnitude, duration, and path of eustatic, tectonic, and sediment-
supply variations. Two-dimensional forward simulation offers a powerful platform for
deriving an integrated model and obtaining plausible values for process rates. The simulation gains its strength by integrating information from many different disciplines and scales including high-resolution seismic studies, modern process studies, and sequence stratigraphic studies of ancient rocks. By building the large picture while recording small details, the simulation can allow the geologist to "see the forest and the trees." We can never expect a computer simulation to displace field observations, but it may help us formulate questions that provide well-defined reasons to re-examine or collect more data.

In this study, we apply a simulator (PHIL) to constrain the geologic factors that produced the middle Permian stratigraphy exposed in Last Chance Canyon of the Guadalupe Mountains, New Mexico. The section formed the margin of the Delaware basin of New Mexico and western Texas. We simulate the stratigraphy documented by Sonnenfeld and Cross (1992) in a cross-section parallel to stratigraphic dip (Fig. 1). Comparisons of the stratigraphic section with results from PHIL show a close tie to stratigraphic geometries (Fig. 2), lithofacies distributions (Fig. 3), and the timing and location of erosion. Comparison of the geometry of stratigraphic horizons with model results (Fig. 4) also demonstrates the present capacity of the simulator to reproduce stratigraphic patterns and lithologies by imposing an appropriate eustatic, tectonic subsidence, and sediment-supply history (Fig. 5). Synthetic stratigraphic sections extracted from the simulation results are similar to measured stratigraphic sections (Fig. 1). Computer analysis of the sequence stratigraphy illustrates the results of applying sequence stratigraphic rules for defining sequence boundaries, maximum flooding surfaces, and top of lowstand prograding complexes to a situation containing higher-order cyclicity (Fig. 6). Depositional systems show a high-order alternation between shelf deposition and slope/basinal gravity-flow deposits during upper lowstand progradation (Fig. 7). The primary porosity state at the end of deposition can be inferred from the porosity
Figure 1. Stratigraphic cross-section through the Last Chance Canyon (Sonnenfeld, 1991).
<table>
<thead>
<tr>
<th>Facies Key</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fissile (wackestone) / mudstone</td>
</tr>
<tr>
<td>Laminated graded packstone / mudstone / wackestone</td>
</tr>
<tr>
<td>Allopaic skeletal wackestone / packstone</td>
</tr>
<tr>
<td>Cherty intraclast rudstone / floatstone</td>
</tr>
<tr>
<td>Nodular lime rudstone / wackestone / mudstone</td>
</tr>
<tr>
<td>Cherty microskeletal-spicule wackestone</td>
</tr>
<tr>
<td>Crinoid-bryozoan wackestone / bafflestone</td>
</tr>
<tr>
<td>Very silty brachiopod-sponge boundstone / wackestone</td>
</tr>
<tr>
<td>Allopaic coated grain packstone / grainstone</td>
</tr>
<tr>
<td>Cherty brachiopod-fusulinid wackestone</td>
</tr>
<tr>
<td>Fusulinid grainstone / packstone</td>
</tr>
<tr>
<td>Massively intraclast-skeletal-peloidal floatstone / rudstone</td>
</tr>
<tr>
<td>Bioturbated wavy-bedded lime mudstone</td>
</tr>
<tr>
<td>Massive dolomite--peloid wackestone / packstone</td>
</tr>
<tr>
<td>Swaley-bedded peloid-coated grain grainstone</td>
</tr>
<tr>
<td>Coated grain-oolitic grainstone</td>
</tr>
<tr>
<td>Fenestral laminite (mudstone to peloid packstone)</td>
</tr>
<tr>
<td>Channelized skeletal-peloid turbidite sandstone</td>
</tr>
<tr>
<td>Thin bedded fusulinid-peloid turbidite sandstone</td>
</tr>
<tr>
<td>Bioturbated fusulinid-peloid dolopackstone-wackestone / sandstone</td>
</tr>
<tr>
<td>Glaucinitic peloidal muddy siltstone</td>
</tr>
<tr>
<td>Thin bedded channelized skeletal turbidite sandstone</td>
</tr>
<tr>
<td>Churned very fine grained silty sandstone</td>
</tr>
<tr>
<td>Sigmoidally trough cross-stratified fusulinid sandstone</td>
</tr>
<tr>
<td>Swaley-bedded, parallel laminated to bioturbate very fine grained sandstone</td>
</tr>
</tbody>
</table>
Figure 2. Simulated stratal history of the Last Chance Canyon cross-section.
Figure 3. Simulated lithofacies of the Last Chance Canyon cross-section.
Facies of the Last Chance Canyon Section
Figure 4. Comparison of simulated sections and stratigraphic sections in Last Chance Canyon.

The black lines are the depths of the observations. Vertical strips indicate where the model is higher than...
Observations.
The results are higher than the observation.
Figure 5. Relative sea-level, eustatic, and tectonic subsidence history from the Last Chance Canyon region.
Figure 6. Simulated sequence stratigraphy of the Last Chance Canyon cross-section.
Stratigraphy of the Last Chance Canyon Section

Lower Lowstand

Prograding Complex
Simulated Depositional Systems

Figure 7. Simulated depositional systems of the Last Chance Canyon cross-section.
Figure 8. Simulated porosity of the sediments at the time of deposition for the Last Chance Canyon cross-
For the Last Chance Canyon Section

canyon cross-section.
depth functions for each lithology (Fig. 8). A chronostratigraphic diagram shows the modeled relationship between tectonic subsidence, eustasy, siliciclastic sediment influx, coastal onlap, and sedimentation in the deep basin (Fig. 9). The paleobathymetry plot (Fig. 10) shows that onlap within carbonate sediment is most often marine onlap (Melton, 1947). In general, siliciclastic sediments form marine onlap during gravity-flow sedimentation and coastal onlap during shelf sedimentation. Non-marine siliciclastic onlap is limited to the overlying Grayburg Formation deposited after a major seaward shift in coastal onlap. A seismic model (Fig. 11) illustrates the seismic reflection pattern the might be obtained under ideal conditions.

**GENERAL GEOLOGY**

The stratigraphy of Last Chance Canyon, Guadalupe Mountains, New Mexico (Fig. 12) consists of upper Leonardian through lower Guadalupian shelf margin and basinal carbonates inter-fingered with basinal turbidites and thin sandstone intervals on the shelf (University of Wisconsin carbonate research group (1986); Kerans and Nance, 1991). Geology of Last Chance Canyon was previously studied by Boyd (1958), Hayes (1959), Harrison (1966), and McDermott (1983) among others. Sarg and Lehmann (1986) published the first sequence stratigraphic framework of the San Andres Formation.

The modeled section in this study is based on a detailed stratigraphic section through Last Chance Canyon (Sonnenfeld, 1991a, b; Sonnenfeld and Cross, 1992). This section exposes offlapping clinoforms containing three sequences characterized by fusulinid, bryozoan grainstone, and sponge-spicule mudstone deposition (Fig. 1). The Glorieta Sandstone is the oldest interval exposed. This unit is a shallow-marine transgressive sandstone equivalent to the Yeso Formation. The Glorieta is overlain by the Cutoff Formation, a siliciclastic-starved transgressive limestone. These units might be the basinal equivalent to the Pipeline Shale. The San Andres Limestone lies above the Cutoff Limestone. The middle of the San Andres Limestone marks the beginning of
Figure 10. Simulated paleobathymetry of the sediments at the time of deposition for the Last Chance Car...
Asymmetry of the Last Chance Canyon Section

10 11 12

Last Chance Canyon cross-section.
Figure 11. Seismic response of the simulated lithofacies distribution. Velocities were calculated from the porosity and density response and is similar to that produced by Vibroseis acquisition techniques.
Porosity and density of the rock. The resulting impedance contrasts were convolved with a zero-phase 5-40 Hz Klauder wavelet. This
Figure 12. Regional stratigraphic cross-section through the Delaware Basin margin in the vicinity of Last Chance Canyon (after Sarg and Lehman, 1986; University of Wisconsin carbonate research group (1986); Kerans and Nance, 1991). The location of the detailed cross-section is marked in the box.
a large siliciclastic influx. Within the basin, up to three-hundred meters of the Brushy Canyon Sandstone overlay the Cutoff Limestone and its basinal equivalents. These turbidites onlap out on the basin margin. The Cherry Canyon Sandstone interfingers with the upper part of the San Andres Limestone. Capping the shelf margin of the San Andres Formation is the Hayes Sandstone. This unit is a shallow marine transgressive sandstone. The datum in this study is the Hayes Sandstone. The Grayburg Limestone lies above the Hayes Sandstone.

Prior to Guadalupian sedimentation, the lowstand platform margin was 10 km farther to the southeast. Guadalupian deposition occurred on a flooded shelf (Fig. 12). The flooding of the shelf may be due to an antecedent period of high-magnitude sea-level fluctuations (>150 meters, Bowman et al 1990) that produced extensive lowstand shelves. This shallow bathymetry may decrease the amount of gravity-flow sedimentation and increase the rate of carbonate sediment production and associated rate of progradation. The flooded shelf setting is observed in other early Guadalupian sections in the Kupfersheifer interval of Holland and in the Himalayas of Pakistan (Vail personal notes).

**CHRONOSTRATIGRAPHIC FRAMEWORK**

The Lower Cherry Canyon Tongue is lower Guadalupian in age and the three sequences (uSA3, 4, 5) in the section belong to the lower part of the Guadalupian. Lower Guadalupian fusulinids and conodonts have been found in the Last Chance Canyon region in the lower limestone's below the Lower Cherry Canyon Sandstone Tongue (Wilde, 1986; Wilkinson et al., 1991).

The absolute time scale for the Permian (Fig. 13) is imprecise and changes with each revision (Harland et al., 1982; Ross and Ross, 1987; Harland et al., 1990; and Wilde, 1990). The degree of precision of the ages used in this study cannot be supported, these ages are only used to identify events within the model. Large differences in absolute age imply the duration of the sequences and processes could
change by a factor of two. The shape of the relative sea level curve should be consistent and simply expand or compress with changes in the absolute ages. If the actual duration is shorter, all sedimentation, tectonic subsidence and erosion rates must increase accordingly. The eustatic histories must also compress into the new interval. However, the shape of the relative sea level curve, including the magnitudes of the sea level falls, should not change. Likewise, if the orientation of the stratigraphic cross-section was oblique to the true progradation direction, the actual sediment production rates would be less but the shape of the curve should remain approximately the same.
For this model, we prefer to use the longer duration assigned by the Ross and Ross (1987) time scale. During the Leonardian and Guadalupian at least 17 sequences formed over a span of 20-22 million years (My) which averages to approximately one sequence per 0.75-0.85 My. The model was scaled so the sequence that is completely recorded in the canyon has a duration of one million years (257.7 to 256.7 Ma). With this time-scale the high-frequency cycles are more consistent with Milankovitch ages and bundling. The model has a duration of 1.74 My from 258.1 Ma to 256.36 Ma, and is sampled by 202 layers with a time step of 8.5 ky. The Nyquist frequency (the highest frequency the model can resolve), therefore, is 0.0588 cycles/ky or a period of 17 ky/cycle. This is higher resolution than the 19 and 21 ky/cycle (0.047 cycles/ky) commonly associated with the precession of the equinox.

**STRATAL COMPONENTS, COMPONENT GROUPS, SEQUENCES AND INCOMPLETE SEQUENCES**

We proposed a stratigraphic framework where tectonic, eustatic, and sediment supply cycles build a hierarchy of packages that compose the stratigraphic record and differ in onlapping relationships, stacking patterns, and internal character (Bowman and Vail, 1994b). The hierarchy is as follows:

1. continental encroachment megasequence;
2. transgressive-regressive supersequence;
3. sequence -- Type I, Type II, and incomplete;
4. systems tract -- lowstand, shelf margin, transgressive, and highstand;
5. stratal component groups; and
6. stratal component.

Each package is composed of the subordinate packages. The following scheme classifies stratigraphic intervals based on the accommodation recorded in the interval (Fig. 14).

As suggested in earlier work (Bowman and Vail, 1994b), we propose that the basic rock unit of a stratigraphic interval is a stratal component. This is the stratigraphic
package that PHIL generates in each time step. A component may exhibit either backstepping, aggrading, forestepping, or degrading accommodation and is bounded by flooding and cutting surfaces. Components include downstepping (d), forestepping (f), upstepping (u), and backstepping (b) intervals, basin-floor fan lobe (f), slump blocks (s), erosion (e), and channel-levee lobe (o). Depending on the position of the interface with respect to wavebase, the base of an aggrading component may be erosional (ravinement) or depositional.

Figure 14. Components form in response to relative changes in sea level. Backstepping, upstepping, and forestepping form in response to a relative rise in sea level. Downstepping components form during a stillstand or small relative rise but are deposited between two components formed during relative falls. Levee-channel lobes, slumps, and basin floor fan lobes form in response to a relative fall in sea level.
Component's group into upstepping-backstepping (u-b, the most downward direct component is listed first), forestepping-upstepping (u-f), downstepping-forestepping (d-f) pairs to form stratal component groups.

Stratal component groups into trends of similar accommodation patterns bounded by major through-going surfaces. These surfaces typically form systems tract boundaries. The upper lowstand systems tract changes from d-f to f-f. However, the shallow-marine components are commonly eroded during subsequent downstepping. The late lowstand systems tract changes from f-u to f-b. The transgressive systems tract changes from u-b to b-b. The highstand systems tract changes from f-u to f-d.

Sequences are composed of a complete series of lowstand, transgressive and highstand systems tracts. An incomplete sequence is missing one or more systems tracts. High rates of sediment supply, subsidence or uplift may cause a eustatic cycle to form a complete sequence in one location and an incomplete sequence in another location. A sequence may be complete in one location but under a different set of circumstances may be incomplete.

When several sequences form a lowstand, transgressive and highstand pattern, they group into a supersequence. Likewise, when several supersequences form a lowstand, transgressive and highstand pattern, they group into a megasequence.

MATCHING OF SIMULATION RESULTS WITH SECTION

The primary controls (in parentheses) were adjusted to match the following depositional features:

1) gradients of the stratal lines (geometry of depositional interfaces, erosion),
2) lithofacies distribution (sediment supply, depositional system),
3) stratal path (sea-level, sediment supply),
4) location of downlap and onlap terminations (sea-level, sediment supply),
5) thickness of intervals (tectonic subsidence rates, sea-level, sediment supply),
6) sequence boundary geometry (sea-level, depositional system),
7) erosional surfaces (sea-level, depositional system), and
8) number and duration of dominant and high-frequency packages (sea-level).

These features define the physical state of this cross-section. Derivation of the principal model parameters is discussed in the following section.

The computer simulation results include a series of time steps captured during the development of the model include the initial unconformity (Fig. 15), before the formation of the G2 sequence boundary (Fig. 16), the top of slope fan surface (Fig. 17), the maximum flooding surface (Fig. 18), a Type I sequence boundary during the upper lowstand systems tract (Fig. 19), and formation of the unconformity at the beginning of the slope fan complex or late lowstand prograding complex (Fig. 20).

Application of Simulator

A relative sea-level curve that reproduces the stratal geometries and lithofacies distributions within Last Chance Canyon was generated by adjusting the sea level history until the stratal patterns within the cross-section minimize the difference between the observed stratal patterns and those in the model results. We focused on matching the position of the offlap break through time (stratal path). Eustasy was adjusted to create the high-frequency vertical changes in the stratal path. A constant rate of tectonic subsidence controls the depths and thicknesses of the intervals. The horizontal differences between the observed and modeled stratal paths are attributed to differences in age or changes in siliciclastic sediment supply. The average time increment between data values is 9 ky. Most sea level values have been adjusted to within 0.5 meter.

Siliciclastic sediment supply was adjusted to reproduce the timing and cross-sectional area of siliciclastic sediment depicted in the original cross-section. Siliciclastic sediment is not present in most aggrading component groups. There are no down stepping components preceding the aggradational components. The gravity flow components, composed of both siliciclastic and carbonate sediment, onlap the sequence
Figure 15. Configuration at the time of the initial depositional profile (258.1 Ma).
Species of the Last Chance Canyon Section

<table>
<thead>
<tr>
<th>10</th>
<th>11</th>
<th>12</th>
</tr>
</thead>
</table>
Simulated Lithofacies of

Figure 16. Configuration before the formation of the G2 sequence boundary (257.79 Ma).
Biofacies of the Last Chance Canyon Section
Figure 17. Configuration at the formation of the top slope fan surface (257.664 Ma).
Interpretations of the Last Chance Canyon Section
Figure 18. Configuration at the formation of the maximum flooding surface (257.594 Ma).
Facies of the Last Chance Canyon Section
Figure 19. Configuration at the formation of G3, a type 1 sequence boundary at the beginning of the lowstand
acies of the Last Chance Canyon Section

slowstand systems tract -- upper prograding complex (257.455 Ma).
Simulated Lithofacies of

Figure 20. Configuration at the formation of the unconformity at the beginning of the lowstand systems tract.
Systems tract -- late prograding complex (256.7 Ma).
boundaries. Therefore, the siliciclastic sediment is deposited during relative sea-level falls. Carbonate production is removed by gravity-flow processes as well.

Stratal patterns can be understood in terms of their relationship to a relative sea-level and sediment supply history. The primary observation that defined this history is the stratal path. The stratal path is the present position of the offlap break through time. The offlap break is the point where the clinoform begins to rollover to the steeper slopes of the depositional front. It has been suggested that the offlap break commonly corresponds with fairweather wavebase, which is typically at 10 meters water depth (Vail, 1990). However, in Last Chance Canyon the offlap break formed at 20 meters. The presents of fusulinid grainstone/packstones (Fig. 1) at this depth suggest that it is above storm weather wavebase, suggesting this was a high-energy coast. The initial bathymetric profile, relative changes of sea-level, sediment supply, and post-depositional deformation control the stratal path. The stratal path and location of onlap and downlap termination's are linked to relative changes of sea level by the depositional interface.

The relative fall of sea-level produces predictable changes in the ratios of systems tract volumes within a sequence. Compared to earlier sequences in the Permian and to other settings, the lowstand prograding complex and highstand systems tracts are poorly developed. Low subsidence rates or falling base-level reduces or eliminates the capacity for the system to transgress by enhancing relative fall and the seaward partitioning of the siliciclastic sediment supply to the platform margin and basin. The results have bearing on sequence stratigraphic interpretations such as the position of sequence boundaries and maximum flooding surfaces.

**Stratal Path**

From the G1 sequence boundary to the G2 sequence boundary the stratal path is characterized by upward decreasing aggradation and increasing rates of progradation. The thicknesses and stratal path are matched in the model by adjusting the siliciclastic
sediment supply and sea level. From the G2 sequence boundary until surface 2 of the
G2 sequence, no carbonate or siliciclastic component preserved an offlap break.
Abundant truncation along the G2 sequence boundary and a major influx of siliciclastic
sediment lying basinward indicates the margin was intensely eroded by slumping,
active storm and fairweather wave processes. It was probably unstable for shelf
deposition at this time. Termination of higher frequency stratal lines below G2 is
matched with minor shoreface erosion. Onlap above the G2 and G3 sequence boundary
is marine onlap. The stratal lines terminate above marine sediment (Fig. 10). The
dolomitic peloid wackestone facies in the overlying components below surface 3
suggests that the location was below sea-level, and the stratal termination's are marine
onlap. The carbonate profile produces this relationship correctly. The onlap is 31
meters below the offlap break of the previous highstand. The onlap is a function of the
interaction between the bathymetry on the sequence boundary, the crustal response to
sediment and water loading during the fall, and the geometry of the depositional
interface. In this case, the position of onlap was produced by a relative fall of 32.5
meters. The additional 1.5 meters is necessary to remove the space that was created by
tectonic subsidence and flexure loading.

The next three surfaces (4, 5, and 6) exhibit progradation and increasing rates of
aggradation. Surface 4 onlaps out at the offlap break of surface 3. A reduction in the
rate of siliciclastic sediment influx continues during the formation of surfaces 3 through
5. The stratal path between surfaces 5 and 6 indicate increasing rates of
accommodation. During deposition of this component the depositional interface
continued to prograde approximately 50 meters (0.58 meters/ky).

Surface 6 is the G3 sequence boundary. Above this surface, gravity flow
deposits are prevalent. This is in response to an increase in the siliciclastic influx rate
accompanied by Type I relative falls of sea level. The rollover at the offlap break on
surface 7 is not as sharp as surface 5 and 6. Surface 7 shows a slight retreat of the offlap break.

The offlap breaks on surfaces 7 through 15 show very high rates of progradation (3.2 meters/ky) with minimal aggradation. It is difficult to determine the amount of topset erosion. However, the consistent depth of the offlap breaks below the toplap surface suggests that erosion was less than ten meters. The progradation is also accompanied by high-frequency sea-level falls that delivered large volumes of siliciclastic sand to the basin. The sand fills the lower portions of the clinoform and provides a shallower base that enhances carbonate progradation. However, the siliciclastic influx also reduces carbonate production, inhibiting progradation. When the section was modeled without siliciclastic sediment, the distance of shelf progradation was equal (Fig. 21). This suggests the volume of sediment stored along the margin is dominantly controlled by the carbonate growth-rate. Sediment is transported to the basin when more sediment is delivered to the shelf margin than can be stored.

At the lowstand in sea level (surface 15), at the end of the upper lowstand, sediment begins to onlap the unconformity that formed during the fall.

**Lithofacies Distribution**

The lithofacies distribution is a function of the interaction of sediment supply and the depositional system with relative changes of sea-level and bathymetry. During the deposition of carbonate components, base-level is rising and the stratal patterns build upward and outward. Siliciclastic sediment is rare in these intervals. The correlation between relative falls of sea level and siliciclastic sediment supply suggests that a relative fall of sea-level is necessary to deliver siliciclastic sands to the basin margin (Fig. 22).

At present, outcrop facies distributions are specified by the textures and types of biota (Sonnenfeld and Cross, 1992). Simulated facies are defined by grain size,
Simulated Lithofacies of t

Figure 21. Last Chance Canyon cross-section without siliciclastic sediment.
mineralogy, and texture. Biota observed in Last Chance Canyon include algae, fusulinids, brachiopods, bryozoans, and sponges. The lithofacies distributions on the basin margin are controlled by energy regimes that roughly correspond to paleobathymetry and shelf physiography. The dolomitic peloid wackestone forms in the "reef" interface between the lower tidal range and the fairweather wavebase. The fusulinid wackestone/packstone/ grainstone interval corresponds to the "fore slope" between fairweather and storm weather wavebase. The very silty brachiopod-sponge boundstone/wackestone corresponds to the basal "fore slope" designated as fine grainstones.

The siliciclastic and carbonate sediment components are plotted as two separate intervals in the lithofacies plot. Carbonates are plotted on the lower layer. More realistically, it would be better to plot them as a mixture but it is nearly impossible to illustrate the infinite gradation of the mixture. During deposition of a component group, flooding produces a siliciclastic-free, carbonate-prone environment, whereas falling produces an influx of siliciclastics, which reduces carbonate production.
A lack of siliciclastic sediment in the basal portions of clinoforms above the G2 sequence boundary forms a significant departure of the model from the observations. This sand is deposited below the offlap break, during relative falls of sea level but within clinoforms, not as basin floor fans, slope fans, or slumps. At present it is not possible to introduce siliciclastic sediment directly on the slope without bypassing the shelf when there is space on the shelf for deposition. This indicates that siliciclastic sediment may be transported across the shelf through submarine incised valleys and channels but is stabilized in the fore slope of the clinoform.

**Seismic Response**

The seismic response of the model results (Fig. 11) shows many steeply dipping reflectors. The condensed section on top of the slope fan forms the strongest and most continuous reflector. The G2 sequence is very subtle. However, the G3 sequence boundary is clear and is defined by onlapping surfaces above a steeply dipping reflector.

Loss of high frequency information in typical seismic data through this interval is due to 1) larger fresnel zone, 2) higher velocities, and 3) lower frequency content of transmitted wavelet. The deeper the section is buried, the wider the region the fresnel zone averages. The velocities predicted by the modeled lithofacies/porosity state are lower than typical measured velocities at 1 to 1.5 seconds of two-way travel-time. The faster velocities would compress the reflections into a shorter interval of time. The Klauder wavelet also contains an equal balance of frequencies, which adds more high-frequency information.

**DESCRIPTION OF VARIABLES**

In total, 80 variables define the siliciclastic, evaporite and carbonate depositional systems and erosion activity (Table 1). The carbonate and siliciclastic depositional interface define the geometry of stratal surfaces. Sediment supply controls the volume and type of sediment. Sediment supply and relative changes of sea-level control the
volume of stratal components, stacking patterns and erosive surfaces within the Last Chance Canyon section. Relative change of sea-level defines the distribution of stratal packages and timing of siliciclastic influx into the platform margin region.

Types and Geometry of Depositional Interfaces

The geometry of stratal lines in the Last Chance Canyon cross-section is controlled by interfaces specific to the carbonate and siliciclastic depositional systems and the wedge form of suspension and gravity-flow deposition. Specifications for the siliciclastic depositional interface remain constant throughout the model. Gradients for carbonate depositional interface are constant except the fore slope, which increases from 0.1 to 0.27 after the formation of the G2 sequence boundary (Table 1). This increase could be due to a change in the cementation or depositional systems. More likely, it is due to a constant gradient with a change in the angle of progradation from 20° to the cross-section to oblique. Regional studies support a change in progradation from south-southeast to southeast at this time (W. Tyrell, pers. comm.).

Table 1. Values used for the depositional systems in the simulation of Last Chance Canyon.

<table>
<thead>
<tr>
<th>Type</th>
<th>Gradient</th>
<th>Degrees</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonate</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land surface</td>
<td>0.000001</td>
<td>0.00°</td>
</tr>
<tr>
<td>Tidal flat</td>
<td>0.0001</td>
<td>0.00°</td>
</tr>
<tr>
<td>Back Reef</td>
<td>0.023</td>
<td>1.3°</td>
</tr>
<tr>
<td>Fore slope</td>
<td>from 258.1 to 257.5 Ma</td>
<td>0.08</td>
</tr>
<tr>
<td></td>
<td>from 257.5 to 256.1</td>
<td>0.25</td>
</tr>
<tr>
<td>Siliciclastic</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fluvial plain</td>
<td>0.00003</td>
<td>0.001°</td>
</tr>
<tr>
<td>Coastal plain</td>
<td>0.00001</td>
<td>0.0006°</td>
</tr>
<tr>
<td>Shoreface</td>
<td>0.01</td>
<td>0.57°</td>
</tr>
<tr>
<td>Depositional front</td>
<td>0.02</td>
<td>1.1°</td>
</tr>
</tbody>
</table>

The tidal flat, lagoon, reef, and fore slope gradients define the carbonate depositional interface. These names are generic through geologic time and do not necessarily imply specific facies. The tidal flat gradient (0.0001) is very shallow because tidal flat sediments are observed approximately 25 kilometers up depositional
dip from the Last Chance Canyon region on the Algerita's Escarpment. The geometry of the lagoon is a response to production and depositional processes and is not defined. The lagoon consists of a broad shallow sea between the higher relative growth rates on the carbonate shelf-margin that isolate the broad shallow sea between the margin and the tidal flat. We have not attempted to study the tidal range and have set it arbitrarily at 1 m. We define the average gradient measured just landward of the offlap break as the back reef interface gradient (0.023). The transition width from the reef to fore slope gradient, or rollover width, is 1.2 km to approximate the roundness of the offlap break on a carbonate stratal surface. The reef-top depth is 20 meters of water-depth and corresponds to fairweather wavebase. We assign the maximum gradient measured on a carbonate stratal surface, seaward of the offlap break, to the fore slope gradient (0.25). The fore slope gradient increases from 0.08 to 0.25 after formation of the G2 sequence boundary. This is much steeper than a typical siliciclastic depositional front (0.065). In this model, very little carbonate sediment deposited on the carbonate platform slope is transported basinward by slumping processes. This is because carbonate sediments are stable on shallow gradients formed by siliciclastic sediment at the base of each component group.

We define the siliciclastic depositional interface with a fluvial plain, coastal plain, shoreface, and depositional front. Gradients within the siliciclastic depositional interface are slightly steeper than mud-dominated deltaic settings (Table 1). The fluvial (0.00003) and coastal plain (0.00001) gradients are empirically set very shallow. The coastal plain width is arbitrarily set at 15 km because no information is available. These values do not effect the results because the model edge is 7 kilometers from the cross-section and fluvial sediments are not observed within 35 km of the region. The shoreface gradient was set to 0.01. This value is slightly less than the reef gradient and preserves the sand above the offlap break. The offlap break, which may correspond to fairweather wavebase (Vail et al., 1991), is set at 20 meters of water-depth to match the
observed maximum depth for the offlap break below a horizontal portion of the stratal line. Here the offlap break is deeper than fairweather wavebase because massive, well-sorted fusulinid grainstone indicate sediments near the rollover were deposited below fairweather wavebase and above storm wavebase. The facies below the horizontal portion of the stratal surface indicate water depths greater than zero. The transition from shoreface to depositional front environment (rollover width) is set at 1.2 km to match the roundness of the offlap break. The maximum stable gradient in the marine environment for siliciclastic sediment (called the depositional front) is set at 0.02. This is slightly shallower than a typical mud-dominated setting (0.065). Even with steeper gradients, most sediment deposited on the carbonate platform slopes are unstable and transported basinward by gravity-flow processes. The best approximation was obtained when the suspension distance is 4 km. This distance is short compared to settings with a large clay component (typically 25-55 km) suggesting, as expected, that the sand settles much faster than clay. Where siliciclastic sediment is deposited above a sediment interface too steep for coastal deposition, it is removed and deposited as slumps on an interface that dips less downslope.

The model produces gravity-flow deposits in the form of basin-floor fans and slope-fan complexes during a relative fall of sea-level with rates greater than 30 mm/ky in settings with a bathymetric contrast greater than a pre-defined bathymetric minimum (iteratively determined as 100 meters in this model). When this occurs, all sediment lying within the basin is removed and transported downward to the deep basin by gravity-flow processes. If the bathymetric contrast is less than the minimum required bathymetric contrast, the sediment remains on the platform margin.

**Sediment Supply**

In this simulation, we applied a sediment supply characterized by an in situ carbonate factory with constant production rates that are decreased by a variable siliciclastic source (Bowman and Vail, 1992). Carbonate sedimentation is a product of
carbonate production and redistribution. Carbonate production is characterized by four contributions including: 1) unrestricted traction-load (150 mm/ky), 2) shelf-margin traction-load (100 mm/ky), 3) unrestricted fine-grained suspension-load (170 mm/ky), and 4) pelagic sediment (0 mm/ky) that may produce a total combined peak accumulation rate of 420 mm/ky (Fig. 23). The shelf-margin production is restricted laterally as a normal distribution function (defined in equation 7, Bowman and Vail, 1994) centered at the first location from the open-basin with a depth equal to the depth of maximum shelf-margin production (20 meters in this model). The horizontal width of the normal distribution function is 1.5 km. The depth of maximum shelf production is 20 meters. The widths of the depth functions of unrestricted and shelf-margin production are 50 meters. There is no pelagic carbonate production during the Permian. Therefore the pelagic production rate is set to 0 mm/ky.

The rate of siliciclastic sedimentation and volume of suspended siliciclastic sediment passing through each location limits the rates of carbonate production. Any siliciclastic deposit that is subsequently removed by slumping or turbidite deposition also reduces carbonate growth. The role of siliciclastic sediment in reducing carbonate growth is defined by an upper limit of siliciclastic sedimentation (C) of 150 mm/ky, above which no carbonate production occurs. Siliciclastic sedimentation rates (S) below 150 mm/ky are reduced by a factor calculated with the following equation:

\[ R = 1 - \sqrt{\frac{S}{C}} \]

Here, R is the reduction factor. Concentrations of fusulinids within the sandstones indicate carbonate production continues during the relative fall but is removed by gravity-flow processes. Interaction of sea level and outer shelf/slope physiography dictated that the carbonate "factory" width contracted during relative falls of sea level (Sonnenfeld and Cross, 1992).

The siliciclastic supply is defined as 20% traction load sediment and 80% suspended sediments because there is minor evidence of shoreface or deltaic sand in the
Carbonate Productivity Functions

mm/ky

Depth (m)

Shelf Margin Production
Unrestricted Production
Fine-grained Production

Figure 23. Carbonate production rates used to produce carbonate sedimentation in the model.

cross-section. Fluvial-deltaic processes were subordinate to eolian and marine processes. This suggests that siliciclastic sediment is unstable along the shore. The model introduces pulses of siliciclastic sediment that are restricted to periods of relative fall of sea level (Fig. 22). The sorting and surface texture of the sand grains indicate they were transported to the basin margin by eolian processes (Fischer and Sarnelein, 1988). This suggests that the similarity of the stratigraphic geometries to those produced by suspension processes may be attributed to suspension within the air column. Siliciclastic sediment is almost entirely silty to very fine-grained, well-sorted quartz sand. A broad shelf is interpreted from regional sections and serves as a likely area for storage of siliciclastic sediment during relative rises. Relative falls in sea level facilitated eolian transport to the shelf-margin. At times these sands may have spilled
directly into the basin and were carried to locations below storm wavebase. In most cases, marine reworking during subsequent transgressions obscures the eolian nature of precursor Guadalupian shelf sandstone's.

Isopachs of the Lower Cherry Canyon Tongue sandstone (Fig. 24) show the siliciclastic sediment is thicker in the Last Chance Canyon region than in the surrounding region (Sonnenfeld, 1991). Perhaps a local fluvial or submarine channel concentrated the sediment.

**Erosion**

The interaction of relative changes of sea-level and the depositional system controls the timing and distribution of erosional surfaces. PHIL reconstructs the deposition and erosion of missing section layers in a manner consistent with the depositional systems and base-level history. In the Last Chance Canyon model, erosion is characterized by surface beveling, or locality-specified erosion. A surface beveling equation with a constant of 2 mm/ky eroded exposed surfaces (Bowman and Vail, 1993). The channel erosion rate is 0 mm/ky because fluvial-deltaic deposits are not reported within the San Andres basin-margin strata. Shoreface erosion did not contribute to an improved model. In this model we specified erosion by location, erosion rate and timing to remove sediment at the shoreface depth and deeper due to submarine canyon formation.

Most erosion occurs on third-order sequence boundaries due to lowering of base-level. Evidence of less than 20 meters of erosion is present in the truncated stratal lines below the G2 sequence boundary. Sporadic breccias are found in the outcrop lying on sequence boundaries and are likely remnants of the erosional processes that were active during marine erosion. The stratigraphic cross-section of Sonnenfeld and Cross (1992) does not show breccias observed on the sequence boundary at the base of depositional sequence G2.
Figure 24. Isopach of the Lower Cherry Canyon sand in the Last Chance Canyon region (from Sonnenfeld, 1991).
Erosional topography on sequence boundaries is only approximate. Predicting the precise location and scarp shape requires knowing positions of channels and shape of coastlines in 3-dimensions. It is difficult to determine the magnitude of erosion on the top of USA4. The abrupt angles on the toplap of surfaces 7, 8 and 9 indicate there may be up to 10 meters missing along that contact. However, preservation of the offlap breaks of these surfaces indicate that overall erosion must be less than 20 meters. The maximum erosion rate applied is 125 mm/ky.

**Initial Bathymetric Profile**

The initial bathymetric profile was digitized from the G1 sequence boundary and vertically shifted to match onlap position and lithofacies at the onset of deposition above the sequence boundary. Flattening occurred after deposition due to compaction. The initial bathymetry is adjusted using the depth differences at the end of the model to account for this. The stratigraphic path of the Last Chance Canyon section begins at the left margin of the cross-section, at a height of 0 meters above the initial bathymetric profile. A 2000 meter thick compacting substrate with an initial porosity of 25% and a compaction rate of 0.0003 m\(^{-1}\) simulated the underlying sedimentary column.

**Tectonic Subsidence History**

The tectonic subsidence rate was determined by holding long-term eustasy constant and backstripping a composite Leonardian through Guadalupian section (Fig. 25). Subsidence analysis suggested that tectonic subsidence rates were constant at a rate less than 10 mm/ky. This rate is very low indicating the basin was stable. More than 50% of the subsidence during the deposition of this interval is due to flexure loading by the sediment load or compaction of a 2000 meter substrate (Fig. 5). However, it is not possible to define the true tectonic component until similar studies are completed in other regions and eustasy for this interval is understood.

The model accounts for subsidence due to flexural loading of the lithosphere by sediment and water. In the Last Chance Canyon region, the effective elastic crustal
Figure 25. Subsidence History calculated from backstripping a section through the Lowe #1 Indian Hills Well, Guadalupe Mountains, New Mexico and a composite from other wells for the section above the San Andres Formation.

thickness of 11 km, which is thick enough to distribute a load equally throughout the cross-section. This equates to a flexural wavelength of 35 km. It is impossible to precisely constrain the true value from this model because the cross-section is too short (total model length is 18 km) for one to observe flexural bending. The flexural algorithm projects the loads deposited at the model margins beyond the boundary of the model and calculates the flexural response within the modeled cross-section. The magnitudes of the loads decrease to zero at a distance of 15 km.

Burial compaction moderately affects stratal geometries and the quantitative history of the section. The model results of the Last Chance section were buried to 500 meters and compacted according to 500 meters of burial. The compacted section was
compared with the uncompacted results (Fig. 26). Compaction is modeled as a function of burial depth using constants in Table 2 with the equation 1 (Dickinson et al., 1987):

\[ \phi(z) = \frac{\phi_0}{1 + z r_c} \]  

(1)

The compacting substrate was modeled as a 2000 meter thick layer with an initial porosity of 25% and compaction rate of 0.0003 to reproduce the compaction related subsidence recorded in the section.

Table 2. Compaction parameters used in the compaction functions.

<table>
<thead>
<tr>
<th>Lithology Name</th>
<th>Grain Density (kg/m³)</th>
<th>Initial Porosity (Φ₀)</th>
<th>Rate Constant (Rc) (m⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Coarse Sand to Cobbles</td>
<td>2650</td>
<td>0.4</td>
<td>0.0001</td>
</tr>
<tr>
<td>Quartz Silt Size</td>
<td>2650</td>
<td>0.3</td>
<td>0.001</td>
</tr>
<tr>
<td>Quartz Silt/Clay</td>
<td>2650</td>
<td>0.4</td>
<td>0.002</td>
</tr>
<tr>
<td>Quartz Sand/Clay</td>
<td>2650</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Interbedded Quartz/Silt</td>
<td>2650</td>
<td>0.45</td>
<td>0.0005</td>
</tr>
<tr>
<td>Interbedded Silt/Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.002</td>
</tr>
<tr>
<td>Substrata lithology</td>
<td>2750</td>
<td>0.25</td>
<td>0.0003</td>
</tr>
<tr>
<td>Kaolinite Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Mixed-Layer Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Glaucnite Clay</td>
<td>2750</td>
<td>0.5</td>
<td>0.003</td>
</tr>
<tr>
<td>Silt/Coal</td>
<td>2450</td>
<td>0.6</td>
<td>0.008</td>
</tr>
<tr>
<td>Clay/Coal</td>
<td>2300</td>
<td>0.85</td>
<td>0.009</td>
</tr>
<tr>
<td>Coal</td>
<td>2000</td>
<td>0.92</td>
<td>0.01</td>
</tr>
<tr>
<td>Peloidal Oolitic grainstone</td>
<td>2800</td>
<td>0.45</td>
<td>0.0001</td>
</tr>
<tr>
<td>Carbonate Fine grainstone</td>
<td>2800</td>
<td>0.45</td>
<td>0.001</td>
</tr>
<tr>
<td>Carbonate Coarse grainstone</td>
<td>2800</td>
<td>0.5</td>
<td>0.0005</td>
</tr>
<tr>
<td>Carbonate Boundstone</td>
<td>2800</td>
<td>0.5</td>
<td>0.002</td>
</tr>
<tr>
<td>Micrite</td>
<td>2800</td>
<td>0.65</td>
<td>0.004</td>
</tr>
<tr>
<td>Algal Laminates</td>
<td>2800</td>
<td>0.4</td>
<td>0.0005</td>
</tr>
<tr>
<td>Dolomite</td>
<td>2900</td>
<td>0.4</td>
<td>0.0001</td>
</tr>
<tr>
<td>Gypsum</td>
<td>2330</td>
<td>0.1</td>
<td>0.00001</td>
</tr>
<tr>
<td>Calciturbidite Sand</td>
<td>2800</td>
<td>0.6</td>
<td>0.0005</td>
</tr>
<tr>
<td>Calciturbidite Silt to Sand</td>
<td>2800</td>
<td>0.6</td>
<td>0.0005</td>
</tr>
<tr>
<td>Calciturbidite Fine Silt</td>
<td>2800</td>
<td>0.6</td>
<td>0.0005</td>
</tr>
</tbody>
</table>

Comparing post-Guadalupian sections throughout the Delaware Basin suggest that before Laramide and Basin and Range uplifts (McKnight, 1986), this section may have been buried to at least 500 meters. The last step of the model buries and compacts the section by 500 meters of burial: When typical compaction functions are used, the simulated section compacts to 85% of the final modeled thicknesses (Fig. 26). This
required increasing the slope angles, magnitudes of the sediment supply and carbonate production rates, and subsidence rates by a factor of 1.18 times of those measured from the cross-section. Early differential compaction during burial locally alters the stratal geometries below the depocenter, increasing the thickness of the interval. Compaction requires increasing the depositional gradients by approximately 10% to account for subsequent flattening. Compaction of uncemented sediment in the basin would decrease the slopes at the basin margin. In this case, the slopes would continue to rotate during progradation producing a fan geometry, expanding toward the basin, which is not observed.

The initial bathymetry, relative changes of sea-level, sediment supply and post-depositional compaction control the thickness of an interval. There are no indications of synsedimentary rotation of strata during deposition of the three sequences. Matching thicknesses at the cross-section margins is, therefore, sufficient to constrain total subsidence. The thickness from the Hayes sandstone to the G1 sequence boundary in both the model and the cross-section is 60 meters on the west side and approximately 210 meters on the east side.

Eustasy

An approximately 1 million-year variation (third-order) is the dominant control on the stratigraphy of Last Chance Canyon. However, the rock record is rarely organized enough for intervals to fall into well-behaved sinusoidal patterns. Carbonate growth and erosion rates regulate the timing of events in the stratal path. Carbonate growth rates were assumed to be constant. By iteratively adjusting the sea-level history to match the stratal path, a sea-level history is defined within 0.5 meter of the amplitude (Fig. 5). Vertical changes in the stratal path were attributed to eustasy and horizontal changes to sediment supply. When more carbonate sediment was required over a short interval, more time was introduced to that segment by shifting the younger portion of
Figure 26. Post-depositional compacted section through the Last Chance Canyon.
the curve forward in time. When more siliciclastic sediment was required, the siliciclastic sediment supply for that time was increased or the interval of the sea level fall was stretched.

The program makes the adjustments to sea level and tectonic subsidence at the beginning of deposition of a layer. This frequently produces a flooding event. The stratigraphic response is similar to a lag time for one-dimensional models such as those observed by Read et al., (1991), and Goldhammer and Dunn (1991). In two dimensions, the response is similar to the parasequence geometry of Van Wagoner (1990) with components bounded by flooding events at the base.

DISCUSSION

Future comparison with other Permian sections will help separate eustasy from local tectonic subsidence. The scale and rates of processes necessary to produce this section provide a well-defined basis to compare with other settings.

Assuming the eustatic and tectonic contributions of relative changes of sea-level have been partitioned correctly, the mechanisms causing base-level change can be inferred based on the shape of the eustatic cycles. A transition from a previous period of high magnitude falls during the Leonardian to low-magnitude cycles in the Guadalupian suggests a climate change may have taken place at the end of the Leonardian. The dominant cyclicity eustasy in the interval examined (Fig. 5) contains three major sea level cycles with an unmeasured fall for the first cycle (greater than 45 meters) and an approximately 32 and 38 meter sea level fall for the second and third cycle.

High-frequency eustatic cycles were modeled directly by adjusting eustasy to match the stratal patterns. Documentation of Milankovitch-scale cyclicity is strongly dependent upon the choice of a time-scale, precise dating capabilities beyond present capabilities, and the assumption that all the components or component groups are counted and assigned their appropriate duration. With this caveat, over the 1.74 My
duration of the simulation more than 24 cycles sub-dominant cycles occur forming component groups. These cycles average 74 ky per cycle. This periodicity may correspond to a 75 ky eccentricity cycle (Berger, 1984). As an alternative, we could build the model with a 2.26 My duration and the third-order cycles would correlate with the 1.3 My eccentricity mode. The high-frequency cycles would have a duration of 94 ky corresponding to the 98 ky eccentricity cycle.

**Relative Changes Of Sea Level and the Timing of Systems Tract Boundaries**

Selection of third-order sequence boundaries G1, G2, and G3 are straightforward using the criteria of the first major downward shift of coastal onlap. Stratigraphic terminations that onlap onto a regional surface indicated by erosion or an increase in grain size and siliciclastic sediment define sequence boundaries. This selection agrees with previous sequence stratigraphic correlation's in the Last Chance region (Sarg and Lehmann, 1986). Using a time scale from Ross and Ross (1987), the falls correspond to model ages of 258.7, 257.7, and 257.455 Ma. Although marine onlap formed on the cutting surfaces of many component groups by onlapping turbidite sands, the shallow marine carbonates rarely exhibit onlap of the stratal surfaces (perhaps above surface 16, Fig. 1). Surface 16 may have formed in response to a high-frequency fall with an unusual high magnitude.

Most onlap in the section is marine rather than coastal onlap. With coastal onlap the depositional interface intersects the underlying surface at or above sea level. Coastal onlap is common in siliciclastic rocks as fluvial transport of traction-load sediment is linked to sea level. With marine onlap the depositional interface intersects the underlying surface below sea level. Marine onlap is common in carbonate settings and suspended siliciclastic sediments because sedimentation is not linked to the intersection of the non-marine sediment source and sea level. This is common in carbonate settings because sediment deposition is not dependent on filling to sea-level.
Superimposing a clinoform geometry with a tidal flat and back reef interfaces determines marine base level. The back reef interface is projected landward and upward from the offlap break, at a gradient of 0.023, until the surface reaches the base of the tidal range. The tidal flat is projected landward and upward at a gradient of 0.0001 until the surface reaches the upper tidal range. Marine erosion removes all the sediment that lies above this interface and transports it to regions below base level. Sediment production must fill the shallow landward regions in the tidal environment with sediment produced in deeper water such as the lagoon and platform margin. Shallow marine currents preferentially transport sediment to regions below base-level landward of the offlap break before the base level interface migrates seaward.

The program determined the position of sequence boundaries by recording the first surface after a relative rise where relative sea level fell faster than -30 mm/ky. Systems tract boundaries were determined by selecting surfaces within a sequence the contained the most flooded cells (maximum flooding surface) and least flooded cells (top lowstand surface).

The three surfaces (4, 5, and 6, Fig. 1) above G2 exhibit progradation and increasing rates of aggradation. This is characteristic of a lowstand prograding complex. Surface 4 formed during a relative fall because it onlaps out at the offlap break of surface 3. Obvious choices for the top of the lowstand prograding complex and maximum flooding surface do not exist in this section. Slow subsidence rates restricted the creation of accommodation space necessary to produce a well-developed transgression and enhanced progradation.

A simulation of the Last Chance Canyon section without the siliciclastic sediment supply (Fig. 21) produces a slightly smaller volume of sediment. The well-developed lowstand prograding complex in the pure carbonate model suggests that siliciclastic influx hampers progradation by reducing carbonate production. Carbonate production is decreased because the average water depth increased with siliciclastic
deposition of gravity-flow deposits deeper in the basin. The load of the siliciclastic sediment did not affect total subsidence.

It is not completely clear whether surface 5 or 6 is the best pick for the maximum flooding surface. These surfaces form downlap surfaces for overlying component groups. A regionally extensive glauconitic, micritic siltstone bed overlies surface 5. The carbonate facies above surface 5 marks the most landward extent of the glauconite-rich facies. Surfaces 5 and 6 extend landward beyond the margin of the cross-section.

PHIL selected a maximum flooding surface by choosing a component within the sequence with the maximum extent covered by water and the deepest average water depth. The program analyzes the cross-section for the surface area below sea level at the beginning and end of every time-step. The maximum flooding surface is the time-slice between two sequence boundaries with the largest area below sea-level. On the basis of this criterion, the maximum flooding surface is surface 4. In the original section, a younger surface was selected based on the time of minimum siliciclastic input to the proximal slope, inferred to correspond to the maximum landward retreat of the fluvial and eolian systems on the shelf. Perhaps the small phase lag between carbonate and siliciclastic stacking patterns is a function of the complex relationship between siliciclastic poisoning of carbonates and depth-dependent carbonate productivity functions. Additionally, strike variations due to subtle changes in progradation directions between components may affect the maximum landward extent of fusulinid facies on the original cross-section.

PHIL selected a different "third-order" sequence boundary based on the beginning of a relative fall that is significantly younger than a previous choice based on an unconformity. Surface 7 is a sequence boundary because it marks the first major downward shift. The new interpretation suggests that most of the siliciclastic-rich prograding components form an upper lowstand systems tract or forced regression. The
component groups 8 through 17 show downstepping aggradation accompanied by increasing progradation and gravity flow sedimentation. Many components in this interval have onlap terminations of carbonate intervals and siliciclastic sands that record relative falls greater than 10 meters. During deposition of these components, third-order base-level falls enhanced the fall of “fourth-order” eustatic cycles and minimized the transgressive phase of the cycles. Coupled with high carbonate productivity on the platform margin, the relative fall enhanced the siliciclastic sediment supply relative to earlier fourth-order sequences (e.g., USA3), thereby enhancing progradation and limiting the systems ability to backstep during transgressive phases.

Long-term relative changes of sea-level probably have little effect on the timing of systems tract boundaries in this case because the long-term changes are much slower than the short-term cycles. The systems tract boundaries form within high-frequency cycles with rates that overwhelm the third-order or slower changes.

CONCLUSIONS

The results of the simulation provide a quantitatively defined history of relative changes of sea-level produced with long-term, “third-order” and higher-order variations. We generated this history of relative changes of sea-level by matching the base-level trends recorded in the stratal patterns. Vertical changes in base-level are attributed to eustasy, long-term changes to tectonism, and horizontal changes to sediment supply. If more carbonate sediment was required over a short interval, more time was introduced to that segment by shifting one side of the curve forward in time.

The results have bearing on designating which surfaces are sequence boundaries and maximum flooding surfaces. We have selected the “third-order” maximum flooding surface as the component within the “third-order” sequence that flooded the platform the most. In the original section, the maximum flooding surface was selected based on the time of minimum siliciclastic influx to the slope, inferred to correspond to the maximum landward retreat of the fluvial and eolian systems on the shelf.
Additionally, subtle changes along strike in progradation direction between component groups may affect the maximum landward extent of fusulinid facies on the original cross-section.

On the basis of downstepping accommodation of the interval from surface 7 to 15, and the influx of sands at surface 7 suggested we move the sequence boundary from surface 15 to surface 7. This interval is best characterized as the upper lowstand prograding complex or a forced regression.

Last Chance Canyon is a classic section for displaying prograding clinoforms. The model results show this interval recorded an episode of modest third-order relative sea level cycles (30-60 meter amplitudes), a variable and modest siliciclastic-influx rate, and the initial depositional-profile is very shallow.

The “third-order” base-level fall enhanced the fall of the “fourth-order” cycles and minimized the transgressive phase of cycles. Coupled with high carbonate productivity on the platform margin, the relative fall enhanced the siliciclastic sediment supply relative to earlier stratal components (e.g., uSA3), thereby enhancing progradation and limiting the depositional systems ability to backstep during transgressive phases.

Tectonic uplift or a long-term sea-level fall may produce predictable changes in the ratios of systems tract volumes within each sequence. Transgressive systems tracts in Last Chance Canyon are poorly developed compared with earlier sequences in the Permian and other settings. High rates of siliciclastic influx may contribute to the poor development of the lowstand prograding complex. High siliciclastic-influx during the relative fall may inhibit carbonate sediment production. Falling long-term base-level may also increase the duration of gravity-flow sedimentation and minimize the capacity of the depositional system to build a lowstand prograding-complex. Slumping and deposition as coarse-grained gravity-flow deposits remove more sediment than in a pure carbonate setting.
GLOSSARY OF TERMS

Accommodation space: Space created or removed, landward of the stormweather wavebase. Sediment may fill the space below base-level. The sediment supply and erosion control the amount of retrogradation or progradation that may occur. Seaward of the stormweather wavebase, stresses at the air or water surface do not influence depositional processes and the resulting stratal surfaces.

Aggradation: Vertical rise of base-level recorded by accumulation of sediment.

Changes in accommodation space: The sum of tectonic subsidence, eustasy, and erosion. The rate of change of accommodation space controls the amount of aggradation or incision that may occur.

Backstepping: Landward retreat of each stratal component.

Base-level: A theoretical surface that sedimentation processes attempt to reach. Base-level is a locally defined surface influenced by eustasy, tectonic subsidence, and sediment supply. When the depositional interface is below base-level, deposition may occur. When the depositional interface is above base-level, erosion may occur. Base-level migrates basinward when all the space below base-level is filled. Base-level is also a function of the sediment grain-size or type. Base-level for a large grain may be shallower than base-level for a smaller grain.

Bathymetry: The depth of the basin below sea level. The measurement is positive below sea level.

Component: See stratal component.

Component group cycle: A eustatic cycle with a smaller amplitude than a dominant eustatic cycle. The amplitude is insufficient to develop a full series of systems tracts.

Downstepping: Seaward progradation and falling of each stratal component.

Eustasy: Mean sea level globally defined relative to the geoid. The volume of water stored as ice, average temperature of the ocean, the size of ocean basins as well as other factors control eustasy.
Eustatic change of sea level: a change in sea level that is equal in all locations.

Fairweather wavebase: Water depth to which fairweather wave action moves fine to coarse grained sediment (range 8-20 meters, average 10 meters of water depth).

Forced regression: the process of prograding during a relative fall of sea level.

Forestepping or Progradation: Basinward migration of the shoreline and associated depositional profile between stratal components.

Gravity flow transport: Transport of coarse-grained sediment through the shallow marine setting to the deep marine setting. Coastal or deltaic transport processes generally produce a decrease in grain size away from the source. Gravity flow processes provide a mechanism that can transport coarse-grained sediment beyond fine-grained sediments.

Incomplete sequence: A stratigraphic package that is locally missing one or more system tracts, and that can probably be correlated with a time-equivalent complete sequence. Deposition may be above either a Type I or Type II sequence boundary.

Influx surface: A surface along which a notable decrease in the sediment influx rate or energy level (represented by the grainsize of sediment) occurs. The opposite is a deprivation surface.

My: millions of years.

Ma: Millions of years ago.

m: meters.

Offlap: Stratal surfaces dipping at a greater angle than those above or below. Offlap indicates progradation into deeper water.

Offlapping: Prograding component groups or stratal components.

Offlap break: A break in slope on a depositional clinoform that occurs most often near the fairweather wavebase.

Paleobathymetry: Depth of sediment-water interface at the time of deposition, as recorded in the paleobathymetric indicators within the sediments.
Parasequence: A relatively conformable succession of genetically related beds or bedsets bounded by marine-flooding (deprivation) surfaces or their correlative surfaces. In special positions within the sequence, parasequences may be bounded either above or below by sequence boundaries (Van Wagoner et al., 1990).

Perched lowstand: a prograding series of clinoforms deposited during a relative fall of sea level.

Relative change of sea level: The sum of tectonic subsidence (thermal subsidence, thrust loading), flexure loading of sediment, and eustatic changes. It controls shelfal accommodation and the timing of gravity flow transport. If the sediment supply is constant, relative change of sea level will determine the amount of progradation or retrogradation that may occur.

Retrogradation: Local landward progradation of stratal surfaces from a barrier island. Usually accompanied by erosion of the barrier island shoreface.

Shelf-slope break (continental shelf edge): Bathymetric feature that divides the shallow gradients of the shoreface from the steep gradients of the slope.

Sequence boundary: Base of a depositional sequence marked by a relative fall of sea level indicated by onlap, incision, erosional truncation, or an increase in grain size.

Sedimentation: All erosion and deposition processes.

Sedimentation interface: Topographic surface (sediment and air interface) above sea level and bathymetric surface (sediment and water interface) below sea level.

Measurement is positive below sea level, negative above sea level.

Deprivation surface: A surface along which a notable decrease in the sediment influx rate or energy level (represented by the grain size of sediment) occurs. The opposite is an influx surface.

Stratal component: A relatively conformable succession of genetically related beds or bedsets bounded by a correlative surface that contains a single depositional episode or a series of similar episodes without a significant change. These include
forestepping, downstepping, backstepping, and upstepping packages, slump, individual lobe of a basin floor fan or leveed-channel, or packages of shingled turbidites, channel fill, point bar migration. They are often bounded by a Deprivation surface (marine-flooding surface), influx surface (erosion surface), glide surface or their correlative surfaces.

Stratal path: The location of a stratal feature or facies boundary through time. Examples include the position of the offlap break or coastline.

Tectonic subsidence: Driving subsidence or uplift. Typically caused by thermal cooling of the crust, flexure loading by crustal shortening, diapirs driven by density contrasts and faults (McKenzie, 1978).

Topography: The height of the earth surface above sea level. Measurement is positive above sea level. To avoid confusion between topographic and bathymetric reference frames we fixed our reference surface to bathymetry to define the sediment interface.

Type I sequence boundary: Base of a depositional sequence marked by coastal onlap below the offlap break of the previous highstand. Gravity-flow transport with deposition as turbidites occurs above a Type I sequence boundary.

Type II sequence boundary: Base of a depositional sequence marked by a coastal onlap above the offlap break of the previous highstand. Evidence of gravity flow of coarse-grained sediment is minimal to absent.

Upstepping: Vertical aggradation during deposition of a stratal component.
References


Barrell, J., 1917, Rhythms and the measurements of geologic time: GSA Bulletin, v. 28, p. 50-60


Bice, D., 1988, Synthetic stratigraphy of carbonate platform and basin systems: Geology, v. 16, p. 703-706


Collier, R.E. Ll., M.R. Leeder, and J.R. Maynard, 1990, Transgressions and regressions: a model for the influence of tectonic subsidence, deposition and eustasy, with application to Quaternary and Carboniferous examples: Geol. Mag., v. 127, no. 2, p. 117-128


Damuth, J.E., R.D. Flood, R.O. Kowsmann, R.H. Belderson, and M.A. Gorini, 1988, Anatomy and Growth Pattern of Amazon Deep-Sea Fan as Revealed by Long-Range Side-Scan Sonar (GLORIA) and High-Resolution Seismic Studies: AAPG Bulletin, v. 72, no. 8, p. 885-911


Greenlee, S.M., F.W. Schroeder, and P.R. Vail, 1988, Seismic stratigraphy and
geohistory analysis of Tertiary strata from the continental shelf off New Jersey;
Calculation of eustatic fluctuations from stratigraphic data: The Geology of North
of America, p. 437-444
Haq, B. U., J. Hardenbol, and P. R. Vail, 1987, Chronology of Fluctuating Sea Levels
since the Triassic: Science, V. 235, p. 1156-1167
Haq, B., J. Hardenbol, and P.R. Vail, 1988, Mesozoic and Cenozoic Chronostratigraphy
approach, SEPM Special Publication No. 42, p. 71-108.
Harbaugh, J.W., 1966, Mathematical simulation of marine sedimentation with IBM
7090/7094 computers: computer contribution 1: Wichita, Kansas, Kansas Geological
Survey, 52 p.
Harland, W.B., R.L. Armstrong, A.V. Cox, L.E. Craig, A.G. Smith, D.G. Smith, 1989,
Hubbard, D.K., and D. Scaturo, 1985, Growth rates of seven species of scleractinean
325-338
Huston, M., 1985, Variation in Coral growth rates with depth at Discovery Bay, Jamaica:
Coral Reefs, v. 4, p. 19-25
Jacquin, T.E., A. Arnaud-Vanneau, H. Arnaud, C. Ravenne, and P.R. Vail, 1991,
Systems tracts and depositional sequences in a carbonate setting: a study of continuous
outcrops from platform to basin at the scale of seismic lines: Marine and Petroleum
Geology, v. 8, no. 2, p. 122-139
Jervey, M.T., 1988, Quantitative geological modeling of siliciclastic rock sequences and
their seismic expression: in Wilgus, et al, (eds.), Sea-Level changes: an integrated
approach, Soc. Econ. Paleon. and Min. Special Publication No. 42, p. 47-69


Mount, J.F., 1984, Mixing of siliciclastic and carbonate sediments in shallow shelf environments: Geology, v. 12, p. 432-435
Mutti, E., 1989, Submarine sand mounds and their relations to turbidite systems reworked by bottom currents: Giornale di Geologia, Bologna; v. 51, Supplement No. 4


Powell, J.W., 1895, Physiographic features: National Geographic Monographs, v. 1, no. 2, p. 33-64
Quinlan, G.M., and C. Beaumont, 1984, Appalachian thrusting, lithospheric flexur, and the
Paleozoic stratigraphy of the eastern Interior of North America: Canadian Journal of
Earth Sciences, v. 21, p. 973-996


York, 569 p.

in Sequence Stratigraphy: The Influence of Flexural Isostasy and Compaction: Journ.
of Geophysical Research, v. 96, no. B4, p. 6931-6949

Rona, P. A., 1973, Relations between rates of sediment accumulation on continental
shelves, sea-floor spreading, and eustasy inferred from the central North Atlantic:

Ross, C.A., and J.R.P. Ross, 1987, Late Paleozoic sea-levels and depositional sequences:
in Ross, C.A., and D. Haman, (ed.), Timing and depositional history of eustatic
sequences, constraints on seismic stratigraphy: Cushman Foundation for
Foraminiferal Research Special Publication No. 24

Royden, L., and Keen, C.E., 1980, Rifting process and thermal evolution of the contin-
ental margin of eastern Canada determined from subsidence curves: Earth and

Sahagian, D.L., 1987, Epeirogeny and eustatic sea level changes as inferred from
Cretaceous shoreline deposits: Applications to the central and western United States;
Journal of Geophysical Research, v. 92, p. 4895-4904

Sangree, J. B., P. R. Vail, and R. M. Sneider, 1988, Evolution of Facies Interpretation of
the Shelf-Slope: Application of the New Eustatic Framework to the Gulf of Mexico,

Saucier, R.T., 1974, Quaternary geology of the lower Mississippi Valley: Research Series 6, Arkansas Archaeological Survey.


Sonnenfeld, M.D., 1991, Day One -- High-Frequency Cyclicity within Shelf-Margin and Slope Strata of the Upper San Andres Sequence, Last Chance Canyon: in Sequence Stratigraphy, Facies, and Reservoir Geometries of the San Andres, Grayburg, and
Queen Formations, Guadalupe Mountains, New Mexico and Texas, Permian Basin Section -- SEPM Annual Field Trip, Publication 91-32, p. 11-52.


University Ph D. Dissertation, 530 p.


Vail, P.R., 1987, Seismic stratigraphy interpretation using sequence stratigraphy, part 1:
of Seismic Stratigraphy.

Vail, P.R., F. Audemard, S.A. Bowman, P.N. Eisner, G. Perez-Cruz, 1991, The
Stratigraphic Signatures of Tectonics, Eustasy and Sedimentation; in Einsele, G. et
617-659.

Vail, P.R., Mitchum, R.M., Jr., and Thompson, S., III, 1977a, Seismic stratigraphy and
global changes of sea level: in C. Payton, editor, Stratigraphic interpretation of

Vail, P.R., Todd, R.G., and Sangree, J.B., 1977b, Seismic stratigraphy and global
changes of sea level, Part 5, Chronostratigraphic significance of seismic reflections:
in C.E. Payton, (ed.), Stratigraphic interpretation of seismic data; AAPG Memoir

Vail, P.R., and Wornardt, W., Jr., 1990, An integrated approach to exploration and
development in the 90's: Well log-seismic sequence stratigraphy analysis:
Transactions of the Gulf Coast Association of Geological Societies, v. XLI, p. 630-
650.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M. and Rahmanian, V.D., 1990,
Siliciclastic sequence stratigraphy in well logs, core and outcrop: Concepts for high-


PLEASE NOTE:

Oversize maps and charts are filmed in sections in the following manner:

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

The following map or chart has been refilmed in its entirety at the end of this dissertation (not available on microfiche). A xerographic reproduction has been provided for paper copies and is inserted into the inside of the back cover.

Black and white photographic prints (17" x 23") are available for an additional charge.
Figure 1a) Vertical incidence seismic model of the stratigraphic sim
Model (0 - 40 Hz Klauder wavelet)

velocities are for sea water fill pores, a function of mineral density, and a calculation for bulk modulus of the framework.

*Seismic Line through the Baltimore Canyon, offshore New J*
the framework that is a function of the porosity.

New Jersey
Figure 1b) Uninterpreted seismic line 6 of the Neogene section of the Baltimore Ca
Figure 1c) Comparison of the depth converted sequence boundaries for the Neogene section of the Baltic Sea. The difference between the model and the data is colored according to the age of the observed layer.
The section of the Baltimore Canyon Trough, offshore New Jersey with the model results. The observation is to the age of the observation. The average absolute difference for 3915 points of comparison is 33 meters.

Level Analysis

Relative Sea Level
Figure 1d) Stratal age model for the cross-

SL (m)
104 -135 0 km

6.2
model for the cross-section.

Chronostratigraphy
90
Figure 1e) Calculated relative changes in sea-level, tectonic and total subsidence history at the offlap break boundaries (yellow solid), top slope fan (brown dash), top of lowstand (pink dash), maximum flooding surface is represented by the band of colors across the top of the figure.
Eustasy
Tectonic Subside

Total Subsidence

0 Ma
16 11 6

Model Porosity
9 80 100 120 140

...
Figure 1f) Predicted chronostratigraphy. Solid line on the left depth in which the sediments were deposited.
Solid line on the left margin in the sea level history (fall is to the right). Circles represent the sediment supply history.
Predicted Lithofacies

<table>
<thead>
<tr>
<th>80</th>
<th>100</th>
<th>120</th>
<th>140</th>
</tr>
</thead>
</table>

(Colors represent the sediment supply history (increase to the right).)
Figure 1g) Predicted porosity distribution based on the lithofacies distribution and associated depth (meters) and predicted depth (meters) for different km ranges.
sediment deposition and associated compaction functions.

Revised Depositional Systems

80  100  120  140
Figure 1h) Predicted lithofacies. A siliciclastic ar
A siliciclastic and carbonate layer is plotted for each stratal component. The lithofacies on the upper surface is...
component. The lithofacies on the upper surface is projected downward.

Equivalent depth 1 km

Figure 3
Figure 1i) Predicted depositional systems.

Model Paleo
Figure 1) Comparison of a lithostratigraphic section with a proximal sp.
Grainsize trends are indicated by the right margin of the profile. Surface stratigraphic age is represented by the color bands on the left margin of the left margin represent missing time (the wider the arrow, the more time is
Graphic section with a proximal spontaneous potential (SP) and resistivity (R) log, right margin of the profile. Surfaces are coded by surface type (see legend Fig. 1k). The color bands on the left margin of the section (see legend in Fig. 1d). The arrows on the wider the arrow, the more time is missing.
Figure 1j) Calculated paleobathymetry.

Figure 1k) Calculated sequence stratigraphy. Sequence boundaries (orange) correspond to the first surface with a relative fall less than -30 mm/ky. The top of the lowstand surface (pink) is the surface with the most cells below sea level within a sequence.
The first surface with a relative fall less than -30 mm/ky. The top of the slope fan (brown) is the last surface with the most cells above sea level within a sequence. The maximum flooding surface (green) is the surface...
Pliocene-Miocene

Upper Miocene

Middle and Lower Miocene

Plate 1.
Baltimore Canyon

30.449 Ma

PHIL Project
Vertical Exagge
Plate 1. Simulation Results of the Baltimore Canyon Section, offshore New Jersey

Scott A. Bowman
Peter R. Vail
Rice University
Houston, Texas 77005

PHIL Project Number: 7 693 16 126
Vertical Exaggeration = 40 x

July, 1
Plate 1. Simulation Results of the
Baltimore Canyon Section, offshore New Jersey

Scott A. Bowman
Peter R. Vail
Rice University
Houston, Texas 77005

MIL Project Number: 7 693 16 126
Vertical Exaggeration = 40 x

July, 14, 1993