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A SEQUENCE STRATIGRAPHIC AND ISOTOPIC STUDY OF UPPERMOST PENNSYLVANIAN-LOWER PERMIAN CARBONATE STRATA, OROGRANDE BASIN, NEW MEXICO

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A SEQUENCE STRATIGRAPHIC AND ISOTOPIC STUDY OF UPPERMOST PENNSYLVANIAN-LOWER PERMIAN CARBONATE STRATA, OROGRANDE BASIN, NEW MEXICO

by

Jesse T. Koch

A DISSERTATION

Presented to the Faculty of
The Graduate College at the University of Nebraska
In Partial Fulfillment of Requirements
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Major: Geosciences (Geology)

Under the Supervision of Professor Tracy D. Frank

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Determining how the tropics respond to glaciation is important for improving our
global understanding of icehouse worlds. The impetus behind this study is to identify the
far-field impacts of Gondwanan glaciation during the Early Permian. Impacts including
climate and sea-level change should be evident in sensitive carbonate systems, such as
the upper Paleozoic paleotropical strata in the Orogrande Basin. A sequence stratigraphic
and stable isotopic approach is used to examine the effects of late Paleozoic climate
change in the Orogrande Basin during the acme and subsequent demise of the late
Paleozoic ice age.

Sequence stratigraphic analysis suggests the occurrence of a major unconformity
at the Pennsylvanian-Permian boundary, which coincided with the onset of major
Gondwanan glaciation. Additionally, two transgressive events during the Sakmarian and
Artinskian broadly correlate with the end of major glacial epochs in eastern Australia.

Stable isotopic analysis suggests periods of subaerial exposure from the Gzhelian
through the early Sakmarian, which are associated with lower $\delta^{13}$C values. $\delta^{13}$C values
return to global averages after the mid-Sakmarian. Two smaller negative isotope
excursions ($1.5\%$ decrease in $\delta^{13}$C values) occur within upper Sakmarian-Kungurian
strata. These lower $\delta^{13}$C values are associated with a facies change to shallower
conditions that coincided with periods of renewed expansion of glacial ice across parts of
Gondwana. Isotopic interpretations are consistent with the relative sea level
reconstructions inferred from the sequence stratigraphic analysis.

Examination of other paleotropical records suggests that the overall stratigraphic
pattern inferred for the Orogrande Basin reflects a global signal. Paleotropical carbonate
platforms experienced significant subaerial exposure in the earliest Permian, which is
represented by a sequence boundary or a basinward shift in facies. This eustatic drop
coincided with the initiation of major glaciation across Gondwana. Results from this study suggest that carbonate strata from the Orogrande Basin responded to glacial epochs evident in the Gondwanan record. This suggests that these glacial events were of global significance.
DEDICATION

…to my wife, Ambr.
ACKNOWLEDGEMENTS

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This project could not have happened without the help of my advisor, Dr. Tracy Frank. Her knowledge about the late Paleozoic, carbonates, stable isotopes, etc. has been a crucial asset to the success of this project. Before I arrived at Nebraska, I was relatively new to the topics of carbonate sedimentology, but now I have learned to truly appreciate all things carbonate. As a result of Dr. Frank’s guidance and support, I am a much better scientist, and I feel well-prepared for my post-Nebraska life. Thank you for your continued support and encouragement.

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my professional career. I cannot say thank you enough for teaching me these invaluable skills.

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Lastly, but most importantly, I want to thank my wonderful wife, Ambr, for her unwavering support during both of my graduate degrees. You inspire me to do my best and are always a source of encouragement. Our furry children (Darwin, Caeli, and Pooka) have helped keep me sane during this process through their unconditional support.

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# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>TITLE / SECTION</th>
<th>PG.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Preface</td>
<td>.......................................................................................................................... i</td>
<td></td>
</tr>
<tr>
<td>Title Page</td>
<td>.......................................................................................................................... i</td>
<td></td>
</tr>
<tr>
<td>Abstract</td>
<td>........................................................................................................................ ii</td>
<td></td>
</tr>
<tr>
<td>Dedication</td>
<td>......................................................................................................................... v</td>
<td></td>
</tr>
<tr>
<td>Acknowledgements</td>
<td>.................................................................................................................. vi</td>
<td></td>
</tr>
<tr>
<td>Table of Contents</td>
<td>.................................................................................................................... viii</td>
<td></td>
</tr>
<tr>
<td>List of Figures</td>
<td>......................................................................................................................... xiv</td>
<td></td>
</tr>
<tr>
<td>List of Tables</td>
<td>....................................................................................................................... xviii</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Introduction ......................................................................................................................... 1</td>
<td></td>
</tr>
<tr>
<td>Overview</td>
<td>.............................................................................................................................. 1</td>
<td></td>
</tr>
<tr>
<td>Recent Research on the Late Paleozoic Ice Age</td>
<td>............................................................................. 1</td>
<td></td>
</tr>
<tr>
<td>Motivation for this Study</td>
<td>.................................................................................................................. 4</td>
<td></td>
</tr>
<tr>
<td>Dissertation Structure</td>
<td>.................................................................................................................... 5</td>
<td></td>
</tr>
<tr>
<td>Chapter 2</td>
<td>............................................................................................................................... 6</td>
<td></td>
</tr>
<tr>
<td>Chapter 3</td>
<td>............................................................................................................................... 7</td>
<td></td>
</tr>
<tr>
<td>Chapter 4</td>
<td>............................................................................................................................... 7</td>
<td></td>
</tr>
<tr>
<td>Appendices</td>
<td>............................................................................................................................. 8</td>
<td></td>
</tr>
<tr>
<td>References</td>
<td>.............................................................................................................................. 9</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Sequence stratigraphic and facies analysis of the Orogrande Basin (New Mexico, USA): Imprint of late Paleozoic glaciation ............................................................................. 11</td>
<td></td>
</tr>
<tr>
<td>Abstract</td>
<td>.............................................................................................................................. 12</td>
<td></td>
</tr>
<tr>
<td>Introduction</td>
<td>............................................................................................................................. 13</td>
<td></td>
</tr>
<tr>
<td>Geologic Background</td>
<td>...................................................................................................................... 15</td>
<td></td>
</tr>
<tr>
<td>Orogrande Basin</td>
<td>......................................................................................................................... 15</td>
<td></td>
</tr>
<tr>
<td>Stratigraphy</td>
<td>............................................................................................................................ 17</td>
<td></td>
</tr>
</tbody>
</table>
Upper Pennsylvanian Units ................................................................. 20
Hueco Group .................................................................................. 21
Abo Formation .............................................................................. 23
Methods .......................................................................................... 24
Lithofacies Descriptions ................................................................. 31

Subaerial to Nearshore Siliciclastic Facies .......................... 31
  Lithofacies 1a: Red Mudrock ....................................................... 31
  Lithofacies 1b: Fine to Medium-Grained Sandstone .......... 35
  Lithofacies 1c: Pebble Conglomerate .................................. 35
  Lithofacies 1d: Thin-Bedded Siltstone and Mudrock ...... 36

Nearshore Restricted Carbonate Facies (Upper Ramp) ....... 36
  Lithofacies 2: Interbedded Microbial Laminite and
   Laminated Carbonate Mudstone ............................................. 36
  Lithofacies 3a: Sandy Mudstone to Wackestone ............. 37
  Lithofacies 3b: Gastropod/Ostracod Wackestone .......... 37

Open Marine Carbonate Facies (Middle-Upper Ramp) ...... 39
  Lithofacies 4: Oncoidal Wackestone, Packstone,
   and Grainstone .................................................................... 39
  Lithofacies 5a: Fusulinid Packstone to Grainstone ....... 39
  Lithofacies 5b: Fossiliferous Wackestone, Packstone,
   and Grainstone .................................................................. 40
  Lithofacies 5c: Dasycladacean Algae Wackestone and
   Packstone ........................................................................... 40

Carbonate Boundstone Facies (Middle-Lower Ramp) ...... 41
  Lithofacies 6a: Phylloid Algae Floatstone/Boundstone ..... 41
  Lithofacies 6b: Tubiphytes-Rich Floatstone/Boundstone .. 41

Offshore Facies (Lower Ramp to Basin) ................................. 42
  Lithofacies 7: Crinoid-Brachiopod Wackestone,
   Packstone, and Grainstone ................................................ 43
  Lithofacies 8: Laminated Mudrock and Shale ................. 43
Depositional Environments...............................................................................................................44
Subaerial to Nearshore Siliciclastic Facies.................................................................44
Nearshore Restricted Carbonate Facies (Upper Ramp)........................................47
Open Marine Carbonate Facies (Middle-Upper Ramp)........................................48
Carbonate Boundstone Facies (Middle-Lower Ramp)........................................49
Offshore Facies (Lower Ramp to Basin)...............................................................50
Sequence Stratigraphy.......................................................................................................51
Gzhelian (Holder Formation; Bursum Formation; Bough Units).................................................................................................................................................52
Asselian to Middle Sakmarian (Hueco Canyon Formation)............................55
Middle Sakmarian to Middle Artinskian (Cerro Alto Formation)...............................57
Late Artinskian (Alacran Mountain Formation)..............................................58
Synthesis and Discussion..............................................................................................59
Conclusions.....................................................................................................................61
Acknowledgements.........................................................................................................63
References.......................................................................................................................63

3 Carbon isotope stratigraphy of uppermost Pennsylvanian-Lower Permian marine carbonates in south-central New Mexico: Implications for glaciation, sea level, and depositional patterns..........................................................74
Abstract.......................................................................................................................75
Introduction.....................................................................................................................76
Geologic setting..............................................................................................................77
The Orogrande Basin....................................................................................................77
Stratigraphy....................................................................................................................79
Uppermost Pennsylvanian Strata..............................................................................80
Lower Permian Strata.................................................................................................82
Samples and analytical methods................................................................................83
Facies and Depositional Trends....................................................................................85
Facies Association A: mudstone and wackestone
(restricted to intertidal environment) ................................................................. 92
Facies Association B: wackestone-grainstone
(shallow subtidal environment) .............................................................................. 93
Facies Association C: boundstone/floatstone
(phyllloid algae-Tubiphytes bioherms) ................................................................. 94

Results ......................................................................................................................... 95
  Overview ....................................................................................................................... 95
  Stratigraphic Trends ..................................................................................................... 95

Discussion ................................................................................................................... 100
  Preservation of Primary Isotopic Signals ................................................................. 100
  Current understanding of the late Paleozoic ice age ..................................................... 106
  Brief Overview of Global Early Permian Proxy Trends .............................................. 108
  A Climate Signal Preserved in the Orogrande Basin? .................................................. 110

Conclusions .................................................................................................................. 113
Acknowledgements ........................................................................................................ 115
References ................................................................................................................... 116

4 The Pennsylvanian-Permian transition in the low-latitude carbonate record and the onset of major Gondwanan glaciation ......................................................... 130
  Abstract ....................................................................................................................... 131
  Introduction .................................................................................................................. 132
  Background .................................................................................................................. 134
  Methodology and Terminology .................................................................................. 136
    Biostratigraphic Correlations .................................................................................... 142
    Global Correlations .................................................................................................. 143
    Data Limitations ........................................................................................................ 143
  Regional Stratigraphic Records .................................................................................... 144
    United States Midcontinent ....................................................................................... 144
    Orogrande Basin, New Mexico, USA ........................................................................ 146
    Permian Basin, Texas, USA ....................................................................................... 147
## LIST OF FIGURES

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>FIGURE</th>
<th>PG.</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Introduction</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>Figure 1a-b: Study area maps of the Orogrande Basin</td>
<td>2</td>
</tr>
<tr>
<td>2</td>
<td>Sequence stratigraphic and facies analysis of the Orogrande Basin (New Mexico, USA): Imprint of late Paleozoic glaciation</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>Figure 1: Study area map of the Orogrande Basin</td>
<td>16</td>
</tr>
<tr>
<td></td>
<td>Figure 2: Chronostratigraphic framework for the Orogrande Basin</td>
<td>18</td>
</tr>
<tr>
<td></td>
<td>Figure 3: Graphic column of the Holder/Abo Formations (northern Sacramento Mountains)</td>
<td>25</td>
</tr>
<tr>
<td></td>
<td>Figure 4: Graphic columns of the upper Pennsylvanian cores from the Northwest Shelf (Permian Basin)</td>
<td>26</td>
</tr>
<tr>
<td></td>
<td>Figure 5: Graphic column of the Hueco/Abo Formations (northern San Andres Mountains)</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>Figure 6: Graphic column of the Hueco Canyon Formation (Franklin Mountains)</td>
<td>28</td>
</tr>
<tr>
<td></td>
<td>Figure 7: Graphic column of the Cerro Alto Formation (Franklin Mountains)</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td>Figure 8: Graphic column of the Alacran Mountain Formation (Franklin Mountains)</td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>Figure 9: Photos of Lithofacies 1a-3a</td>
<td>34</td>
</tr>
<tr>
<td></td>
<td>Figure 10: Photos of Lithofacies 3b-6a</td>
<td>38</td>
</tr>
<tr>
<td></td>
<td>Figure 11: Photos of Lithofacies 6b-8</td>
<td>42</td>
</tr>
<tr>
<td></td>
<td>Figure 12: Depositional model for the Orogrande Basin</td>
<td>45</td>
</tr>
<tr>
<td></td>
<td>Figure 13: Lithostratigraphic framework of outcrops/cores</td>
<td>53</td>
</tr>
<tr>
<td></td>
<td>Figure 14: Sequence stratigraphic interpretation for the carbonates of the Orogrande Basin</td>
<td>54</td>
</tr>
</tbody>
</table>
3 Carbon isotope stratigraphy of uppermost Pennsylvanian-Lower Permian marine carbonates in south-central New Mexico: Implications for glaciation, sea level, and depositional patterns.................................................................74
  Figure 1: Study area map of the Orogrande Basin..................................................78
  Figure 2: Chronostratigraphic framework for the Orogrande Basin.................................81
  Figure 3: Plot of δ^{13}C values by stratigraphic section..............................................84
  Figure 4: Plots (by time interval and lithostratigraphic unit) of δ^{13}C versus δ^{18}O values..............................................................................................................96
  Figure 5: Composite δ^{13}C values for the study area.................................................98
  Figure 6: Cathodoluminescence images..........................................................................101
  Figure 7: Petrographic and outcrop images showing diagenetic alteration..........................102

4 The Pennsylvanian-Permian transition in the low-latitude carbonate record and the onset of major Gondwanan glaciation........................................................................130
  Figure 1: Map of study localities..................................................................................133
  Figure 2: Global stratigraphic framework......................................................................139
  Figure 3: Biostratigraphic framework for uppermost Pennsylvanian-Lower Permian strata.................................................................140
  Appendices..........................................................................................................................200
  A. Graphic columns of outcrops and cores...............................................................200
     Figure 1: List of symbols used for graphic logs....................................................201
     Figure 2: Graphic log from the lower Sacramento Mountains (0-100 m)..................202
     Figure 3: Graphic log from the lower Sacramento Mountains (100-200 m)............203
     Figure 4: Graphic log from the lower Sacramento Mountains (200-290 m)............204
Figure 5: Graphic log from the Getty #5 Willard Beatty Core (0-67 m)...........................................................................................................................205
Figure 6: Graphic log from the Nicor DDH-2 Core (0-39 m)...........................................................................................................................206
Figure 7: Graphic log from the Tres Papalotes Core (0-16.5 m)......................................................................................................................207
Figure 8: Graphic log from the Unocal State 1-33 Core (0-33.5 m)......................................................................................................................208
Figure 9: Graphic log from the lower San Andres Mountains (0-90 m)...............................................................................................209
Figure 10: Graphic log from the Franklin Mountains (0-100 m)............................................................................................................210
Figure 11: Graphic log from the Franklin Mountains (100-200 m)...............................................................................................211
Figure 12: Graphic log from the Franklin Mountains (200-300 m)...............................................................................................212
Figure 13: Graphic log from the Franklin Mountains (300-400 m)...............................................................................................213
Figure 14: Graphic log from the Franklin Mountains (400-500 m)...............................................................................................214
Figure 15: Graphic log from the Franklin Mountains (500-600 m)...............................................................................................215
Figure 16: Graphic log from the Franklin Mountains (600-700 m)...............................................................................................216
Figure 17: Graphic log from the Franklin Mountains (700-783 m)...............................................................................................217
Figure 18: Graphic log from the upper San Andres Mountains (0-101 m)...............................................................................................218
Figure 19: Graphic log from the upper Sacramento Mountains (0-100 m)...............................................................................................219
Figure 20: Graphic log from the upper Sacramento Mountains (100-186 m) ................................................................. 220

B. Additional lithofacies photographs ......................................................... 221

Figure 1: High-frequency sequence from the Sacramento Mountains (Holder Formation) ................................................. 222

Figure 2: Small phylloid algae bioherm in the Holder Formation (Sacramento Mountains) .................................................. 223

Figure 3: Shallowing upward cycles in the Hueco Canyon Formation (Franklin Mountains) ............................................... 224

Figure 4: Large phylloid algae bioherm at the top of the Alacran Mountain Formation (Franklin Mountains) .................... 225

Figure 5: Large phylloid algae bioherms in the lower Alacran Mountain Formation (Franklin Mountains) .......................... 226

Figure 6: Close-up view of phylloid algae in the Franklin Mountains) .................................................................................. 227

C. Miscellaneous figures not included in publications ................................ 228

Figure 1: $\delta^{13}$C values plotted against time interval ............................... 229

Figure 2: $\delta^{18}$O values plotted against time interval ................................. 230

Figure 3: GIS map of study area (#1) ....................................................... 231

Figure 4: GIS map of study area (#2) ....................................................... 232
# LIST OF TABLES

<table>
<thead>
<tr>
<th>CHAPTER</th>
<th>TABLE</th>
<th>PG.</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>Sequence stratigraphic and facies analysis of the Orogrande Basin (New Mexico, USA): Imprint of late Paleozoic glaciation</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>Table 1: Facies associations for upper Paleozoic strata of the Orogrande Basin</td>
<td>32</td>
</tr>
<tr>
<td>3</td>
<td>Carbon isotope stratigraphy of uppermost Pennsylvanian-Lower Permian marine carbonates in south-central New Mexico: Implications for glaciation, sea level, and depositional patterns</td>
<td>74</td>
</tr>
<tr>
<td></td>
<td>Table 1: Raw stable isotopic data from measured sections and cores used in this study</td>
<td>86</td>
</tr>
<tr>
<td>4</td>
<td>The Pennsylvanian-Permian transition in the low-latitude carbonate record and the onset of major Gondwanan glaciation</td>
<td>130</td>
</tr>
<tr>
<td></td>
<td>Table 1: Summary of stratigraphic data</td>
<td>137</td>
</tr>
</tbody>
</table>
CHAPTER 1: INTRODUCTION
OVERVIEW

This dissertation examines stratigraphic and stable isotopic records from the Orogrande Basin, which was situated in the paleotropics during the late Paleozoic and was the site of carbonate deposition (Fig. 1a-b). This basin is an ideal place to examine the imprint of Gondwanan glaciation in a paleotropical carbonate system; it contains good exposures of upper Paleozoic strata. The goal of this study is to construct a robust sequence stratigraphic and isotopic framework that can be examined in the context of the Gondwanan record of glaciation. Lower Permian strata are chosen because emerging evidence suggests the acme of late Paleozoic glaciation occurred during the Early Permian (Isbell et al., 2003; Fielding et al., 2008a-c). The data and interpretations presented here allow examination of the far-field effects of polar glaciation during the acme and subsequent demise of the late Paleozoic ice age (LPIA). Results suggest that carbonate strata from the Orogrande Basin responded to glacial epochs evident in the Gondwanan record, which implies that these glacial events were of global significance.

RECENT RESEARCH ON THE LATE PALEOZOIC ICE AGE

The LPIA was the longest icehouse period of the Phanerozoic, and previous studies have viewed this icehouse world as a single protracted glacial epoch, lasting between 60-80 Ma (e.g., Veevers and Powell, 1987; Crowley and Baum, 1991; 1992; Frakes et al., 1992). However, a review of Gondwanan records by Isbell et al. (2003) demonstrated that the LPIA could be divided into three distinct periods of glaciation (two in the Carboniferous and one in the latest Carboniferous-Sakmarian), separated by times
Figure 1: Early Permian (~280-290 Ma) paleogeographic maps of the Orogrande Basin. Both maps modified after Blakey (2007). A) Global map showing Pangea and the Orogrande Basin. B) Close-up view of the study area.
of non-glacial conditions. The concept of discrete glacial intervals was further resolved by work in eastern Australia (Fielding et al., 2008a), where the LPIA can be divided into eight distinct glacial epochs (four in the Carboniferous and four in the Permian), each lasting 1-8 million years, which are separated by non-glacial intervals of equal duration. Additionally, studies based on compilations of data across Gondwana indicate that ice centers remained small and restricted to higher elevations during the Carboniferous (Isbell et al., 2003). By contrast, multiple ice centers expanded during the Early Permian in places like eastern Australia, Antarctica, India, southern Africa, South America, and the Middle East, which suggests an acme in the earliest Permian (Isbell et al., 2003; Fielding et al., 2008b; 2008c and references therein).

An Early Permian glacial acme is also supported by geochemical proxy records of ice volume, \( p\text{CO}_2 \) (atmospheric carbon dioxide), and temperature. Carbon isotope records from marine carbonates, soil-forming minerals, and sedimentary organic matter record a \( \sim 1\% \) to \( 2\% \) increase in global \( \delta^{13}\text{C} \) values in the Early Permian, which are interpreted as representing a drop in atmospheric \( p\text{CO}_2 \) (Ekart et al., 1999; Montañez et al., 2007; Frank et al., 2008; Birgenheier et al., 2010). Additionally, compiled isotopic data by Frank et al. (2008) suggested a \( \sim 2\% \) increase in \( \delta^{18}\text{O} \) values from low-latitude marine carbonates is consistent with a significant expansion of global ice volume in the earliest Permian.

The idea of a major expansion of glacial ice during the Early Permian is not without controversy. Northern Pangean cyclothem records have been interpreted as recording the most reliable eustatic record of Gondwanan glacial fluctuations, especially for the Carboniferous (Heckel, 1986; 1994; Wright and Vanstone, 2001; Heckel, 2008).
This claim is based mostly on the fact that these predominantly epicontinental records formed in an ice-distal location, free from the effects of isostatic loading from glacial ice. Interpretations made from these paleotropical deposits suggest a peak in glaciation during the Late Pennsylvanian, which is based on larger perceived relative sea-level fluctuations (e.g., Veevers and Powell, 1987; Heckel, 1994; Gonzalez-Bonorino and Eyles, 1995; Crowell, 1999; Heckel, 2008). However, a review by Isbell et al. (2003) concluded that the Carboniferous glacial epochs were not extensive enough to have been the primary mechanism that controlled the well documented base level changes in the Euramerican cyclothem record (Veevers and Powell, 1987; Heckel, 1986; 1994; 2008). Despite recent advancements in the global understanding of the LPIA, the apparent discrepancy between Euramerican and Gondwanan records highlights the continuing need for global correlations.

MOTIVATION FOR THIS STUDY

The impetus behind this study is to identify the far-field impacts of Gondwanan glaciation during the Early Permian. In this context, the Orogrande Basin represents a largely overlooked region that could provide clues regarding the paleotropical response to major fluctuations in sea level, climate, and glacial ice volume during the late Paleozoic. This region contains good exposures of the Pennsylvanian-Permian boundary interval, which represents a time period of major glacial expansion across Gondwana (Fielding et al. 2008b-c). This study represents the first basin-scale sequence stratigraphic and stable isotopic analysis of the Orogrande Basin. The main reason this area has been
understudied likely relates to the proximity of the heavily studied Permian Basin and lack of recoverable hydrocarbon reserves.

The hypothesis tested and falsified in this dissertation is that the growth and decay of Early Permian ice centers is not reflected in the sequence stratigraphic and stable isotopic records of Lower Permian carbonates in the Orogrande Basin. In order to test this hypothesis, this project consisted of three tasks:

1.) Conduct a sedimentologic and facies analysis and place strata into a sequence stratigraphic framework. The sequence stratigraphic framework would then be used to correlate specific sedimentologic responses to individual Gondwanan glacial epochs.

2.) Generate stable isotope records ($\delta^{13}C$ and $\delta^{18}O$) for carbonate strata in the basin. These isotopic trends could then be compared to the record of Gondwanan glaciation in order to determine the paleotropical response to individual glacial epochs.

3.) Conduct a literature review of the most highly resolved paleotropical carbonate successions spanning the Pennsylvanian-Permian boundary. The resulting analysis would allow for synthesis of stratigraphic records, which could then be used to interpret global eustatic changes associated with the onset of major Gondwanan glaciation during the Early Permian.

DISSERTATION STRUCTURE

This dissertation is organized as a series of three manuscripts that have been or will be submitted for publication. The three manuscripts are divided into separate chapters (Chapters 2-4), and each chapter (i.e., manuscript) is formatted according to the
respective journal’s guidelines. Specifically, the formatting guidelines used are from the
*Geological Society of America Bulletin* (Chapter 2), *Sedimentology* (Chapter 3), and the
*Journal of Sedimentary Research* (Chapter 4). Figures and tables are included within the
text of each chapter to help the readability of this dissertation. Chapter 1 (i.e., this
chapter) is an introduction to the project, and Chapter 5 presents conclusions and
synthesizes the results from the three manuscripts. The Appendices (A-D) include
additional figures and information not contained in Chapters 2-4.

CHAPTER 2

Chapter 2 is entitled “Sequence stratigraphic and facies analysis of the Orogrande
Basin (New Mexico, USA): Imprint of late Paleozoic glaciation,” by Jesse T. Koch and
Tracy D. Frank. This paper examines the sedimentologic and lithofacies distributions of
uppermost Pennsylvanian-Lower Permian carbonate strata in south-central New Mexico.
The result is a large-scale sequence stratigraphic framework that can be broadly
correlated to discrete glacial events from Gondwana (Fielding et al. 2008a-c). A major
unconformity is present at the Pennsylvanian-Permian boundary, which coincided with
the onset of major Gondwanan glaciation. Additionally, two transgressive events during
the mid-Sakmarian and mid-Artinskian correlate with the end of major glacial epochs
(Fielding et al., 2008a-c). Results from this study suggest that Early Permian glacial
epochs identified in the Gondwanan record created a stratigraphic imprint in the
Orogrande Basin. This suggests that these glacial epochs were of global significance.
CHAPTER 3

Chapter 3 is entitled “Carbon isotope stratigraphy of uppermost Pennsylvanian-Lower Permian marine carbonates in south-central New Mexico: Implications for glaciation, sea level, and depositional patterns,” by Jesse T. Koch and Tracy D. Frank. This study examined the $\delta^{13}C$ record from uppermost Pennsylvanian-Lower Permian carbonates in southern New Mexico. The distribution of $\delta^{13}C$ and $\delta^{18}O$ values, petrographic, and outcrop data in Gzhelian through lower Sakmarian strata suggest periods of subaerial exposure, which incorporated lower $\delta^{13}C$ values into carbonate strata during meteoric diagenesis. $\delta^{13}C$ values return to global averages during the early Sakmarian-Kungurian (Frank et al., 2008 and references therein). Two smaller isotope excursions (1.5‰ decrease in $\delta^{13}C$ values) are inferred for upper Sakmarian-middle Artinskian and Kungurian strata. These lower $\delta^{13}C$ values are not associated with subaerial exposure, but rather a facies change to shallower (and more restricted circulation) conditions that coincided with periods of renewed expansion of glacial ice across parts of Gondwana (Fielding et al., 2008a-c). Results demonstrate the global significance of glacial events recorded in eastern Australia, and that paleotropical carbonate systems responded to these major climate changes during the Early Permian.

CHAPTER 4

Chapter 4 is entitled “The Pennsylvanian-Permian transition in the low-latitude carbonate record and the onset of major Gondwanan glaciation,” by Jesse T. Koch and Tracy D. Frank. This study is a literature review and synthesis of global stratigraphic records for the paleotropics spanning the Pennsylvanian-Permian boundary. Stratigraphic
data suggest that carbonate successions record a significant eustatic drop at or near the Pennsylvanian-Permian boundary, which coincides with the initiation of major glaciation across Gondwana (Fielding et al., 2008b-c). Additionally, a series of mid-Sakmarian-Kungurian transgression events are interpreted to reflect the asynchronous deglaciation of Gondwana. The global stratigraphic framework developed in this paper will allow for better correlations among stratigraphic records from Gondwana and Euramerica, which will ultimately improve our understanding of the far-field effects of glaciations.

APPENDICES A-D

Appendices A-D contain materials that are not included in the three papers (Chapters 2-4). Appendix A, entitled “Graphic columns of outcrops and cores,” includes detailed graphic logs (with descriptions) of all outcrops and cores analyzed for this dissertation. Appendix B, entitled “Additional lithofacies photographs,” contains additional field images of important features (e.g., phylloid algae bioherms, high-frequency sequences, and shallowing upward cycles). Appendix C, entitled “Miscellaneous figures not included in publications,” contains additional stable isotope ($\delta^{13}C$ and $\delta^{18}O$ trends) and GIS figures. Appendix D, entitled “Miscellaneous fusulinid identifications,” contains unpublished fusulinid identifications by Dr. Greg Wahlman (Wahlman Geological Services, Houston, TX) from outcrop and core samples.
REFERENCES


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CHAPTER 2:
SEQUENCE STRATIGRAPHIC AND FACIES ANALYSIS OF THE
OROGRANDE BASIN (NEW MEXICO, USA): IMPRINT OF LATE
PALEOZOIC GLACIATION

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ABSTRACT

The Permian transition from deep icehouse to greenhouse conditions represents an important interval of time at the end of the late Paleozoic ice age, however, relatively few sequence stratigraphic studies have documented these changes. This study utilizes sequence stratigraphic methodology to construct a paleotropical relative sea level record of carbonate strata in the Orogrande Basin, which is used to assess the global significance of recently resolved stratigraphic records that indicate the timing, nature, and duration of multi-million year Gondwanan glacial epochs. These records suggest a synchronous buildup of glacial ice during the Early Permian. Strata are divided into five facies associations that include terrigenous clastic deposits, nearshore restricted carbonates, open marine carbonates, phylloid algae/Tubiphytes bioherms, and deeper water facies that may have been deposited below the photic zone. A major subaerial unconformity (i.e., sequence boundary) separates Pennsylvanian from Permian strata, and significant transgressive events in the mid-Sakmarian and mid-Artinskian are recorded as shifts to deeper depositional facies (i.e., large bioherms and fusulinid packstones). Development of the Pennsylvanian-Permian sequence boundary coincided with major ice expansion across Gondwana, and likely represents a eustatic drop. Transgressions correspond to deglaciation events across Gondwana and likely represent post-glacial eustatic rise. These results suggest that stratigraphic changes in the Orogrande Basin correlate with large-scale (multi-million year) glacial epochs and deglaciation events from Gondwana. This study demonstrates that recently documented eastern Australian glacial epochs were of global significance, which enhances our understanding of how the paleotropics responded to significant glaciation during the late Paleozoic icehouse world.
INTRODUCTION

Understanding the paleotropical response to major climate perturbations during ice ages is a challenging task, however, relating ice-proximal and ice-distal records is important to help our understanding of the degree of synchronicity of glaciations. One such example is the late Paleozoic ice age (LPIA), which represents the longest icehouse interval of the Phanerozoic. Despite all that is known about this icehouse world, inherent complexities in the climate system and poorly constrained chronostratigraphic control for upper Paleozoic strata make global comparisons difficult.

The main goal of this study is to use sequence stratigraphy to construct a paleotropical relative sea level record and compare it with recently published high-resolution stratigraphic records from Gondwana (Fielding et al., 2008a; 2008b). The eastern Australian record is perhaps the most highly resolved from all of Gondwana due to detailed biostratigraphic/radiogenic age control. It has the potential to be correlated with the Euramerican record, and more specifically for this report, the Orogrande Basin in New Mexico. In eastern Australia, eight distinct glacial epochs are recognized. The glacial epochs range from 1-8 million years in duration, and they are separated by non-glacial intervals of equal duration. Each of these glacial epochs contains multiple 4th and 5th-order advance and retreat cycles (Birgenheier et al., 2009). The four non-overlapping Carboniferous glaciations (termed C1-C4) occurred from the Late Mississippian through the mid-Pennsylvanian. For the Permian, four discrete glaciations (P1-P4) occurred from the Early Permian through the mid-Permian, with the P1 and P2 glaciations being the most widespread. Recently published work from Gondwana suggests that the maximum extent of glacial ice occurred during the Early Permian (Fielding et al., 2008b), which
contradicts previous Euramerican cyclothem studies suggesting a glacial acme during the Late Pennsylvanian (Heckel, 1986; 1994; 2008).

The Euramerican cyclothem framework for Carboniferous strata has long been considered an accurate and reliable eustatic record of Gondwanan glacioeustasy (Heckel, 1986; 1994; Wright and Vanstone, 2001; Heckel, 2008). These studies have created relative sea level curves, and previous researchers have had some success correlating far-field sea-level changes to the onset and demise of the LPIA. For example, Wright and Vanstone (2001) suggested the onset of glacioeustasy occurred at 330 Ma (Viséan), which was recorded in low-latitude cyclothem deposits. Isbell et al. (2003) reviewed the Gondwanan stratigraphic record and determined that the LPIA consisted of three distinct glacial epochs (i.e., Glacial I, II, and III); they concluded that the pre-Permian glacial epochs (Glacial I and II) were not extensive enough to have been the primary mechanism that controlled the well documented base level changes in the Euramerican cyclothem record (Heckel, 1986; 1994; 2008).

In this paper, uppermost Pennsylvanian-Lower Permian strata from the Orogrande Basin in south-central New Mexico are examined. This basin contains some of the most continuous marine carbonate sections of Upper Pennsylvanian-Lower Permian strata in the southwestern U.S. (Jordan, 1975; Raatz, 2002; Wahlman and King, 2002). Carbonate strata from this region are analyzed in terms of sequence stratigraphy, facies distributions, and paleoenvironmental interpretations. Previous sequence stratigraphic interpretations have been restricted to the terrigenous clastic units (Abo Formation; see Mack et al., 2003; Mack, 2007), or limited to relatively short intervals of time (Holder Formation; see
Rankey et al., 1999). Uppermost Pennsylvanian-Lower Permian carbonates throughout the basin remain largely understudied with respect to sequence stratigraphy.

This study helps to provide a link between paleoequatorial regions and ice-proximal localities during the acme and subsequent demise of the late Paleozoic ice age, which ultimately helps to improve our global understanding of this unique and complex icehouse world. Comparing Gondwanan and Euramerican stratigraphic records helps to enhance our understanding of how paleoequatorial regions responded to major climate perturbations during global icehouse conditions.

GEOLOGIC BACKGROUND

Orogrande Basin

The Orogrande Basin of south-central New Mexico was a relatively shallow basin that formed in Middle Mississippian time (Fig. 1; Seager et al., 1976; Candelaria, 1988). The basin was located within 5° of the equator during the Late Pennsylvanian-Early Permian (Ross and Ross, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000). Regional studies suggest that the Orogrande Basin was sporadically connected to the Permian Basin (e.g., Northwest Shelf/Delaware Basin) in Texas and the Pedregosa Basin in Mexico (Jordan, 1975; Candelaria, 1988; Raatz, 2002). During the Pennsylvanian, regional tectonic activity created zones of uplift in south-central New Mexico (e.g., Pedernal, Diablo, Florida uplifts), which appear to be associated with the Ouachita-Marathon orogenic belt and the assembly of Pangea (Candelaria, 1988; Raatz, 2002). These uplifted areas reached maximum relief during the Late Pennsylvanian-earliest Permian, which resulted in significant volumes of clastic sediment being deposited into
Figure 1: Study area map of the Orogrande Basin; modified after Candelaria (1988). Locations of outcrops are denoted with black dots and core locations are denoted with red dots.
the Orogrande Basin (Jordan, 1975; Candelaria, 1988). These terrigenous clastics were shed into the basin, primarily from the north and east; open marine carbonate deposition dominated the southern part of the study area (Fig. 1; Jordan, 1975; Candelaria, 1988; Raatz, 2002). By the end of the Wolfcampian accommodation in the basin was limited due to clastic/carbonate infill and a reduction in subsidence (Jordan, 1975; Candelaria, 1988; Raatz, 2002).

Stratigraphy

Deposits in the Orogrande Basin range from open marine tropical carbonate strata in the south to mixed siliciclastic-carbonate strata in the northern portion of the basin (Jordan, 1975; Candelaria, 1988; Rankey et al., 1999). Upper Virgilian strata (and adjoining Northwestern Shelf region) consist of the clastic dominated Bursum (Laborcita) Formation; the phylloid algae bioherm dominated Bough units, and the mixed clastic-carbonate Panther Seep/Holder Formations (Fig. 2; Cys and Mazzullo, 1985; Schoderbek and Chafetz, 1988; Candelaria, 1988; Raatz, 2002). Coeval units in the western part of the basin (Robledo Mountains) are considered to be part of the Madera Group (Kues, 2001).

Permian strata consist primarily of the open to restricted marine carbonate dominated Hueco Group and the terrestrial clastic dominated Abo Formation (Fig. 2; Jordan, 1975; Seager et al., 1976; Mack and James, 1986; Candelaria, 1988; Mack et al., 1988; Wahlman and King, 2002; Mack et al., 2003; Mack, 2007). A zone of transition occurs between Hueco and Abo strata in the central portion of the basin where the units interfinger. In this transitional zone, marine carbonates of the Hueco Group show
Figure 2: Chronostratigraphic framework for the study area. Time scale is based on Gradstein et al. (2004). Conodont and global fusulinid zones after Gradstein et al. (2004) and references therein. U.S. Midcontinent fusulinid zones after Ross and Ross (1988; 1995).
characteristics consistent with a more restricted environment and the siliciclastic strata of
the Abo Formation show a greater marine influence, including evidence for estuarine
deposition (Jordan, 1975; Wahlman and King 2002; Mack et al., 2003).

The Pennsylvanian-Permian boundary discussed in this report is based on the
Global Stratotype Section and Point (GSSP) for the base of the Permian System, which
has been placed in Kazakhstan (Davydov et al., 1998). Conodont and fusulinid studies
have correlated the Pennsylvanian-Permian boundary in Kazakhstan with stratigraphic
sections in the U.S. Midcontinent, the Permian Basin in Texas, and the Orogrande Basin
in New Mexico (Ross, 1963; Ritter, 1995; Wahlman and King, 2002). This boundary
shift reassigns strata originally considered lower Wolfcampian to the Upper
Pennsylvaniaian. Units such as the Bursum Formation and Bough units that have
traditionally been viewed as Permian are now assigned to the Upper Pennsylvaniaian
(Wahlman and King, 2002). Rasbury et al. (1998) reported radiogenic ages (U-Pb) from
paleosol calcite in the Sacramento Mountains (Holder and Bursum-Laborcita Formations)
of $302.2 \pm 2.4$ Ma for the Pennsylvanian-Permian boundary. Utilizing current GSSP
numerical ages (Ramezani et al. 2007), a reevaluation of the original U-Pb ages and
stratigraphic positions presented in Rasbury et al. (1998) would place the System
boundary at the top of the Bursum (Laborcita) Formation in the Orogrande Basin.

Wolfcampian strata unconformably overlie Pennsylvanian deposits (Schoderbek
and Chafetz, 1988; Wahlman and King, 2002). This erosional unconformity near the
Pennsylvanian-Permian boundary appears to extend across much of the southwestern
United States (Ross, 1986). The stratigraphic intervals analyzed are the Upper
Pennsylvaniaian strata of the Holder Formation, informal Bough members (including the
Saunders unit, and Bursum Formation. Lower Permian strata examined are the Hueco Group and Abo Formation. Figure 2 shows a chronostratigraphic framework of the units examined for this study, and Figure 1 shows their respective locations.

*Upper Pennsylvanian Units*

Upper Pennsylvanian strata in the Orogrande Basin are primarily divided into the Upper Virgilian Holder Formation (Panther Seep-equivalent) and the uppermost Virgilian Bursum Formation. The Holder Formation is exposed in the Sacramento Mountains and consists of alternating open marine limestone and terrestrial clastic units deposited as cyclothems (Rankey et al., 1999). The unit reaches a maximum thickness of 275 m in the northern Sacramento Mountains, and biostratigraphic data from fusulinids suggest a Virgilian age (Cline, 1959; Schoderbek and Chafetz, 1988; Rankey et al., 1999). Holder strata are the shallow shelf equivalent to the Panther Seep Formation (in the Franklin and San Andres Mountains), which was deposited in the central to southern part of the basin (Schoderbek and Chafetz, 1988).

A small portion of the Delaware Basin (i.e., Northwest Shelf) extends into southeastern New Mexico. Informal Bough members (A-D) and a Saunders unit have been described as part of the Lower Permian Hueco Formation by Cys and Mazzullo (1985) and Malek-Aslani (1985). However, redefinition of the boundary in the region indicates these units are now considered latest Pennsylvanian in age (Wahlman and King, 2002). The Bough members reach a maximum thickness of ~120 m and consist mainly of phylloid algae bioherm and open shelf deposits (Cys and Mazzullo, 1985; Malek-Aslani, 1985).
The Bursum Formation (equivalent to the Laborcita Formation) consists of open marine limestone and terrigenous clastic units that range from thin, incomplete sections in the Dona Ana Mountains (Seager et al., 1976) to more complete sections that reach a maximum thickness of 450 m in the Jarilla Mountains (Candelaria, 1988). Bursum deposits are localized and do not occur across the entire basin (Jordan, 1975; Raatz, 2002). Biostratigraphic studies utilizing fusulinids assign Bursum strata (and equivalent units) to the uppermost Pennsylvanian (Steiner and Williams, 1968; Wahlman and King, 2002).

Hueco Group

The Lower Permian Hueco Group is present across the Orogrande Basin, but is thickest in the south (Jordan, 1975). These open marine carbonate strata thin towards the north and become more restricted as they interfinger with the terrestrial Abo Formation. Biostratigraphic studies using fusulinids suggest all Hueco units were deposited during Wolfcampian time (Williams, 1966; Simo et al., 2000; Wahlman and King, 2002; Krainer et al., 2005).

In the southern part of the basin (Franklin and Hueco Mountains), Hueco strata thicken to over 800 m and are divided into three distinct formations, the Hueco Canyon, Cerro Alto, and Alacran Mountain Formations (Williams, 1963; Jordan, 1975; Candelaria, 1988; Wilson and Jordan, 1988; Raatz, 2002). The Hueco Canyon Formation (up to 400 m thick) consists of bioherm units and open marine wackestones to packstones deposited in a shelf to shoaling environment (Jordan, 1975; Wilson and Jordan, 1988). In the Hueco Mountains the Hueco Canyon Formation is underlain by a conglomeratic unit.
(Powwow Conglomerate; up to 30 m thick), which is related to clastic shedding from the adjacent Pedernal Uplift (Jordan, 1975; Candelaria, 1988; LeMone, 1988). Powwow clastics are not present in coeval strata in the Franklin Mountains (Jordan, 1975). The Cerro Alto Formation (less than 300 m thick) consists primarily of fossiliferous wackestones and phylloid algae bioherm units, which were deposited in an open marine shelf environment; clastics are more abundant than in the Hueco Canyon Formation (Jordan, 1975; Wilson and Jordan, 1988). The overlying Alacran Mountain Formation (less than 200 m thick) is the youngest unit in the Hueco Group. This unit consists of bioherms and open marine wackestones-packstones (Jordan, 1975; Wilson and Jordan, 1988). In the Hueco Mountains a ~30 m shale unit (Deer Mountain Red Shale Member) is present, which has been interpreted to represent clastic influx from adjacent uplifted areas (Williams, 1966; Jordan, 1975; Wilson and Jordan, 1988). The Deer Mountain Shale is not present in coeval strata in the Franklin Mountains (Jordan, 1975).

In the central to northern part of the basin (Robledo, Dona Ana, and San Andres Mountains), Hueco units have been designated a formation and are subdivided into lower, middle, and upper members (Jordan, 1975; Seager et al., 1976; Mack et al., 1988; Raatz, 2002; Wahlman and King, 2002). These members are distinguishable based on faunal changes in the carbonate strata. The lower Hueco (~150 m thick) consists mainly of micrite to fossiliferous packstone deposited in a predominantly carbonate shoal environment (Jordan, 1975; Wahlman and King, 2002). The middle Hueco (~80 m thick) is dominated by micrite to fossiliferous wackestone deposited in open marine to restricted environment (Wahlman and King, 2002). The upper Hueco (~100 m thick) consists of wackestones deposited in an open to restricted environment (Seager et al., 1976;
Wahlman and King, 2002). The middle and upper Hueco units are separated by a 150 m thick section of terrestrial to marginal-marine Abo strata (Jordan, 1975; Mack et al., 1988; Wahlman and King, 2002).

In the northernmost part of the Orogrande Basin (Rhodes Canyon in the northern San Andres Mountains), the open to restricted marine Hueco Formation thins to < 120 m and is not divided into individual members. Hueco units in the northern part of the basin interfinger with overlying Abo Formation strata, which creates an Abo/Hueco transition zone (Kottlowski, 1975). Similar Hueco tongues are present in the lower Abo Formation in the Sacramento Mountains (Jordan, 1975).

Abo Formation

The Abo Formation in the central to northern portion of the basin (Robledo, Dona Ana, San Andres, and Sacramento Mountains). This terrigenous clastic dominated formation is observed either as a tongue within a Hueco unit (as observed in the Robledo Mountains) or as an unconformable unit that overlies Hueco Formation or Upper Pennsylvanian strata (Jordan, 1975; Raatz, 2002; Wahlman and King, 2002). The Abo Formation has a maximum thickness of 500 m in the northern part of the basin (Raatz, 2002); it thins to the southeast and is not present in the Franklin and Hueco Mountains (Jordan, 1975). The lowest beds at the type locality consist of thin wackestones with open marine fauna. Up-section from the limestones, terrestrial clastic red beds consisting of sandstones, conglomerates, and shales begin to dominate Abo strata. These units have been interpreted as marginal marine to coastal plain/fluviatile deposits (Jordan, 1975; Mack et al., 2003).
METHODS

This investigation is based on measured sections through five outcrop and four core sections, primarily of the Holder Formation, Bursum Formation, informal Bough members, and the Hueco Group (Figs. 3-8). Outcrops and cores were chosen based on stratigraphic completeness and accessibility. The cores used for this study are housed at the New Mexico Bureau of Geology and Mineral Resources in Socorro, New Mexico. From both cores and outcrops, a total of 243 additional hand samples were analyzed to supplement core and outcrop descriptions. Of the 243 samples, 138 were made into thin-sections and analyzed using standard petrographic procedures. Information from measured sections, core logs, and thin-sections are combined to determine facies associations and depositional environments. The facies associations and depositional environment interpretations are used to develop a sequence stratigraphic framework.

Utilizing the current placement of the Pennsylvanian-Permian boundary (Wahlman and King, 2002), this study analyzes uppermost Pennsylvanian-Lower Permian deposits from all major areas of the Orogrande Basin. The limited lateral extent of outcrops across the Orogrande Basin necessitates that sequence stratigraphic interpretations for certain intervals of time rely, in part, on stratigraphic sections in one area of the basin. For example, the majority of Early Permian interpretations are based on outcrops in the Franklin Mountains because this locality provides the most continuous carbonate record of Hueco deposition in the Orogrande Basin (Jordan, 1975); coeval units in the northern part of the basin are dominated by terrigenous clastics of the Abo Formation. This methodology represents the best possible way to study sequence stratigraphic trends for carbonate strata across the study area.
Figure 3: Graphic column of the Holder/Abo Formations in the northern Sacramento Mountains (Dry Canyon area east of Alamogordo, NM).
Figure 4: Graphic columns of the studied cores from the Northwest Shelf portion of the Delaware Basin. A) Unocal State No. 1-33 Core; B) Tres Papalotes Core; C) Getty No. 5 Williard Beatty Core; D) Nicor DDH-2 Core.
Figure 5: Graphic column of the Hueco/Abo Formations in the northern San Andres Mountains (Rhodes Canyon, White Sands Missile Range). P.S.: Panther Seep Formation.
Figure 6: Graphic column of the Lower Permian Hueco Canyon Formation in the Franklin Mountains. P.S.: Panther Seep Formation.
Figure 7: Graphic column of the Lower Permian Cerro Alto Formation in the Franklin Mountains.
Figure 8: Graphic column of the Lower Permian Alacran Mountain Formation in the Franklin Mountains.
LITHOFACTIES DESCRIPTIONS

Upper Paleozoic rocks in the Orogrande Basin are divided into five facies associations (Table 1). Lithofacies in the study area are divided into both carbonate and clastic units. Each lithofacies is defined based upon dominant lithology, overall fossil content, and sedimentary structures. Clastic units contain both terrestrial and offshore lithologies, and carbonate strata range from a peritidal to an outer ramp setting. Lithofacies are arranged according to inferred water depth on Table 1; the overall depositional profile is described in terms of a carbonate ramp setting. The carbonate ramp model for late Paleozoic tropical carbonate systems is based on many previous studies suggesting a general lack of rimmed platforms during this time period (e.g., James, 1983; Wahlman, 2002).

Subaerial to Nearshore Siliciclastic Facies

Subaerial and nearshore siliciclastic facies include clastic mudrock, fine to medium-grained sandstone, and pebble conglomerates (Table 1). These lithologies were observed in the Sacramento Mountains section, cores, the San Andres Mountains section, and the Franklin Mountains sections (Figs. 3-8, respectively).

Lithofacies 1a: Red Mudrock

Red mudrock and siltstone comprises Lithofacies 1a (Fig. 9a). These deposits are thinly bedded (< 1 cm) to laminated and do not contain marine fossil material. Pedogenic features such as slickensides, blocky structures, and rhizoliths are common. Lithofacies 1a is found in cores and in the San Andres/Sacramento Mountains sections (Figs 3-5).
## TABLE 1. FACIES ASSOCIATIONS FOR UPPER PALEOZOIC STRATA OF THE OROGRANDE BASIN

<table>
<thead>
<tr>
<th>Facies Association</th>
<th>Attributes</th>
<th>Lithofacies Code</th>
<th>Interpreted Environment</th>
<th>Occurrence in Studied Units</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Subaerial to Nearshore Siliciclastic Facies</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Red Mudrock</td>
<td>Thin-beded clastic red bed deposits; pedogenic features (silex, bentonite, etc.) common.</td>
<td>1a</td>
<td>Paleosols and coastal plain environment.</td>
<td>Tres Papaloites Core; Nicor DDH-2 Core; Getty No. 5 Core; San Andres Section; Sacramento Section.</td>
</tr>
<tr>
<td>Fine to Medium-Grained Sandstone</td>
<td>Thin to thick-beded to cross-beded sandstone; rare aridified fragments of crinoids, brachiopods, and fusulind. Basal scum/erosion surfaces common; single storey beds are laterally continuous; often occur as part of mixed elasticarbonate-cemented deposits.</td>
<td>1b</td>
<td>Low accommodation fluvial environment.</td>
<td>Nicor DDH-2 Core; San Andres Section; Sacramento Section.</td>
</tr>
<tr>
<td>Pebble Conglomerite</td>
<td>Thin to thick-beded to cross-beded conglomerate; rare aridified fragments of crinoids, brachiopods, and fusulind. Basal scum/erosion surfaces common.</td>
<td>1c</td>
<td>High-energy fluvial environment.</td>
<td>Nicor DDH-2 Core; San Andres Section; Sacramento Section.</td>
</tr>
<tr>
<td>Siltstone-Mudstone</td>
<td>Thin-beded clastic siltstone/mudstone: rare gastropods and fusulind. pedogenic features absent.</td>
<td>1d</td>
<td>Nearshore shallow marine environment.</td>
<td>Franklin Mountains (Cerro Alto-Alacran Mountain Fmns.).</td>
</tr>
<tr>
<td><strong>Nearshore Restricted Carbonate Facies (Upper Ramp)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Interbedded Microbial Laminitie and Laminated Carbonate Mudstone</td>
<td>Laminated microbial structures and laminated micrite common; fenestral fabrics; contains gastropods, ostracods, and foraminifers.</td>
<td>2</td>
<td>Shallow subtidal to lower intertidal environment.</td>
<td>Franklin Mountains (Panther Seep Fm.; Hueco Canyon Fmns.).</td>
</tr>
<tr>
<td>Sandy Mudstone to Wackestone</td>
<td>Massive to thin-beded deposits with rare fossil material; occasional abraded grains of ostracods or brachiopods; quartz sand present throughout unit.</td>
<td>3a</td>
<td>Shallow restricted marine environment; elastic influenced.</td>
<td>San Andres Section; Sacramento Section; Franklin Mountains Section (Hueco Canyon Fmns.).</td>
</tr>
<tr>
<td>Gastropod/ Ostracod Mudstone to Wackestone</td>
<td>Massive to thin-beded unit dominated by gastropods and/or ostracods; no other fossil taxa observed.</td>
<td>3b</td>
<td>Shallow restricted marine environment.</td>
<td>San Andres Section; Sacramento Section; Franklin Mountains Section (Panther Seep Fm.; Hueco Canyon Fmns.).</td>
</tr>
<tr>
<td><strong>Open Marine Carbonate Facies (Middle-Upper Ramp)</strong></td>
<td></td>
<td></td>
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<tr>
<td>Ooloidal Wackestone, Packstone, and Grainstone</td>
<td>Thin to thick-beded to cross-beded oolitic deposits; nuclei are rounded/abraded open marine taxa; occasional ooids.</td>
<td>4</td>
<td>High energy middle to upper ramp environment.</td>
<td>Unocal State 1-33 Core; Getty No. 5 Core; San Andres Section; Sacramento Section; Franklin Mountains (Hueco Canyon-Cerro Alto-Alacran Mountain Fmns.)</td>
</tr>
<tr>
<td>Fusulinid Packstone to Grainstone</td>
<td>Thick to thin-beded fusulinid-rich units with occasional brachiopods, trilobites, bryozoans, rare phylloid algae, and foraminifers; grains are usually abraded.</td>
<td>5a</td>
<td>Middle to upper ramp setting; open marine salinity; within the euphotic zone.</td>
<td>Unocal State 1-33 Core; Tres Papaloites Core; Getty No. 5 Core; Sacramento Section; Franklin Mountains (Cerro Alto-Alacran Mountain Fmns.).</td>
</tr>
<tr>
<td>Fossiliferous Wackestone, Packstone, and Grainstone</td>
<td>Bedded to massive deposits with diverse open marine fauna; grains often abraded; can contain intraclasts and occasional limestone granule to pebble breccias.</td>
<td>5b</td>
<td>Middle to upper ramp setting; open marine salinity; within the euphotic zone; likely storm-influenced.</td>
<td>Unocal State 1-33 Core; Tres Papaloites Core; Nicor DDH-2 Core; Getty No. 5 Core; San Andres Section; Sacramento Section; Franklin Mountains (Hueco Canyon-Cerro Alto-Alacran Mountain Fmns.).</td>
</tr>
<tr>
<td>Dasyycladacean Algae Wackestone-Packstone</td>
<td>Massive unit dominated by Dasyycladacean Epimastopora algae with other open marine faunas including crinoids, foraminifers, brachiopods, ostracods, and bryozoans.</td>
<td>5c</td>
<td>Middle to upper ramp setting; open marine salinity; within the euphotic zone.</td>
<td>San Andres Section; Sacramento Mountains Section.</td>
</tr>
<tr>
<td>Facies Association</td>
<td>Attributes</td>
<td>Lithofacies Code</td>
<td>Interpreted Environment</td>
<td>Occurrence in Studied Units</td>
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<td><strong>Carbonate Boundstone</strong></td>
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<td>Facies (Middle-Lower Ramp)</td>
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<tr>
<td>Phylloid Algae Floatstone/Boundstone</td>
<td>Massive to thick-bedded units dominated by phylloid algae; some <em>Tubiphytes</em> present along with other open marine taxa such as crinoids, trilobites, brachiopods, and bryozoans.</td>
<td>6a</td>
<td>Bioherm or carbonate mound environment within the photic zone; middle to lower ramp setting; open marine salinity; patch bioherms present on upper ramp.</td>
<td>Unocal State 1-33 Core; Tres Papalotes Core; Nicor DDH-2 Core; Getty No. 5 Core; Sacramento Section; Franklin Mountains (Hueco Canyon-Cerro Alto-Alacran Mountain Fmns.).</td>
</tr>
<tr>
<td><em>Tubiphytes</em>-rich Floatstone/Boundstone</td>
<td>Massive unit dominated by <em>Tubiphytes</em>, some phylloid algae present along with other open marine taxa such as crinoids, trilobites, brachiopods, and bryozoans.</td>
<td>6b</td>
<td>Bioherm or carbonate mound environment within the euphotic zone; middle to lower ramp setting; open marine salinity.</td>
<td>Unocal State 1-33 Core; Tres Papalotes Core; Getty No. 5 Core; Franklin Mountains (Hueco Canyon-Cerro Alto-Alacran Mountain Fmns.).</td>
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<tr>
<td><strong>Offshore Facies (Lower Ramp to Basin)</strong></td>
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<tr>
<td>Crinoid-Brachiopod Wackestone, Packstone, and Grainstone</td>
<td>Thick to thin-bedded and massive deposits with echinoderms, productid brachiopods, bryozoans, foraminifers, rare fusulinids, ostracods, gastropods, and trilobites; distinct lack of photozoans compared to Lithofacies 5a-c.</td>
<td>7</td>
<td>Lower ramp; dysphotic to aphytic; storm-influenced; normal marine salinity.</td>
<td>Tres Papalotes Core.</td>
</tr>
<tr>
<td>Laminated Mudrock/Shale</td>
<td>Thin-bedded light gray to black laminated mudrock/shale; occasional crinoids, brachiopods, and foraminifers.</td>
<td>8</td>
<td>Basin; aphytic zone; likely disoxic conditions.</td>
<td>Tres Papalotes Core; Nicor DDH-2 Core.</td>
</tr>
</tbody>
</table>
Figure 9: Facies images from the Orogrande Basin. See Table 1 for additional details. A) Red mudrock of Lithofacies 1a-Sacramento Mountains section, 195 m; B) Sandstones of Lithofacies 1b; arrow points to a small channel form in the Holder Formation-Sacramento Mountains Section, 164 m; C) Conglomerates of Lithofacies 1c-Abo Formation in the Sacramento Mountains Section, 175 m; D) Siltstones of Lithofacies 1d; arrow points to the siltstone unit, which is bounded by fusulinid packstones-Cerro Alto Formation, 5 m; E) Laminated carbonate mudstones of Lithofacies 2 (denoted with black arrow)-Hueco Canyon Formation, 9.5 m; F) Sandy carbonate wackestone of Lithofacies 3a; quartz grains are denoted with black arrows, San Andres Mountains section, 78 m.
This lithofacies is thinnest in Pennsylvanian units of the Sacramento Mountains and Northwest Shelf where it ranges from 5 cm to ~10 m thick (Holder Formation and Bough units; see Figs. 3-4). Lithofacies 1a is thickest in Permian units (Abo Formation) where it commonly exceeds 10-15 m in thickness (Figs. 3 and 5). This lithofacies is absent in the southern part of the basin (i.e., Franklin Mountains) and is thickest along the basin margins (Fig. 1).

**Lithofacies 1b: Fine to Medium-Grained Sandstone**

Lithofacies 1b is moderately sorted, sub-rounded, fine to medium-grained sandstone (Table 1). Sandstone units are quartz dominated, but also contain marine fossil fragments (e.g., fusulinids, brachiopods, crinoids, etc.). The stratigraphic architecture of the sandstones range from tabular, sheet-like units (up to 1 m thick) to channel-form deposits with erosional bases that are up to 2 m thick (Fig. 9b). Many of the units contain tabular and trough cross-sets up to 5 cm thick. Lithofacies 1b is found in both Pennsylvanian and Permian units (Bursum Formation and Abo Formation) from the Nicor DDH-2 core (Fig. 4d) and Sacramento Mountains section (Fig. 3).

**Lithofacies 1c: Pebble Conglomerate**

Clast-supported, sub-rounded/rounded, pebble conglomerate comprises Lithofacies 1c (Table 1, Fig. 9c). These units are similar to the sandstone strata of Lithofacies 1b in that they contain fossil fragments and range from tabular to channel-form deposits, but attain greater thickness ranging from 1-9 m thick. Planar bedding to weak tabular cross bed sets up to 10 cm thick are common. Lithofacies 1c is found in
both Pennsylvanian and Permian strata (Bursum Formation, Holder Formation, and Abo Formation) in the Nicor DDH-2 core and Sacramento/San Andres Mountains sections (see Figs. 3, 4d, and 5). This unit is not observed in the central portion of the basin (i.e., Franklin Mountains).

Lithofacies 1d: Thin-Bedded Siltstone and Mudrock

Lithofacies 1d is light gray to tan colored, thin-bedded siltstone and mudrock (Table 1, Fig. 9d). These deposits range from laminated to 5 cm thick beds. Sparse marine fossils are found in the unit (e.g., gastropods and fusulinids). This unit is observed interbedded with carbonate strata in the Permian portion of the Franklin Mountains sections (Figs. 6-8), specifically the Cerro Alto and Alacran Mountains Formations. Lithofacies 1d is also found in the Upper Pennsylvanian strata below the Hueco Canyon Formation in the Franklin Mountains (Fig. 6).

Nearshore Restricted Carbonate Facies (Upper Ramp)

Nearshore carbonate facies include laminated carbonate mudstone, sandy mudstone to wackestone, and gastropod/ostracod mudstone to wackestone (Table 1). These lithologies were observed in the Sacramento Mountains section, the San Andres Mountains section, and the Franklin Mountains sections (Figs. 3, 5-6).

Lithofacies 2: Interbedded Microbial Laminite and Laminated Carbonate Mudstone

Lithofacies 2 is laminated carbonate mudstone (Table 1, Fig. 9e). Beds are up to 5 cm thick. Sparse fossils include gastropods, ostracods, and green algae. Fenestral
fabric and laminated microbial structures are common. Lithofacies 2 occurs in the Upper Pennsylvanian and lowermost Permian strata in the Franklin Mountains (Panther Seep and Hueco Canyon Formations; Fig. 6). At this location in the Franklin Mountains, a total of 5 m of Lithofacies 2 is present.

Lithofacies 3a: Sandy Mudstone to Wackestone

Lithofacies 3a consists of sandy carbonate mudstone to wackestone. It is commonly massive to weakly bedded (~10 cm thick). Sand consisting mainly of fine to medium-grained quartz (Fig. 9f) occurs throughout the unit. Fossil debris is generally rare, but occasional abraded grains of ostracods or brachiopods occur. Lithofacies 3a is found in the Sacramento Mountains section (Holder/Abo Formations), San Andres Mountains section (Panther Seep, Hueco, Abo Formations), and Hueco Canyon Formation in the Franklin Mountains (Figs. 3, 5, and 6).

Lithofacies 3b: Gastropod/Ostracod Wackestone

Gastropod/ostracod mudstones to wackestones comprise Lithofacies 3b. As with Lithofacies 3a, it is commonly massive to weakly bedded (~10 cm thick). Gastropods and ostracods are abundant (Fig. 10a). Other fossil debris is generally rare, but occasional abraded grains of brachiopods and phylloid algae occur. Lithofacies 3b is found in the Sacramento Mountains section (Abo Formation), San Andres Mountains section (Panther Seep and Abo Formations), and Panther Seep/Hueco Canyon Formations in the Franklin Mountains (Figs. 3, 5, and 6).
Figure 10: Facies images from the Orogrande Basin. See Table 1 for additional details.

A) Gastropod (denoted with black arrow) wackestone of Lithofacies 3b-San Andres Mountain section, 4 m; B) Algal coated grains of Lithofacies 4; arrow points to a coated phylloid algae grain-Hueco Canyon Formation, 190 m; C) Fusulinid packstone of Lithofacies 5a-Cerro Alto Formation, 5 m; D) Fossiliferous packstone of Lithofacies 5b; upper right arrow points to a Tubiphytes grain, arrow in the center points to a bryozoan grain, and arrow in the lower right points to an echinoderm- Getty No. 5 Core, 3 m; E) Epimastopora algae (black arrows) of Lithofacies 5c-Sacramento Mountains section, 95 m; F) Phylloid algae grain (black arrow) within Lithofacies 6a, Sacramento Mountains section, 68 m.
Open Marine Carbonate Facies (Middle-Upper Ramp)

Open marine carbonate facies deposited in an upper to middle ramp environment include oncoidal wackestone-grainstone, fusulinid packstone-grainstone, fossiliferous wackestone-grainstone, and Dasycladacean algae wackestone-packstone (Table 1). There is an abundance of photozoans in this facies association (e.g., phylloid algae, fusulinids, dasycladacean green algae), which distinguishes it from offshore carbonate facies (Lithofacies 7; see Table 1). These lithologies were observed in all outcrop and core sections (Figs. 3-8).

Lithofacies 4: Oncoidal Wackestone, Packstone, and Grainstone

Lithofacies 4 consists of skeletal wackestone to grainstone (Table 1). This unit is very common throughout the study area, where it forms massive to planar bedded units up to 2 m thick (Figs. 3-8). Fossil content (commonly abraded) is abundant with a diverse and open marine fauna including: phylloid algae, brachiopods, foraminifera, fusulinids, gastropods, ostracods, echinoderms, bryozoans, and trilobites. The distinguishing characteristic of Lithofacies 4 is a calcimicrobial coating (*Girvanella*) on the majority of grains (Fig. 10b).

Lithofacies 5a: Fusulinid Packstone to Grainstone

Fusulinid packstone to grainstone comprises Lithofacies 5a (Fig. 10c). This lithofacies forms planar bedded units up to 2 m thick. In the Franklin Mountains, it is commonly interbedded with Lithofacies 1d (see Figs. 10c). Although other fossil debris is present (e.g., brachiopods, echinoderms, foraminifera), the dominant taxa are large
fusulinids. Lithofacies 5a is common in the Sacramento Mountains section (Holder Formation), the cores, and in the Cerro Alto/Alacran Mountain Formations in the Franklin Mountains (see Figs. 3, 4, 7 and 8).

Lithofacies 5b: Fossiliferous Wackestone, Packstone, and Grainstone

Lithofacies 5b consists of bioturbated fossiliferous wackestone to grainstone. This lithofacies forms massive to planar bedded units up to ~30 m thick, and contains a diverse open marine fauna including phylloid algae, corals, dasycladacean green algae, brachiopods, foraminifera, fusulinids, gastropods, ostracods, echinoderms, bryozoans, and trilobites (Fig. 10d). This facies is commonly associated with Lithofacies 4 throughout the study area. Occasionally, Lithofacies 5b contains thin (< 0.5 m thick) limestone breccias (granule to pebble-sized grains), but these deposits are not prevalent enough to be considered a separate lithofacies (see Figs. 3-6). This lithofacies is the most abundant carbonate unit and is found in all sections studied (Figs. 3-8).

Lithofacies 5c: Dasycladacean Algae Wackestone and Packstone

Lithofacies 5c is similar to Lithofacies 5b except that the dominant taxa are Dasycladacean/Epimastopora green algae (Fig. 10e). This lithofacies forms massive to bedded units that are less than 2 m thick. Besides the dominant taxa, other open marine fossils are abundant including: phylloid algae, Tabulate/Rugose corals, brachiopods, foraminifera, fusulinids, gastropods, ostracods, echinoderms, bryozoans, and trilobites. Lithofacies 5c is found in the Sacramento Mountains section (Holder Formation) and the Panther Seep Formation in the San Andres Mountains section (see Figures 3 and 5).
Carbonate Boundstone Facies (Middle-Lower Ramp)

Carbonate boundstone facies (middle to lower ramp setting) include phylloid algae floatstone/boundstone and Tubiphytes-rich floatstone/boundstone (Table 1). This facies association is found in the majority of sections studied, except the San Andres Mountains section (Figs. 3-4, 6-8).

Lithofacies 6a: Phylloid Algae Floatstone/Boundstone

Lithofacies 6a is a phylloid algae floatstone/boundstone (Table 1). Phylloid algae is the dominant taxon (see Fig. 10f), with other open marine fossils such as brachiopods, foraminifera, fusulinids, gastropods, ostracods, echinoderms, bryozoans, and trilobites common. This lithofacies commonly forms erosion resistant cliffs up to 30 m thick and is often associated with Tubiphytes dominated strata of Lithofacies 6b. Micrite commonly fills the voids between phylloid algae plates and Tubiphytes grains throughout this unit. Lithofacies 6a is found in all sections studied, except the San Andres Mountains section (see Figs. 3-8).

Lithofacies 6b: Tubiphytes-Rich Floatstone/Boundstone

This lithofacies is similar to Lithofacies 6a, except that Tubiphytes is the dominant taxon (Table 1; Fig. 11a). Other marine taxa are less abundant, but include mainly brachiopods, foraminifera, echinoderms, and bryozoans. This lithofacies commonly forms erosion resistant cliffs up to 30 m thick and is often associated with phylloid algae dominated strata of Lithofacies 6a. Lithofacies 6b is found in all sections...
Figure 11: Facies images from the Orogrande Basin. See Table 1 for additional details. A) Tubiphytes (black arrows) boundstone of Lithofacies 6b-Tres Papalotes Core, 5.7 m; B) Productid brachiopod (black arrow) of Lithofacies 7-Tres Papalotes Core, 16.4 m; C) Laminated gray clastic shale of Lithofacies 8-Nicor DDH-2 Core, 38 m.
studied, except the San Andres Mountains section, but is most common in the Franklin Mountains and the Northwest Shelf (see Figs. 3-8).

**Offshore Facies (Lower Ramp to Basin)**

Offshore facies (lower ramp) include fossiliferous wackestone-grainstone and laminated mudrock and shale (Table 1). This facies association is found in the Tres Papalotes and Nicor DDH-2 cores (Fig. 4).

*Lithofacies 7: Crinoid-Brachiopod Wackestone, Packstone, and Grainstone*

Lithofacies 7 consists of crinoid-brachiopod wackestone to grainstone that is commonly bioturbated (Table 1). This lithofacies forms massive to weakly bedded units that are generally less than 10 cm thick. Common open marine taxa present include large productid brachiopods (see Fig. 11b), crinoids, foraminifera, gastropods, ostracods, bryozoans, and trilobites. This lithofacies is unique from Lithofacies 5 because there is a distinct lack of photozoans, and an increase in productid brachiopods, echinoderms, and bryozoans that dominate over photozoan benthos. Lithofacies 7 is restricted to the Tres Papalotes core (Bough/Saunders units) from the Northwest Shelf (see Fig. 4b).

*Lithofacies 8: Laminated Mudrock and Shale*

Lithofacies 8 consists of dark gray/black laminated mudrock/shale (Fig. 11c). This lithofacies forms generally unbioturbated units that are up to 2 m thick. Although fossils are present locally (echinoderms, and brachiopods), it is generally fossil-poor. Lithofacies 8 is commonly associated with Lithofacies 7 or 1. It is observed in the Tres
Papalotes core (Bough/Saunders units) and in the Abo Formation of the Nicor DDH-2 core (see Figs. 4b and 4d).

**DEPOSITIONAL ENVIRONMENTS**

The depositional profile of the Orogrande Basin has been interpreted previously as a series of narrow shallow shelves rimming positive tectonic elements, such as the Pedernal Uplift in the Sacramento Mountains (Rankey et al., 1999). Widespread Upper Pennsylvanian-Lower Permian shallow-water carbonate deposition in the center of the basin (Franklin Mountains) suggests the Orogrande Basin was relatively shallow (Jordan, 1975; Candelaria, 1988). It has been suggested that there was a general lack of rimmed platforms and reef framework builders during the late Paleozoic (e.g., James, 1983; Wahlman, 2002). Rather than large reefs, tropical carbonate environments were characterized by deposition in a ramp setting that contained organic bioherm structures dominated by phylloid algae and *Tubiphytes* (Wahlman, 2002; Forsythe, 2003). These previous inferences are consistent with the depositional style in the Orogrande Basin (Jordan, 1975; Candelaria, 1988), and the depositional profile is herein interpreted as a homoclinal carbonate ramp (Fig. 12). The discussions in the following sections are ordered according to interpreted water depth (i.e., shallow to deeper settings).

**Subaerial to Nearshore Siliciclastic Facies**

Red clastic mudrocks (Lithofacies 1a) locally show a pedogenic overprint, suggesting deposition in a subaerial environment dominated by soil-forming processes (Retallack, 1990). The erosionally based sandstone and conglomerate units (Lithofacies
Figure 12: Idealized Upper Pennsylvanian-Lower Permian depositional profile showing interpreted facies distribution along a carbonate ramp setting for the Orogrande Basin. FWB: fair-weather wave base; SWB: storm wave base. Lithofacies (L.F.) correspond to those found on Table 1. Profile and facies distribution interpreted after discussions in Flügel (2004).
Lithofacies 1 (a-c) commonly forms cyclothem deposits with open marine carbonates (Holder Formation; see Fig. 3), which further suggests deposition in a low-accommodation setting. Since these clastics were deposited in proximity to an ancient shoreline in a low-accommodation setting, they most likely represent an ancient coastal plain environment. This interpretation is consistent with previous studies of the clastic units in the Orogrande Basin, such as the Bursum Formation, Abo Formation, and portions of the Holder Formation (Wilson, 1967; Jordan, 1975; Goldstein, 1988; Rankey et al., 1999; Mack et al., 2003).

The light gray to tan colored, locally fossiliferous, and thin bedded mudrocks/siltstones of Lithofacies 1d are distinct from the pedogenically-altered, red coastal plain deposits of Lithofacies 1a. Additionally, the thin bedded clastics of Lithofacies 1d are cyclically interbedded with shallow marine carbonates of Lithofacies 5 (a-c), which suggest deposition in a shallow marine setting during relative lowstands of sea level. The source of clastic material was likely adjacent uplifted areas, such as the Pedernal Uplift around the edges of the basin (Candelaria, 1988). However, it is important to note that changes in large-scale climatic regimes (tropical cyclones, increased seasonality, etc.) during the Early Permian could have triggered seasonal storm events that brought these clastics into the center of the basin which would not necessitate a relative change in sea level (Tabor and Montañez, 2002; Tabor, 2007; Tabor et al., 2008). These clastic deposits are present in the Franklin Mountains (see Figs. 6-8), and have been previously interpreted as terrestrial-derived clastics in a nearshore marine setting (Jordan, 1975).
Nearshore Restricted Carbonate Facies (Upper Ramp)

The lack of open marine fauna, presence of fenestral fabric, and microbially influenced sedimentation (laminations) in Lithofacies 2 is consistent with deposition in a shallow, restricted carbonate environment (Wilson, 1975; Flügel, 2004). There is a lack of desiccation features (tepees and mudcracks), which further suggests deposition in a lower intertidal to extremely shallow subtidal zone, rather than a supratidal/upper intertidal environment (Flügel, 2004). This lithofacies occurs near the Pennsylvanian-Permian boundary in the Franklin Mountains (uppermost Panther Seep Formation), which suggests a relative lowstand of sea level compared to overlying deeper deposits of the Hueco Group (Jordan, 1975). As such, the laminated carbonate mudstone of Lithofacies 2 is interpreted to record a shallow environment, dominated by peritidal carbonate deposition in an upper ramp setting (Table 1; Fig. 12).

Due to the presence of clastics and low diversity/high abundance organisms such as gastropods and ostracods typical of biostressed environments, Lithofacies 3 (a-b) is herein interpreted as a restricted/hypersaline lagoon environment that was proximal to a clastic source (Flügel, 2004). This environment was likely deposited in an upper ramp setting similar to Lithofacies 2 (Fig. 12). Lithofacies 3 is present throughout the study area as discrete units within deeper marine strata (see Figs. 3, 5, and 6), which suggests periodic shallowing in the basin during portions of the Late Pennsylvanian-Early Permian. Restricted carbonates also occur as thin (< 2 m thick) strata in the lower 60 m of the terrigenous clastic dominated Abo Formation in the Sacramento and San Andres Mountain sections (Figs. 3 and 5). This relationship suggests periodic transgression events, called Hueco tongues, occurred during the early portion of Abo deposition.
Previous studies in the Sacramento Mountains have also described these marine Hueco tongues (transgression events) within the lowermost terrestrial Abo Formation (Jordan, 1975; Raatz, 2002).

**Open Marine Carbonate Facies (Middle-Upper Ramp)**

The presence of calcimicrobial coatings and numerous photozoans (phylloid algae debris, dasycladacean algae, and fusulinids as nuclei) in Lithofacies 4 suggests this environment was well within the photic zone. These oncoids likely formed in a high energy shoal environment, above storm wave base in a middle to upper ramp setting (Wilson, 1975; Flügel, 2004; see Fig. 12). Lithofacies 4 is present throughout the study area, but is most common in the cores (Northwest Shelf) and the Franklin Mountain sections (see Figs. 4, and 6-8). In the Franklin Mountains (Hueco Canyon Formation), Lithofacies 4 is a common constituent in 5-8 m thick shallowing upward cycles, which has been documented by previous studies (Jordan, 1975).

Lithofacies 5 (a-c) is the most common carbonate facies in the study area (see Figs. 3-8). The presence of wackestones-grainstones suggests a wide range of energy conditions in a middle to upper ramp setting (Wilson, 1975; see Fig. 12). Wackestone suggests lower energy conditions, whereas the grainstone units are likely high energy deposits (possibly storms) where the carbonate mud has been winnowed (Flügel, 2004). The high diversity of taxa including: green algae, brachiopods, trilobites, bryozoans, gastropods, and echinoderms suggest deposition in an open marine setting within the photic zone (Wilson, 1975; Flügel, 2004).
Although the environmental interpretation above is consistent with Lithofacies 5 (a-c), a distinction needs to be made between Lithofacies 5a and 5b/5c. Lithofacies 5a (i.e., fusulinid packstone-grainstone) was likely deposited in the photic zone due to the presence of photosynthetic symbionts that likely existed in these large foraminifera (Hallock, 1985). Previous workers have interpreted fusulinid packstones-grainstones as: autochthonous assemblages, downslope accumulations, storm deposits, transgressive lag deposits, or a diagenetic artifact caused by intergranular dissolution during burial (see discussions in Flügel, 2004). Lithofacies 5a is common in the Cerro Alto and Alacran Mountain Formations in the Franklin Mountains (see Figs. 7-8). Because this lithofacies is commonly interbedded with nearshore clastic material (Lithofacies 1d), these fusulinid dominated deposits likely represent transgressive units (of autochthonous assemblages) that are part of high-frequency sea-level cycles. Even though these strata could have potentially formed as storm deposits or downslope accumulations, their cyclicity with nearshore clastic facies (Lithofacies 1d) suggests deposition in the mid-upper ramp environment. The presence of Lithofacies 5a in this study is herein interpreted as a transgressive or deepening facies, compared to the “typical” open marine wackestones-grainstones of Lithofacies 5b and 5c.

**Carbonate Boundstone Facies (Middle-Lower Ramp)**

Due to the high abundance of phylloid algae and encrusting *Tubiphytes* the floatstones/boundstones of Lithofacies 6 (a-b) are interpreted as bioherm deposits. These bioherms occur as erosion-resistant, cliff-forming deposits that are often greater than 10 m thick (Jordan, 1975; Goldstein et al., 1991; Wahlman, 2002; Forsythe, 2003; see Figs.
Previous researchers have determined that these phylloid algae and *Tubiphytes*-dominated deposits likely acted as wave resistant, "reef-like" structures that were common along shelf margins during the Late Pennsylvanian-Early Permian (James, 1983; Goldstein et al., 1991; Wahlman, 2002; Forsythe, 2003). These carbonate mound structures were likely deposited in a middle to lower ramp setting, within the photic zone (euphotic to dysphotic environment) because phylloid algae is thought to have been a type of calcareous green or red algae (Wahlman, 2002; Forsythe, 2003; Flügel, 2004). However, smaller "patch" mounds probably existed within the upper ramp environment (Flügel, 2004; see Fig. 12). Although Lithofacies 6 (a-b) is common throughout the study area, it is especially prevalent in the three cores from the Northwest Shelf (Figs. 4a-c) and the Franklin Mountain sections (Figs. 6-8).

**Offshore Facies (Lower Ramp to Basin)**

Due to a distinct lack of photosynthetic organisms and high abundance of large productid brachiopods and crinoids, Lithofacies 7 is interpreted to record a lower ramp setting. These deposits suggest a deeper depositional setting, compared to the bioherm deposits of Lithofacies 6 (Fig. 12; Flügel, 2004). Lithofacies 7 only occurs in the cores from the Northwest Shelf (not the Orogrande Basin, proper), which supports the notion of a relatively shallow water environment for the Orogrande Basin during the Late Pennsylvanian-Early Permian (Jordan, 1975; Candelaria, 1988; see Fig. 1).

In Lithofacies 8, the lack of major bioturbation, dark color, and lack of fossils likely represents deeper and/or disoxic waters, which is interpreted to have been deposited in a lower ramp setting (Fig. 12). This lithofacies is present in the Tres
Papalotes core, which is interbedded with the relatively deep Lithofacies 7 (Fig. 4b). Lithofacies 8 is also present in the top of the Nicor DDH-2 core (Fig. 4d), which is in the Abo Formation. Although slightly more bioturbation is observed in this unit, as compared with the same unit in the Tres Papalotes core, it is still interpreted as representing a deeper water setting. It is possible that this deep water unit in the Nicor DDH-2 core (lower Abo Formation) is coeval with one of the transgressive Hueco tongues in the lower Abo of the San Andres/Sacramento Mountain sections (Figs. 3 and 5), but this claim is impossible to prove due to a lack of high-resolution age control in the Abo Formation.

SEQUENCE STRATIGRAPHY

The methods of carbonate sequence stratigraphy outlined in Schlager (2005) are utilized in this study. Additionally, sequence stratigraphic concepts related to carbonates outlined in Schlager (1999), Catuneanu (2006), and Catuneanu et al. (2009) have been incorporated into the model for the Orogrande Basin described below. Three trends for carbonate systems deposited in shallow marine settings (platforms/epeiric seas) are summarized: 1.) deepening trends tend to represent the transgressive systems tract (TST); 2.) shallowing upward trends often represent the highstand systems tract (HST); 3.) the lowstand systems tract (LST) is represented by subaerial exposure and/or terrestrial deposits. This model is applied to the facies associations in the studied sections; relative changes in sea level are based on the relative water depth of individual lithofacies. Refer to Figure 12 for the depositional model and relative water depths of individual lithofacies. The sequence stratigraphic model and relative sea-level curve for the carbonate
dominated portion of the Orogrande Basin is summarized on Figures 13 and 14. Below are detailed discussions for the sequence stratigraphic interpretations, separated by chronostratigraphic time interval.

**Gzhelian (Holder Formation; Bursum Formation; Bough Units)**

The interpreted depositional pattern for Upper Pennsylvanian strata in the study area is shallowing upward, high-frequency sequences (4th-5th-order) in a mixed carbonate-clastic system, similar to those of the U.S. Midcontinent (Heckel, 2008). High-frequency sequences are best preserved in the Dry Canyon (Holder Formation) area of the Sacramento Mountains (Fig. 3). A typical high-frequency sequence consists of a basal clastic unit up to 5 m thick (Lithofacies 1a-c), which is overlain by marine carbonate strata up to 5 m thick (Lithofacies 3-6). The basal terrestrial clastic unit likely represents a LST, overlain by open marine carbonates of the HST; a maximum flooding surface (MFS) occurs between these two system tracts. A sequence boundary (SB) separates the marine carbonates from the next overlying sequence. In the Holder Formation 17 of these cycles are documented (see Fig. 3), which is roughly consistent with the 22 high-frequency sequences documented by Rankey et al. (1999) in the same region.

The interpreted sequences most likely reflect high-frequency changes in relative sea level/accommodation, either through glacioeustasy or regional tectonic movements, or some combination of the two. The high abundance of terrigenous clastic sediment in the basin during the Late Pennsylvanian is a result of the increased regional tectonic activity that created numerous uplifted and subaerially exposed areas during this period.
Figure 13: Lithostratigraphic framework for the studied sections. Dashed line represents the Pennsylvanian-Permian boundary, which is the datum for the stratigraphic sections. P.S.: Panther Seep Formation. Dashed line on the inset map represents the line of cross-section from west to east. See Figures 3-8 for additional details regarding individual stratigraphic sections.
Figure 14: Sequence stratigraphic interpretation for the Orogrande Basin. Composite section is based primarily on outcrops in the Franklin and Sacramento Mountains due to the stratigraphic completeness of these successions (see Figs. 3 and 6-8). Relative sea level curve is based on shifts in facies distributions throughout the section. cSB: composite sequence boundary; cMFS: composite maximum flooding surface; cTS: composite transgressive surface; cTST: composite transgressive systems tract; cHST: composite highstand systems tract. Glacial epoch data is from Fielding et al. (2008b; 2008c and references therein). Time scale is based on Gradstein et al. (2004).
of time (e.g., Pedernal Uplift; see Candelaria, 1988; Jordan, 1975). In the southern part of the study area (Franklin Mountains), high-frequency sequences of mixed clastics-carbonates are not present in Lower Permian strata suggesting a decrease of clastics being shed into the basin, which likely resulted from decreased regional tectonic uplift (Candelaria, 1988; Jordan, 1975).

**Asselian to Middle Sakmarian (Hueco Canyon Formation)**

A major sequence boundary separates Pennsylvanian from Permian strata across the Orogrande Basin (e.g., base of the Abo or Hueco Group). This unconformity is also present across much of the southwestern U.S. (see Fig. 14; Ross, 1986; Wahlman and King, 2002). In the Sacramento Mountains section, the base of the thickest Abo conglomerate unit (~9 m thick; Lithofacies 1c) represents the Pennsylvanian-Permian boundary, which suggests a major drop in relative sea level and subsequent erosion of the adjacent Pedernal Uplift (Jordan, 1975; see Fig. 3). A thin pebble conglomerate (~1 m thick; Lithofacies 1c) is present in the lowermost portion of the Abo/Hueco transition zone in the northern San Andres Mountains (Fig. 5), but is stratigraphically higher than the Pennsylvanian-Permian boundary (Giles et al., 2002; see Figs. 5 and 13). This conglomerate (Lithofacies 1c) in the San Andres Mountains likely represents a separate and younger erosion event in the western part of the Orogrande Basin during the earliest Permian. The Pennsylvanian-Permian boundary in the western part of the basin (Robledo Mountains) has been identified by previous workers as an erosional unconformity that separates uppermost Pennsylvanian strata from Lower Permian Hueco units (Wahlman and King, 2002).
The lowermost Permian carbonate strata in the Orogrande Basin (Asselian-middle Sakmarian) are interpreted as a TST (predominantly Lithofacies 6), which overlie the major regional unconformity separating Pennsylvanian from Permian strata (Fig. 14). These open marine bioherm units (Lithofacies 6) represent a deepening compared to underlying clastic dominated uppermost Pennsylvanian units of the Panther Seep/Holder Formations. The bioherm units (Lithofacies 6) do not appear to exhibit obvious cyclicity. However, stable isotopic analysis of the same succession suggests multiple subaerial exposure events (i.e., sequence boundaries) occurred during deposition of the lower Hueco Canyon Formation (Koch, 2010). As such, the overall TST interpreted for this unit likely represents a series of composite sequence sets (see discussions in Coe, 2003), suggesting that this unit is a composite transgressive systems tract (cTST).

The majority of the upper Hueco Canyon Formation consists of multiple shallowing upward cycles (4th-5th-order) that average 5-10 m in thickness. Subaerial exposure events are interpreted for this unit from stable isotope analysis (Koch, 2010). For each cycle, 5-8 m of open marine wackestones-packstones (Lithofacies 5b) are overlain by oncoidal packstones-grainstones (Lithofacies 4; see Fig. 6). These shallowing upward cycles are interpreted as cHST (composite highstand systems tract) deposits. A cMFS (composite maximum flooding surface) is inferred between these two composite system tracts at the point where shoaling upward cycles begin (Fig. 14).

The 4th-5th-order cycles inferred for the Hueco Canyon Formation represent relative sea-level fluctuations that were likely caused by glacioeustatic mechanisms, changes in subsidence rates/accommodation, or a combination of these controls. A composite transgressive surface (cTS) is inferred to be preserved between the Hueco
Canyon and Cerro Alto Formations, which relates to an overall mid-Sakmarian deepening trend. This topic is addressed in the next section below. It should be noted that a composite sequence boundary (cSB) is inferred to exist in the covered section underlying the Cerro Alto Formation (Fig. 14), but lack of available outcrop makes it difficult to identify this surface in the field. Traditional sequence stratigraphic methodology typically invokes a SB between the HST and overlying TST (see Catuneanu, 2006; Catuneanu et al., 2009).

**Middle Sakmarian to Middle Artinskian (Cerro Alto Formation)**

The Cerro Alto Formation consists primarily of a cTST (Lithofacies 5a and 1d) overlain by a cHST (Lithofacies 4 and 5b), separated by a cMFS. This strata stacking pattern is similar to the underlying Hueco Canyon Formation (Fig. 14). However, the basal cTST consists of a deepening trend evident by the appearance of thick fusulinid packstone-grainstone units (see discussions in Flügel, 2004 and Lithofacies 5a above), which are not common in the underlying Hueco Canyon Formation. Depositional cycles within the cTST consist of 1-4.9 m fusulinid packstones-grainstones (Lithofacies 5a) that are overlain by relatively thin (less than 1 m) siltstone beds of Lithofacies 1d (Fig. 7). These depositional cycles within the cTST suggest high order fluctuations in relative sea level as clastics were brought into the basin during relative lowstands of sea level. Overlying each cTST are strata that contain shallowing upward cycles (Lithofacies 4 and 5b), which are interpreted as representing cHST deposits (Figs. 7 and 14).

The inferred cTS at the base of the Cerro Alto Formation separates cHST from cTST deposits. Although not identifiable in the field, a cSB likely exists in the covered
section below the cTS (Fig. 14). If a cSB does not exist, this cTS would be labeled a “drowning unconformity” or “Type 3 sequence boundary” after various authors (e.g., Schlager, 1999; Catuneanu, 2006). Regardless of the name specified and presence (or absence) of a sequence boundary, there is a distinct deepening trend inferred, which suggests an increase in relative sea level in the basin during the mid-Sakmarian.

Another cSB is inferred near the top of the Cerro Alto Formation (Fig. 14), but is unidentifiable in the field. This cSB likely exists in the covered strata near the top of the formation, and separates cHST deposits (Lithofacies 4 and 5b) from cTST deposits of Lithofacies 5 and 6.

**Late Artinskian (Alacran Mountain Formation)**

The Alacran Mountain Formation is similar to the Cerro Alto Formation in that it consists of a basal cTST (Lithofacies 5-6 and Lithofacies 1d). The cTST is interpreted as an overall deepening trend, compared to the upper portions of the underlying Cerro Alto Formation. The cTST is overlain by phylloid algae bioherm deposits (cHST) of predominantly Lithofacies 6, with a cMFS inferred between these composite systems tracts. The cHST is dominated by shallowing upward cycles consisting of large phylloid algae bioherms (Lithofacies 6), overlain by open marine wackestones-grainstones and nearshore clastics of Lithofacies 1d, 4, and 5b (Figs. 8 and 14).

The phylloid algae bioherm deposits (Lithofacies 6) are thicker (> 10 m) and more common in the Alacran Mountain Formation compared to the underlying Cerro Alto Formation, which suggests more stability in overall environmental conditions. However, the presence of thin beds (< 0.10 m) of clastic siltstone (Lithofacies 1d)
throughout the unit suggest continued high-frequency fluctuations in relative sea level
during Alacran Mountain deposition at the end of Artinskian time.

Upper Artinskian Hueco/Abo units are overlain by the terrestrial/restricted
carbonate-dominated Yeso Formation (Jordan, 1975; Candelaria, 1988; Raatz, 2002).
This significant basinward shift in facies throughout the basin at the end of the Artinskian
suggests a lowering of relative sea level, and subsequent formation of a SB.

SYNTHESIS AND DISCUSSION

The paleotropics represent an ideal region in which to study glacioeustasy
because glacioisostatic effects are not present in these systems. The studied interval for
this report (Gzhelian-Artinskian) overlaps with the timing of the P1 and P2 glacial epochs
from eastern Australia (Fielding et al., 2008a; see Fig. 14), as well as glacial epochs in
Siberia and other Gondwanan localities including Antarctica, South America, southern
Africa, the Middle East, India, and Western Australia (Fielding et al., 2008b-c).

The documented SB separating Pennsylvanian and Permian strata in the
Orogrande Basin is coeval with the beginning of the P1 glaciation in eastern Australia, as
well as other glacial epochs from across Gondwana (Fig. 14). The formation of this SB
at the Pennsylvanian-Permian boundary likely reflects a eustatic drop coinciding with the
buildup of significant glacial ice across Gondwana (Fielding et al., 2008c). Recent
research suggests that the Pennsylvanian-Permian sequence boundary can be traced
across the majority of Pangean paleotropical carbonate platforms, which further supports
the link between this eustatic drop and an increase in Gondwanan glacial ice volume
(Koch, 2010). Because tectonic uplift was occurring in and around the Orogrande Basin
during the Late Pennsylvanian-Early Permian (Jordan, 1975; Candelaria, 1988; Raatz, 2002), it is possible that the formation of this sequence boundary was enhanced by regional tectonic uplift.

The record from Siberia is more cryptic and does not seem to fit into any of the above interpretations; it lay between the P1 and P2 glacial epochs from eastern Australia. It should be noted that the record from Siberia is poorly dated compared to other localities (Raymond and Metz, 2004). It is possible that the Siberian record may actually correlate with either the P1 or P2 glacial epochs in eastern Australia, or that the volume of glacial ice in Siberia was too small to create a global imprint.

Two transgressive events are interpreted to have occurred in the mid-Sakmarian and mid-Artinskian, which correspond roughly with the end of both the P1 and P2 glacial epochs in eastern Australia. These events are represented by distinct deepening trends within the facies stacking patterns (Fig. 14). The controlling mechanism for these transgression events was either a dramatic increase in regional subsidence or a rapid eustatic rise. Because previous Orogrande Basin studies have suggested that a decrease in subsidence occurred during the Wolfcampian, which is based on depositional trends and facies associations (Jordan, 1975; Candelaria, 1988; Raatz, 2002), it is likely that rapid eustatic rise controlled the inferred increase in overall water depth during the mid-Sakmarian and mid-Artinskian (Fig. 14).

The mid-Sakmarian transgression event represents the beginning of a significant change in depositional facies in the basin due to the introduction of terrigenous clastics (Lithofacies 1d), interbedded with fusulinid packstones-grainstones (Lithofacies 5a). These stacking patterns are interpreted above as high-frequency (i.e., 4th and 5th-order)
cycles of relative sea-level change. It is important to note that other mechanisms besides sea-level change can bring clastics into the center of a carbonate basin. For example, changes in large-scale climatic regimes such as tropical cyclones, increased seasonality, etc. during the Early Permian could have triggered seasonal storm events that brought these clastics into the southern part of the basin, which would not require a change in relative sea level. Regional paleoenvironmental studies for the southwestern U.S. suggest that the Early Permian was warmer and more arid than the Late Pennsylvanian, but seasonal tropical cyclones were likely present (Tabor and Montañez, 2002; Tabor, 2007; Tabor et al., 2008). Given the large-scale glacial epochs in Gondwana during the Early Permian (e.g., P1 and P2 in eastern Australia; see Fielding et al., 2008a), it is interpreted that these mid-Sakmarian-Artinskian terrigenous clastics (Lithofacies 1d) were deposited as part of high-frequency cycles of relative sea-level change, which were potentially influenced by large and infrequent storm events. It is likely that a change in large-scale atmospheric circulation patterns beginning in the mid-Sakmarian could be related to changes in overall climatic regimes, coinciding with the end of major Gondwanan glaciation (Fielding et al., 2008b-c).

**CONCLUSIONS**

The uppermost Pennsylvanian-Lower Permian strata of the Orogrande Basin are divided into five facies associations that range from terrigenous clastic deposits to deeper, quieter carbonate ramp environments that may have lain below the photic zone. Because the southern (i.e., deepest) part of the basin contained primarily shallow water carbonate deposition and lacked deep basinal facies, the Orogrande Basin was a shallow basin
during the latest Pennsylvanian-Early Permian (Jordan, 1975). As such, the Orogrande Basin is interpreted as being deposited in a carbonate ramp environment with phylloid algae/Tubiphytes bioherms acting as "reef-like" structures, which is consistent with other paleotropical localities during the late Paleozoic (James, 1983; Wahlman, 2002).

Sequence stratigraphic interpretations suggest relative sea-level changes were largely controlled by glacioeustatic mechanisms. The major sequence boundary separating Pennsylvanian-Permian strata in the Orogrande Basin likely formed during a eustatic drop associated with the expansion of glacial ice centers across Gondwana during the earliest Permian (Fielding et al. 2008b). Two transgression events (mid-Sakmarian and mid-Artinskian) likely represent eustatic rise resulting from deglaciation events (P1 and P2 glaciations) across eastern Australia and Gondwana.

Upper Paleozoic deposits from south-central New Mexico provide an excellent opportunity to explore whether a glacioeustatic signal is clearly preserved, and to what extent. The results from this study suggest that large-scale glacial epochs and deglaciation events from Gondwana correlate with stratigraphic changes in the Orogrande Basin (paleotropics) for the Early Permian portion of the LPIA. However, global correlations of high-frequency cycles are difficult given the current chronostratigraphic control for upper Paleozoic strata. Yet, 4th and 5th-order cycles are clearly documented in many locations across Euramerica and Gondwana (e.g., Rankey et al., 1999; Mack et al., 2003; Heckel, 2008; Birgenheier et al., 2009). Future research will focus on the construction of more highly resolved records from the Orogrande Basin, as well as other paleotropical localities, which will ultimately lead to more refined global correlations of upper Paleozoic strata.
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CHAPTER 3:

Carbon isotope stratigraphy of uppermost Pennsylvanian-Lower Permian marine carbonates in south-central New Mexico: Implications for glaciation, sea level, and depositional patterns

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ABSTRACT

Although the Permian was a time of climate transition from deep icehouse to greenhouse conditions, few isotopic records exist that document these changes. The goal of this investigation is to document $\delta^{13}C$ trends from uppermost Pennsylvanian-Lower Permian carbonate strata in the Orogrande Basin in order to assess the global significance of newly resolved stratigraphic records of Gondwanan glacial epochs. Periods of subaerial exposure from the Gzhelian through the early Sakmarian are associated with $\delta^{13}C$ values as low as -5.9‰ and exposure surfaces identified in sequence stratigraphy. During the early Sakmarian-Kungurian, a time of deglaciation across Gondwana, $\delta^{13}C$ values trend toward higher values (+2‰ to +5‰). Two small excursions of decreasing $\delta^{13}C$ values are inferred for upper Sakmarian-Kungurian strata, each represented by a 1.5‰ decrease in $\delta^{13}C$ values. These lower $\delta^{13}C$ values are associated with facies changes to shallower restricted conditions that correspond to periods of expansion of glacial ice across much of Gondwana. In addition to adding to the database of Permian isotopic records, results from this study suggest that marine chemistry, depositional environments, and sea level in the Orogrande Basin were profoundly influenced by epochs of Gondwanan glaciation. These inferences help improve our understanding of how far-field carbonate systems responded to the effects of Gondwanan glaciation and climate change during the acme and waning stages of the late Paleozoic ice age.
INTRODUCTION

Stable isotopic records have provided much insight into the late Paleozoic ice age (LPIA), including the onset, duration, timing, and demise (e.g., Saltzman, 2003; Montañéz et al., 2007; Frank et al., 2008; Grossman et al., 2008; Elrick and Scott, 2010). Much of what is known about late Paleozoic climate has come largely from Euramerican cyclothem and stable isotope records due to a lack of high-resolution stratigraphic and isotopic data from Gondwana (e.g., Bruckschen et al., 1999; Mii et al., 1999, Veizer et al., 1999; Mii et al., 2001; Wright and Vanstone, 2001; Saltzman, 2003; Korte et al., 2005; Heckel, 2008).

Previous studies focused largely on the Carboniferous and were influenced by the view that the most intense glaciation occurred during that time (e.g., Veevers and Powell, 1987; Gonzalez-Bonorino and Eyles, 1995). However, recent stratigraphic studies from Gondwana (Isbell et al., 2003; Fielding et al., 2008a-c) and global stable isotopic studies (Montañéz et al., 2007; Frank et al., 2008; Grossman et al., 2008) have encouraged a re-examination of climate change during the late Paleozoic. This reassessment revealed that isotopic trends, in part, correspond to stratigraphic changes in Gondwanan records. Frank et al. (2008) compiled published records of well-preserved brachiopods and demonstrated that the largest mismatch between Euramerican isotopic trends and Gondwanan stratigraphic records was for Carboniferous data sets; Permian records matched relatively well. In addition, Frank et al. (2008) inferred that the complicated Carboniferous patterns resulted from the presence of numerous small ice centers that were active at different times, thus creating a disorganized global response. Permian isotope records match Gondwanan stratigraphic records better because glaciation became
more synchronous and widespread during the Early Permian (Montañez et al., 2007; Frank et al., 2008).

However, a conspicuous issue evident in previous late Paleozoic isotopic studies is that the Permian record relies on relatively small data sets. It is, therefore, difficult to decipher climate change during the latter part of the LPIA (Frank et al., 2008). The main goal of this study is to generate a high-resolution δ¹³C record for a single paleotropical basin, and to investigate the possible far-field response to glacial epochs in Gondwana during the Early Permian. For this paper, δ¹³C data from uppermost Pennsylvanian-Lower Permian strata in the Orogrande Basin in south-central New Mexico are examined (Fig. 1). This basin contains some of the most continuous marine carbonate sections of Upper Pennsylvanian-Lower Permian strata in the southwestern United States (Jordan, 1975; Raatz, 2002; Wahlman and King, 2002). High-resolution stable isotope records from many Lower Permian successions around the globe (such as the Orogrande Basin) are needed to better decipher climate change during the acme and subsequent demise of the LPIA.

**GEOLOGIC SETTING**

**The Orogrande Basin**

The Orogrande Basin of south-central New Mexico was a shallow and elongate basin that formed during the Middle Mississippian (Fig. 1; Seager et al., 1976; Candelaria, 1988; Raatz, 2002). The basin lay within 5° of the equator during the Late Pennsylvanian-Early Permian (Ross and Ross, 1990; Scotese and Langford, 1995; Golonka and Ford, 2000). Regional studies suggest that the Orogrande Basin was, at
Figure 1: Regional map of the Orogrande Basin; modified after Candelaria (1988). Locations of outcrops examined in this study are denoted with black dots and core locations are denoted with gray dots. Exact location for each section is on caption for Figure 3.
times, connected to the Permian Basin to the east (i.e., Northwest Shelf/Delaware Basin) and the Pedregosa Basin to the south (Jordan, 1975; Candelaria, 1988; Raatz, 2002). Regional tectonic activity during the Pennsylvanian created uplifted zones adjacent to the basin (Pedernal, Diablo, Florida uplifts), which were associated with the Ouachita-Marathon orogenic belt and the final assembly of Pangea (Fig. 1; Candelaria, 1988; Raatz, 2002). Zones of uplift attained maximum relief during the latest Pennsylvanian-earliest Permian, resulting in significant volumes of clastic sediment being deposited around the margins of the basin (Jordan, 1975; Candelaria, 1988). By the end of the Wolfcampian, accommodation in the basin was limited due to terrigenous clastic/marine carbonate infill and a reduction in subsidence (Jordan, 1975; Candelaria, 1988; Raatz, 2002). The region returned to an evaporite/shallow marine environment during the Leonardian and became connected to the Delaware Basin/Northwest Shelf to the east (Candelaria, 1988; Raatz, 2002).

**Stratigraphy**

Continuous sections of uppermost Pennsylvanian-Lower Permian marine carbonates can be found in the Orogrande Basin (Jordan, 1975; Wahlman and King, 2002). Deposits range from open marine tropical carbonate units in the south to mixed siliciclastic-carbonate strata in the north (Jordan, 1975; Candelaria, 1988; Rankey et al., 1999). Lower Permian strata unconformably overlie Pennsylvanian units (Schoderbek and Chafetz, 1988; Wahlman and King, 2002); this unconformity is present across much of the southwestern United States (Ross, 1986).
The Pennsylvanian-Permian boundary discussed herein is based on the Global Stratotype Section and Point (GSSP) for the base of the Permian System, which has been placed in Kazakhstan (Davydov et al., 1998). Conodont and fusulinid studies have correlated the Pennsylvanian-Permian boundary in Kazakhstan with stratigraphic sections throughout North America (Ross, 1963; Ritter, 1995; Wahlman and King, 2002). This boundary shift reassigns lower Wolfcampian strata to the uppermost Pennsylvanian. Stratigraphic units such as the Bursum Formation and Bough members, which have traditionally been viewed as Permian, are now assigned to the uppermost Pennsylvanian (Wahlman and King, 2002).

Upper Pennsylvanian carbonate strata analyzed for this report are mixed siliciclastic-marine carbonate dominated strata of the Holder Formation, uppermost Beeman Formation, informal Bough members (including the Saunders unit), uppermost Panther Seep Formation, and the Bursum Formation. Lower Permian strata examined are the Hueco Group and lower San Andres Formation. Refer to Figure 2 for a chronostratigraphic framework of these units.

*Uppermost Pennsylvanian strata*

Upper Pennsylvanian strata are divided into the Virgilian age Beeman Formation, the Upper Virgilian Holder Formation (Panther Seep-equivalent), and the uppermost Virgilian Bursum (Laborcita) Formation (Fig. 2). The Holder and Beeman Formations are best exposed in the Sacramento Mountains where they consists of alternating cyclothems of open marine limestone and terrigenous clastic units, and large phylloid algae bioherms (Wilson, 1967; Rankey et al., 1999). Strata of the Holder Formation are
Figure 2: Chronostratigraphic framework for the study area based on data presented in Jordan (1975); Cys and Mazzullo (1985); Malek-Aslani (1985); Candelaria (1988); Rankey et al. (1999); Raatz (2002); Mack et al. (2003). Time scale is based on Gradstein et al. (2004). Conodont and global fusulinid zones after Gradstein et al. (2004; and references therein); U.S. Midcontinent fusulinid zones after Ross and Ross (1988; 1995).
the shallow, updip equivalent to the slightly deeper water facies of the Panther Seep Formation (Schoderbek and Chafetz, 1988; Rankey et al., 1999). Biostratigraphic data utilizing fusulinids suggests a Virgilian age for the Holder and Beeman Formations (Cline, 1959; Schoderbek and Chafetz, 1988; Rankey et al., 1999).

Open marine limestone and terrigenous clastic units of the Bursum Formation are discontinuously preserved throughout the basin (Jordan, 1975; Candelaria, 1988; Raatz, 2002). Fusulinid studies assign Bursum strata (and equivalent units) to the uppermost Pennsylvanian (Steiner and Williams, 1968; Wahlman and King, 2002).

A portion of the Delaware Basin (i.e., Northwest Shelf) extends into southeastern New Mexico, east of the Sacramento Mountains (Figs. 1 and 2). Informal Bough units (A-D) and a Saunders member have been described as part of the Lower Permian Hueco Limestone (Cys and Mazzullo, 1985; Malek-Aslani, 1985). However, redefinition of the Pennsylvanian-Permian boundary (Davydov et al., 1998) results in the reassignment of these units to the latest Pennsylvanian. The Bough members consist primarily of open marine carbonates including, phylloid algae bioherms and open marine storm dominated shelf deposits (Cys and Mazzullo, 1985; Malek-Aslani, 1985).

Lower Permian strata

Lower Permian carbonate strata in the southern part of the Orogrande Basin primarily consist of open to restricted marine units of the Hueco Group (Fig. 2; Jordan, 1975; Seager et al., 1976; Mack and James, 1986; Candelaria, 1988; Mack et al., 1988; Wilson and Jordan, 1988; Raatz, 2002; Wahlman and King, 2002; Mack et al., 2003; Mack, 2007). Coeval strata in the northern part of the basin consist of the terrigenous
clastic dominated Abo Formation. A transition zone occurs between Hueco and Abo strata in the central portion of the basin where the units interfinger (Robledo Mountains). In this transitional zone marine carbonates of the Hueco Group show characteristics consistent with a more restricted and shallow environment; the clastic units of the Abo Formation contain marginal marine facies, including fluvio-estuarine deposits (Jordan, 1975; Wahlman and King 2002; Mack et al., 2003). Fusulinid data suggest Hueco Group strata were deposited during the Early Permian (Williams, 1963; 1966; Simo et al., 2000; Wahlman and King, 2002; Krainer et al., 2005).

In the central to northern part of the basin, the Lower Permian terrigenous to marginal marine Abo Formation is overlain by terrigenous clastics/restricted carbonates of the Yeso Formation (Kottlowski, 1975; Candelaria, 1988; Lindsay and Reed, 1992; Mack et al., 2003). During the Leonardian, the region returned to a marine environment dominated by shallow water carbonate deposition of the San Andres Formation (Fig. 2; Kottlowski, 1975; Milner, 1976; Lindsay and Reed, 1992). The San Andres Formation consists of variably dolomitized carbonate mudstones-grainstones, often containing gypsum nodules (Milner, 1976; Lindsay and Reed, 1992). These carbonate units are thought to have been deposited in a restricted to open marine, storm dominated environment (Kottlowski, 1975; Milner, 1976; Lindsay and Reed, 1992; Raatz, 2002).

SAMPLES AND ANALYTICAL METHODS

This study uses samples collected from cores and outcrops throughout the Orogrande Basin (see Figs. 1 and 3). A total of 286 samples from multiple facies (i.e., Facies Associations A-C; see discussions below) were analyzed using standard
Figure 3: Plot of δ¹³C values through each stratigraphic section, from west to east. Section locations (base of section): Florida Shelf (Nicor DDH-2 Core)-19S, 7W, Section 29; Franklin Mountains section 31°55.26'N-106°30.85'W; L. San Andres Mountains section 33°11.63'N-106°41.72'W; U. San Andres Mountains section 33°18.02'N-106°43.52'W; L. Sacramento Mountains section 32°57.20'N-105°54.03'W; U. Sacramento Mountains 32°58.46'W-105°45.46'W. Northwest Shelf Cores (shown as a composite section on figure): Unocal State No. 1-33 Core-15S, 32E, Section 33; Tres Papalotes Core-14S, 34E, Section 28; Getty No. 5 Williard Beatty Core-13S, 33E, Section 35. Time scale is based on Gradstein et al. (2004). PS: Panther Seep Formation.
petrographic methods, and both matrix and brachiopod shell material was utilized for δ¹³C and δ¹⁸O analysis (see Table 1). All isotopic analyses were carried out at the Keck Paleoenvironmental & Environmental Stable Isotope Laboratory (KPESIL) at the University of Kansas using Kiel Carbonate Device III + Finnigan MAT253 isotope ratio mass spectrometer (ThermoFinnigan, Germany). Age determinations for the data set are based on published biostratigraphic frameworks for the Orogrande Basin (Williams, 1966; Jordan, 1975; Simo et al., 2000; Wahlman and King, 2002; Krainer et al., 2005), calibrated to the Gradstein et al. (2004) timescale.

Microsamples (20 to 80 μg) for isotopic analysis were drilled from polished slabs using a microscope-mounted drilling assembly and dental drill bits with 20-500 μm tip diameters. Microsamples were roasted under vacuum at 200°C for 1 h to release any volatile organic compounds. Calcite microsamples and standards were reacted at 75°C under vacuum using phosphoric acid. Phosphoric acid was prepared according to the Stanford University Stable Isotope Laboratory's On-line manual (http://pangea.stanford.edu/research/isotope/dam/manual.html). Results are reported in the standard δ notation as δ¹³C (or δ¹⁸O) relative to VPDB. NBS-18 Carbonatite (NIST Ref. Mat. 8543) & NBS-19 Limestone (NIST Ref. Mat. 8544) were used as quality control standards; analytical precision is better than 0.01‰ for δ¹³C and better than 0.02‰ for δ¹⁸O.

FACIES AND DEPOSITIONAL TRENDS

Facies disposition has been used to infer that uppermost Pennsylvanian-Lower Permian carbonates in the Orogrande Basin were deposited on a homoclinal ramp (Koch,
Table 1. Raw stable isotopic data from measured sections and cores used in this study.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Section</th>
<th>Material</th>
<th>Meters Above Section Base</th>
<th>δ¹³C (‰,VPDB)</th>
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Table 1 (continued). Raw stable isotopic data from measured sections and cores used in this study.

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L. San Andres Mtns.

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Table 1 (continued). Raw stable isotopic data from measured sections and cores used in this study.

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U. Sacramento Mts. | SN-B-2          | San Andés Formation  | mérinate    | 15              | 1.5             | -4.6          |
|        | SN-B-9          | San Andés Formation  | mérinate    | 22              | 3.0             | -4.5          |
|        | SN-C-0          | San Andés Formation  | mérinate    | 46.3            | 3.9             | -2.2          |
|        | SN-C-3          | San Andés Formation  | mérinate    | 49.5            | 4.3             | -2.9          |
|        | SN-D-10         | San Andés Formation  | mérinate    | 61              | 3.5             | -5.7          |
|        | SN-D-15         | San Andés Formation  | mérinate    | 69              | 3.3             | -4.6          |
|        | SN-D-22         | San Andés Formation  | mérinate    | 74              | 4.1             | -3.1          |
|        | SN-D-3          | San Andés Formation  | mérinate    | 84              | 2.5             | -4.6          |
|        | SN-E-0          | San Andés Formation  | mérinate    | 134             | 3.3             | -4.9          |
|        | SN-F-5          | San Andés Formation  | mérinate    | 139             | 3.1             | -2.8          |
|        | SN-F-0          | San Andés Formation  | mérinate    | 153.5           | 4.1             | -2.7          |
|        | SN-F-8          | San Andés Formation  | mérinate    | 161             | 3.3             | -4.2          |
|        | SN-G-1          | San Andés Formation  | mérinate    | 185.5           | 3.9             | -3.3          |

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| 3     | Getty No. 5 Willard Beauty Core | mérinate | 4.4             | -0.3            | -5.3          |
| 4     | Getty No. 5 Willard Beauty Core | mérinate | 5.0             | 0.3             | -5.8          |
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| 6A    | Getty No. 5 Willard Beauty Core | mérinate | 8.8             | 1.0             | -5.4          |
| 6B    | Getty No. 5 Willard Beauty Core | fusa/fuid | 8.8             | 1.0             | -5.4          |
| 7A    | Getty No. 5 Willard Beauty Core | mérinate | 9.1             | 0.5             | -5.1          |
| 7B    | Getty No. 5 Willard Beauty Core | brachiopod | 9.1             | 0.5             | -4.8          |
| 8A    | Getty No. 5 Willard Beauty Core | mérinate | 9.7             | 0.1             | -5.0          |
| 8B    | Getty No. 5 Willard Beauty Core | brachiopod | 9.7             | 0.5             | -5.3          |
| 9     | Getty No. 5 Willard Beauty Core | mérinate | 11.8            | -0.2            | -4.8          |
Table 1 (continued). Raw stable isotopic data from measured sections and cores used in this study.

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Table 1 (continued). Raw stable isotopic data from measured sections and cores used in this study.

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Like other upper Paleozoic tropical carbonate systems (e.g., James, 1983; Wahlman, 2002) these deposits lack evidence of rimmed platforms and reef framework builders during this time period. Instead, relatively shallow-water phylloid algae and Tubiphytes-rich bioherm structures dominated (Wahlman, 2002; Forsythe, 2003), which is consistent with the depositional style in the Orogrande Basin (Jordan, 1975; Candelaria, 1988; Koch, 2010). For the purposes of this paper, carbonate facies have been divided into three facies associations: A) mudstone and wackestone, deposited in a restricted to intertidal environment; B) wackestone-grainstone, deposited in a shallow subtidal environment; C) boundstone/floatstone, deposited as phylloid algae-Tubiphytes dominated bioherms. A more detailed facies analysis is provided in Koch (2010). Refer to Figure 3 for additional details regarding specific facies associations and depositional trends in the Orogrande Basin.

**Facies Association A: mudstone and wackestone (restricted to intertidal environment)**

Carbonate mudstones and wackestones interpreted as deposits of a restricted to intertidal environment occur in sections in the Franklin, San Andres, and Sacramento Mountains (Figs. 1 and 3). Microbially laminated facies, presence of fenestral fabric, and a lack of open marine fauna are consistent with deposition in a very shallow, restricted environment in an upper ramp setting (Wilson, 1975; Flügel, 2004). This facies association also includes strata inferred to have been deposited in a restricted/hypersaline lagoon environment that was proximal to a clastic source, which is based on the presence
of terrigenous clastics and low diversity/high abundance organisms such as gastropods and ostracods typical of stressed environmental conditions (Flügel, 2004; Koch, 2010).

Facies Association A is most common in Upper Pennsylvanian and lowermost Permian strata of the Beeman, Holder, Panther Seep, lower Hueco, Hueco Canyon Formations (i.e., near the Pennsylvanian-Permian boundary; see Fig. 3). This facies is not present in the Cerro Alto-Alacran Mountain Formations in the Franklin Mountains, which suggests most carbonate deposition remained in the subtidal environment in the southern part of the basin during the mid-Sakmarian through Artinskian (Fig. 3). Facies Association A is also present in the lower San Andres Formation, suggesting a return to a very shallow water environment during the early Kungurian.

**Facies Association B: wackestone-grainstone (shallow subtidal environment)**

Strata interpreted as being deposited in a shallow subtidal environment consist of bioturbated photozoan-rich wackestone to grainstone. Fossil content is abundant with a diverse and open marine fauna including: phylloid algae, brachiopods, corals, foraminifera, fusulinids, gastropods, ostracods, echinoderms, bryozoans, and trilobites. The presence of wackestones-grainstones suggests a wide range of energy conditions in a middle to upper ramp setting (Wilson, 1975). Wackestone suggests lower energy conditions, whereas the grainstone units are likely high energy mud-winnowed deposits. Also abundant are oncoids, which suggest a high energy carbonate shoal environment (Flügel, 2004).

Although Facies Association B is widespread throughout the study area, it is most common in the Cerro Alto Formation and parts of the Alacran Mountain Formation in the
Franklin Mountains (Figs. 1 and 3). This suggests that the majority of carbonate deposition in the southern part of the Orogrande Basin was predominantly in an open marine subtidal environment during the mid-Sakmarian through the Artinskian. In the San Andres Formation, lowermost strata consist of Facies Association A (i.e., extremely shallow-water), but then transition into subtidal facies (Facies Association B) higher up in the formation (Fig. 3).

**Facies Association C: boundstone/floatstone (phyllloid algae-\textit{Tubiphytes} bioherms)**

Phyllloid algae-\textit{Tubiphytes}-rich floatstone/boundstone is interpreted as bioherm deposits. Previous researchers have concluded that these phyllloid algae and \textit{Tubiphytes} dominated deposits likely acted as wave resistant, "reef-like" structures that were common along shelf margins during the Late Pennsylvanian-Early Permian (James, 1983; Wahlman, 2002; Forsythe, 2003). These carbonate bioherms were likely deposited in a middle to lower ramp setting, but within the photic zone (i.e., euphotic to dysphotic environment) because phyllloid algae is thought to have been a type of calcareous green or red algae (Wahlman, 2002; Forsythe, 2003; Flügel, 2004).

Facies Association C is most common in Lower Permian strata in the Franklin Mountains and in the Bough Units from the Northwest Shelf (Figs. 1 and 3). In the Franklin Mountains, thick bioherm deposits are dominant in the lower Hueco Canyon and Alacran Mountain Formations. Strata of the Cerro Alto Formation (i.e., mid-Sakmarian-lower Artinskian) are dominated by subtidal units of Facies Association B, which suggest slightly shallower conditions, compared to the overlying Alacran Mountain Formation (Fig. 3).
RESULTS

Overview

Of the 286 samples analyzed for this study, 220 are from matrix material, 39 are from brachiopod shell material, and 27 are from other materials (e.g., crinoids, spar fills, mollusk shell material, etc.). For the entire data set, matrix $\delta^{13}C$ values range from -5.9‰ to +5.4‰; brachiopod $\delta^{13}C$ values range from -5.3‰ to +3.8‰; other material $\delta^{13}C$ values range from -4.6‰ to +4.7‰ (see Fig. 3-4; Table 1). The $\delta^{18}O$ values for matrix samples range from -9.5‰ to -0.8‰; brachiopod $\delta^{18}O$ values range from -9.2‰ to -2.5‰; $\delta^{18}O$ values from other material ranges between -13.0‰ to -1.9‰.

Stratigraphic trends

In the Nicor DDH-2 core (Florida Shelf), a clear stratigraphic trend up-section is not evident, but $\delta^{13}C$ values range from -5.6‰ to -0.5‰ (Table 1; Fig. 3). The Franklin Mountains section is characterized by more variable $\delta^{13}C$ values -1.6‰ to +4.7‰ in the lower part of the section (Hueco Canyon Formation), and more constant $\delta^{13}C$ values of +1.4‰ to +5.4‰ in the upper part of the section (Cerro Alto and Alacran Mountain Formations). A small trend of decreasing $\delta^{13}C$ values is also present in the upper part of the Franklin Mountains section (Excursion 1; see Fig. 3), which coincides with a shift to shallower-water facies (Facies Association B). The San Andres Mountains section is characterized by variable $\delta^{13}C$ values (-5.0‰ to +3.1‰) in the lower part of the section (Hueco/Abo Formations); $\delta^{13}C$ values in the upper part of the San Andres Mountains section (San Andres Formation) are much more consistent, averaging approximately +3‰ (Fig. 3). This trend is consistent with the San Andres Formation in the Sacramento
Figure 4: Carbon vs. oxygen plots for $\delta^{13}C$ and $\delta^{18}O$ values. A.) carbon vs. oxygen plotted by stage; B.) carbon vs. oxygen plotted by measured section or location. Values for the $\delta^{18}O$ value of Permian marine calcite are -2.5‰ to -2.8‰ (Given and Lohmann, 1985) and -0.7‰ to -3.3‰ (Korte et al., 2005); values for the $\delta^{13}C$ of Permian marine calcite are +5.2‰ to +5.8‰ (Given and Lohmann, 1985) and +4.0‰ (Korte et al., 2005).
Mountains (Table 1; Fig. 3). The lower part of the San Andres Formation contains a small trend of decreasing $\delta^{13}C$ values (as low as -5.8‰) that is labeled Excursion 2, which is associated with fairly shallow-water facies (Facies Association A; Fig. 3). The lower part of the Sacramento Mountains Section (Beeman, Holder, and Hueco/Abo Formations) consists of variable $\delta^{13}C$ values (-5.9‰ to +3.1‰), without evidence of a clear stratigraphic trend (Fig. 3). In the cores from the Northwest Shelf, $\delta^{13}C$ values are variable and range from -4.8‰ to +4.6‰.

The combination of data into a single composite section using previously published age constraints for each stratigraphic unit (Cline, 1959; Williams, 1963; 1966; Schoderbek and Chafetz, 1988; Rankey et al., 1999; Simo et al., 2000; Wahlman and King, 2002; Krainer et al., 2005) shows a clear stratigraphic trend (Fig. 5). There is considerably more scatter in Gzhelian-Asselian strata relative to the rest of the succession. With the exception of one outlier ($\delta^{13}C$ value of -5.8‰ in the upper San Andres Mountains section), the $\delta^{13}C$ values from the lower Sakmarian through the Kungurian range from +1.3‰ to +5.4‰. The $\delta^{13}C$ values from the Gzhelian through the early Sakmarian are generally lower and exhibit much more variation, ranging from -5.9‰ to +4.6‰. The overall trend of lower and more variable $\delta^{13}C$ values in Gzhelian-lower Sakmarian strata is consistent across the study area regardless of facies. Lower $\delta^{13}C$ values occur in intertidal to restricted units (Facies Association A), as well as subtidal units of Facies Associations B and C (Fig. 3). However, $\delta^{13}C$ values in Facies Association A tend to be 1‰-3‰ lower than coeval subtidal units in Facies Association B and C. For example, $\delta^{13}C$ values in the shallow subtidal to bioherm deposits (Facies Associations B and C) of the lower Hueco Canyon Formation (Franklin Mountains)
Figure 5: Composite record of $\delta^{13}$C values for the Orogrande Basin compiled from sections shown on Figure 3. Time scale is based on Gradstein et al. (2004); Distribution of Gondwanan glacial epochs from Fielding et al. (2008c and references therein); atmospheric pCO2 data from Ekart et al. (1999) and Montañez et al. (2007); bulk Corg data from Birgenheier (2010); sea level data from Rygel et al. (2008).
average between +1‰ and +3‰. By contrast, coeval δ^{13}C values in the predominantly restricted units (Facies Association A) in the Hueco Formation (lower San Andres Mountains) average between -2‰ and 0‰ (Fig. 3).

Two small transient trends toward lower δ^{13}C values (i.e., Excursions 1 and 2) also occur between the lower Sakmarian-middle Artinskian and in the lower Kungurian, each representing a 1.5‰ decrease in δ^{13}C values (see Fig. 5). On Figure 5, a panel is presented that shows only the δ^{13}C values between 0‰ and +6‰, which emphasizes these smaller excursions. The δ^{13}C values during Excursions 1 and 2 average +2.5‰, which is approximately 1.5‰ lower than δ^{13}C values in middle Sakmarian strata (approximately +4.0‰).

Excursion 1 (upper Sakmarian through middle Artinskian) occurs in the middle to upper Cerro Alto Formation, which is dominated by shallow subtidal strata of Facies Association B (Fig. 3). The beginning of this excursion corresponds to a shallowing of facies. More specifically, shallow subtidal to bioherm deposits (Facies Associations B and C) in the lower Cerro Alto Formation transition into predominantly shallow subtidal deposits (Facies Association B) in the middle Cerro Alto Formation. The end of the excursion corresponds to a facies change from shallow subtidal (Facies Association B) to slightly deeper units of Facies Associations B and C (Fig. 3 and 5).

A similar shallowing of facies is associated with the negative δ^{13}C excursion in the Kungurian (i.e., Excursion 2). In the lower San Andres Formation, intertidal to restricted units (Facies Association A) correspond to lower δ^{13}C values (+1.5‰ to +2.5‰). Overlying the intertidal to restricted deposits are shallow subtidal deposits of
Facies Association B (Figs. 3 and 5). The shift to deeper water conditions in the middle San Andres Formation corresponds to a 1.5‰ increase in δ13C values.

DISCUSSION

Preservation of primary isotopic signals

In the Orogrande Basin, strata older than the early Sakmarian are characterized by relatively low δ13C values. This trend of lower δ13C values for Gzhelian-lower Sakmarian strata occurs systematically throughout the basin, including the San Andres Mountains, Franklin Mountains, Sacramento Mountains, and core samples from the Northwest Shelf (Fig. 4b). A similar δ13C trend for coeval strata from the adjacent Permian Basin was presented in Saller et al. (1999), where δ13C values for uppermost Pennsylvanian strata ranged between -3.5‰ and -5.5‰; Lower Permian (Wolfcamp) carbonate strata had δ13C values that ranged between -2.5‰ and +3.5‰.

On Figure 4, δ13C and δ18O data for Gzhelian through Asselian strata show a wide range of values (Fig. 4a). Specifically, δ13C values average between +4.0‰ and -6.0‰; δ18O values for the same strata average between -1.0‰ and -9.0‰. The presence of variable δ13C values and variable δ18O values (Fig. 4b) is consistent with isotopic trends expected to result from both meteoric and burial diagenesis (Lohmann, 1988; Choquette and James, 1990). Evidence for both meteoric and burial diagenesis is observed in petrographic and outcrop/core analysis (Figs. 6 and 7). Brightly luminescent matrix material from cathodoluminescence analysis (Fig. 6) suggests alteration in the presence of Mn²⁺ incorporated in a reducing environment. Also present are low-Mg calcite brachiopod shells, which can often remain unaltered during diagenesis, making them
Figure 6: Cathodoluminescence images from the study area. A.) Image from the Lower Sacramento Mountains section (73m-Holder Formation). Arrow points to dully luminescent brachiopod shell indicating alteration; B.) Image from the Lower Sacramento Mountains section (262m-Hueco/Abo Formation). Arrows point to mostly non-luminescent brachiopod shell indicating little or no alteration; C.) Image from the Nicor DDH-2 Core (16.6m-Florida Shelf-Bursum Formation). Arrow points to a partially altered productid brachiopod spine; D.) Image from the Getty No. 5 Williard Beatty Core (19.5m-Northwest Shelf-Bough unit); arrow points to an altered phylloid algae fragment; E.) Image from the Franklin Mountains section (653m-Alacran Mountain Formation). Arrow points to a mostly non-luminescent brachiopod shell indicating little or no alteration; F.) Image from the Franklin Mountains section (237m-Cerro Alto Formation). Arrows point to non-luminescent brachiopod shells indicating little or no alteration.
Figure 7: Petrographic and outcrop images showing diagenetic alteration. A.) Petrographic image from the Lower San Andres Mountains section (4.0m-uppermost Panther Seep Formation). Arrow points to stylolitic contacts surrounding a gastropod shell suggesting burial diagenesis. B.) Image from the Unocal State No. 1-33 Core (13.9m-Bough Member B) showing stylolite suggesting burial diagenesis. C.) Image from the Getty No. 5 Williard Beatty Core (64.8m-Bough Member B) showing brecciated exposure surface suggesting meteoric alteration. D.) Petrographic image from the Franklin Mountains section (483m-Cerro Alto Formation) showing a partially dolomitized and compacted crinoid grain (center of image) with stylolitic contacts suggesting burial diagenesis. E.) Image from the Franklin Mountains section (285m-Cerro Alto Formation) showing wavy bedding suggesting pressure dissolution during burial diagenesis. F.) Petrographic image from the Upper Sacramento Mountains section (48m-San Andres Formation) showing broken and compacted ostracod grain suggesting burial diagenesis.
ideal for stable isotopic analysis (Popp et al., 1986; see Fig. 6b, 6e, 6f). Multiple exposure surfaces were documented throughout the study area, often at the top of ~5 m thick shallowing upward cycles (see Fig. 7); this suggests that strata were exposed to the meteoric diagenetic environment. Incorporation of soil-derived CO2 may have contributed to lower \( \delta^{13}C \) values in these strata (Lohmann, 1988). Low \( \delta^{18}O \) values could reflect both burial and meteoric influences (Lohmann, 1988; Choquette and James, 1990). As such, the isotopic composition of these samples likely reflects a mixture of pristine (i.e., primary), meteorically altered, and burial diagenetic signals. The diagenetically altered (burial and meteoric) phases represent a local/regional influence that has destroyed the primary isotopic signal.

The \( \delta^{13}C \) values for Sakmarian through Kungurian strata (Fig. 4a) are higher and less variable, ranging between +1.5‰ and +5.5‰; \( \delta^{18}O \) values for the same strata are between -2.0‰ and -9.0‰. The presence of relatively invariant \( \delta^{13}C \) values and variable \( \delta^{18}O \) values (Fig. 4b) is consistent with isotopic trends expected to result from burial diagenesis (Choquette and James, 1990). Although the \( \delta^{13}C \) values for these younger strata do vary by as much as 4‰, this is still much less than the 10‰ variation documented in the Gzhelian-Asselian strata (Fig. 4a). The distribution of \( \delta^{13}C \) values for Sakmaringian through Kungurian strata is consistent with alteration in a burial diagenetic environment (Choquette and James, 1990). Additionally, a petrographic and outcrop analysis also supports alteration in a burial diagenetic environment. In the Franklin Mountains and Upper Sacramento Mountains, stylolites (outcrop and thin-section), compaction of grains, blocky cement, and wavy bedding from pressure dissolution all provide evidence for burial diagenesis (Fig. 7d-f). Brightly luminescent matrix material
from cathodoluminescence analysis (Fig. 6) also supports a burial diagenetic environment. By contrast, no examples of intraformational exposure surfaces were documented in any of the Sakmarian-Kungurian strata analyzed for this study. A sequence stratigraphic analysis of these strata based on facies relationships and stacking patterns in the basin suggests a few exposure events may be present in certain covered intervals (Koch, 2010), but this cannot be demonstrated because of a lack of outcrop. Even if a few exposure events occurred in these strata, they are not common, as compared to the extensive exposure events in Gzhelian-lower Sakmarian strata. The available stable isotopic, petrographic, and outcrop data suggest the dominant form of alteration was from burial (not meteoric) diagenesis for Sakmarian-Kungurian strata. Because the burial environment tends to preserve primary $\delta^{13}C$ values in marine carbonates (see discussion below; Brand and Veizer, 1981; Banner and Hanson, 1990; Choquette and James, 1990), it is likely that the $\delta^{13}C$ values in Sakmarian-Kungurian strata represent a primary, and potentially global, signal.

Given the trends in carbon versus oxygen presented on Figure 4 and a stratigraphic/petrographic analysis, lower $\delta^{13}C$ values (as low as -5.9‰) for Gzhelian-lower Sakmarian strata in the Orogrande Basin are attributed to subaerial exposure and do not record primary marine compositions. The lower $\delta^{13}C$ values originated from the addition of dissolved inorganic carbon (likely sourced by the oxidation of organic matter) in a meteoric water environment (Lohmann, 1988; Patterson and Walter, 1994; Theiling et al., 2007). These inferences from the Orogrande Basin are consistent with other late Paleozoic isotope studies that have documented lower $\delta^{13}C$ values as reflecting episodic
subaerial exposure of marine carbonates (Algeo, 1996; Immenhauser et al., 2002; Elrick and Scott, 2010).

Conversely, lower Sakmarian through Kungurian strata, which includes Excursions 1 and 2, were not likely exposed for significant periods of time during deposition due to the relatively invariant δ¹³C values (Fig. 4) and lack of exposure surfaces. Although Excursions 1 and 2 are associated with slightly lower δ¹³C values, the total range in values (~4‰) is not large enough to convincingly indicate meteoric diagenesis. Furthermore, the low δ¹³C values during the Asselian-early Sakmarian are still much lower (by several per mil) than δ¹³C values documented for Excursions 1 and 2 (see Figs. 3-5). These strata, however, were impacted by burial diagenesis based on the presence of stylolites, dissolution bedding, and grain compaction (Fig. 7). During burial diagenesis, the δ¹³C value of mineral precipitates quickly equilibrates to the δ¹³C value of the host rock because of the large amounts of carbon in the CaCO₃, relative to the diagenetic fluids (Brand and Veizer, 1981; Banner and Hanson, 1990; Choquette and James, 1990). The δ¹³C values of precipitating cements are generally invariant and resemble the composition of the host rock. The δ¹⁸O values in mineral precipitates are directly impacted by increasing burial and pore water temperatures, resulting in varying isotopic compositions (Banner and Hanson, 1990; Choquette and James, 1990; Brand, 2004). As such, the relatively invariant δ¹³C values in lower Sakmarian through Kungurian strata likely reflect primary depositional compositions that could potentially be globally significant. These inferences are consistent with other isotopic studies that have previously demonstrated the preservation of primary δ¹³C values in upper Paleozoic carbonates (e.g., Saltzman, 2003; Batt et al., 2007).
Current understanding of the late Paleozoic ice age

The LPIA was the longest icehouse period of the Phanerozoic, and previous studies have viewed this icehouse world as a single protracted glacial epoch, lasting between 60-80 Ma (e.g., Veevers and Powell, 1987; Crowley and Baum, 1991; 1992; Frakes et al., 1992). More recently, Isbell et al. (2003) reviewed the stratigraphic data from across Gondwana and demonstrated that the LPIA could be divided into three discrete periods of glaciation, two in the Carboniferous (Glacial I and II) and one in the latest Carboniferous-Sakmarian (Glacial III). These glacial epochs were separated by times of non-glacial conditions. Workers in eastern Australia further resolved the claim of distinct glacial intervals separated by periods of non-glaciation (Fielding et al. 2008a). In eastern Australia, the LPIA can be divided into eight discrete glacial epochs (four in the Carboniferous and four in the Permian), each lasting 1-8 million years; glacial epochs were separated by non-glacial intervals of equal duration.

Recent workers studying ice-proximal deposits across Gondwana have created a detailed stratigraphic framework for glacial epochs during the LPIA (Fielding 2008b; 2008c and references therein). These stratigraphic studies suggest that the maximum extent of glacial ice occurred during the Early Permian, instead of the long-held view of a Late Pennsylvanian acme (Veevers and Powell, 1987; Gonzalez-Bonorino and Eyles, 1995; Heckel, 2008). Proxy records using stable isotopic data from terrestrial and marine carbonates suggest relatively mild global climate conditions during the Gzhelian, which then shifted to a significantly cooler climate during the Early Permian; this climate change was coeval with a dramatic increase in Gondwanan ice volumes (Montañez et al. 2007). Recent studies also suggest glaciation had ended by the end of the Middle
Permian (Fielding et al., 2008b). During the LPIA acme in the Early Permian, glacial deposits are known from across Gondwana, including South America, southern Africa, the Middle East, India, eastern Australia, and Western Australia (Veevers and Tewari, 1995; Fielding et al., 2008a; Holz et al., 2008; Isbell et al., 2008a; Martin et al., 2008; Mory et al., 2008; Stollhoffen et al., 2008). Early Permian glacial deposits also are known from Siberia, but the exact chronostratigraphic position of these sediments is somewhat ambiguous (Raymond and Metz, 2004).

During the latest Pennsylvanian and prior to the LPIA acme, glacial deposits are known from several regions of Gondwana, including South America, southern Africa, the Middle East, India, and Western Australia (Veevers and Tewari, 1995; Holz et al., 2008; Isbell et al., 2008a; Martin et al., 2008; Mory et al., 2008; Stollhoffen et al., 2008). No glacial deposits are known from Antarctica and eastern Australia during the latest Pennsylvanian. A lack of high-resolution stratigraphic control in many of the locations across Gondwana makes it difficult to resolve the exact timing of glacial epochs. A possible exception is the highly resolved stratigraphic record from eastern Australia (Fielding et al., 2008a).

The record of late Paleozoic glaciation in eastern Australia is currently the most highly resolved stratigraphic framework from across Gondwana. This is primarily a result of thick stratigraphically complete successions, combined with well constrained radiogenic and biostratigraphic ages (Fielding et al., 2008a). Eight distinct glacial epochs are recognized, each ranging from 1-8 million years in duration. Each glacial epoch contains multiple advance and retreat cycles (Fielding et al., 2008a; Fielding et al., 2008b; Birgenheier et al., 2009). Separating each glacial epoch are non-glacial intervals.
The four distinct Carboniferous glacial epochs (C1-C4) occurred from the Late Mississippian through the mid-Pennsylvanian (Fielding et al., 2008a). The Permian record consists of four non-overlapping glacial epochs (P1-P4) that occurred from the Early Permian through the mid-Permian, with glacial epochs P1 and P2 being the most extensive (Fielding et al., 2008a). Due to the high stratigraphic resolution of the eastern Australian glacial record, it has the potential to be used as a framework in which to compare $\delta^{13}C$ trends from the Orogrande Basin in New Mexico.

**Brief overview of global Early Permian proxy trends**

Previous stable isotopic compilations for the upper Paleozoic published by Veizer *et al.* (1999), Korte *et al.* (2005), and Grossman *et al.* (2008) have been interpreted to represent global trends, but were based on relatively few data points for the Permian. Frank *et al.* (2008) compiled previously published data sets, including Veizer *et al.* (1999) and Korte *et al.* (2005). They inferred that global climate conditions became much cooler during the earliest Permian, partially based on a ~2‰ increase in $\delta^{18}O$ values suggesting a significant expansion of global ice volume. Additionally, Frank *et al.* (2008) inferred a ~1‰ to 2‰ increase in global $\delta^{13}C$ values during the Pennsylvanian-Permian transition, which is consistent with a drop in global $pCO_2$ levels. The Early Permian drop in $pCO_2$ levels could have partially resulted from an increase in ocean productivity and sequestering of $^{12}C$ associated with glacial conditions.

An earliest Permian shift to colder global conditions is further supported by low $pCO_2$ (atmospheric carbon dioxide) values inferred from materials such as pedogenic carbonate, (Ekart *et al.* 1999; Fig. 5). Montañez *et al.* (2007) used brachiopod data from
marine carbonates, pedogenic carbonate, and fossil organic matter to document short-term δ¹³C and δ¹⁸O variations during the latest Pennsylvanian-Middle Permian. They suggested that changes in pCO₂, eustasy, global temperatures, and ice volume correlate with stratigraphic records documenting glacial epochs from Gondwana (Isbell et al., 2003; Fielding et al., 2008a). More recently, Birgenheier et al. (2010) used organic matter-rich facies to construct a bulk δ¹³C<sub>org</sub> record for eastern Australia. Their results suggest that fluctuations in δ¹³C<sub>org</sub> (variations in burial rates of organic matter) correlate with global trends of pCO₂ documented in previous studies (e.g., Montañez et al., 2007).

Variations in the bulk δ¹³C<sub>org</sub> record from Birgenheier et al. (2010) correlated with high-resolution stratigraphic records from the same strata (Fielding et al., 2008a), suggesting the glacial epochs in eastern Australia correspond to global changes in pCO₂ during the Permian (Montañez et al., 2007).

Near the end of the Early Permian, δ¹³C and δ¹⁸O values began a general decline, which continued through the Late Permian (Korte et al., 2005; Frank et al., 2008). This trend was interpreted as representing a general increase in pCO₂ levels associated with the end of major Gondwanan glacial epochs.

In addition to stable isotopic studies, a recent compilation of eustatic trends has suggested that the largest fluctuations in eustasy (up to ~120 m) occurred from the Late Pennsylvanian to the mid-Sakmarian, which likely resulted from glacioeustasy (Rygel et al., 2008; refer to Fig. 5). After the mid-Sakmarian, glacioeustatic change was limited to less than 70 m, which likely reflects an overall decrease in glacial ice across Gondwana.
A climate signal preserved in the Orogrande Basin?

Extensive subaerial exposure during the latest Pennsylvanian-earliest Permian in the Orogrande Basin suggests a prolonged drop in relative base level, which was overprinted by high-frequency sea-level change that gave rise to depositional cycles. Likely contributing factors include tectonic uplift or eustasy. The latest Pennsylvanian-earliest Permian was a time of tectonic activity in the western part of Euramerica due to the final assembly of Pangea (Ross, 1986; Scotese and Langford, 1995; Golonka and Ford, 2000; Blakey, 2008). However, previous research in the Orogrande Basin suggests that large-scale tectonic activity had slowed substantially by the end of the Pennsylvanian (Jordan, 1975; Candelaria, 1988). Major zones of uplift that achieved maximum relief during the Virgilian (e.g., Pedernal Landmass, etc.) were mostly eroded by the end of the Wolfcampian. Towards the end of the Wolfcampian, the central and northern part of the basin was completely filled with terrigenous clastics, and carbonate deposition was restricted to the southern regions such as the Franklin Mountains (Jordan, 1975; Candelaria, 1988). Relative sea-level rise occurred in the Kungurian, and the region returned to a shallow marine environment (Kottlowski, 1975; Milner, 1976; Lindsay and Reed, 1992; Raatz, 2002). It is unlikely that tectonic uplift was entirely responsible for subaerial exposure of carbonate strata into the mid-Sakmarian.

In light of eustatic and isotopic trends (Rygel et al., 2008; Frank et al., 2008), and Gondwanan stratigraphic records (Fielding et al., 2008a-c), it is inferred that glacioeustatic change was the most likely candidate controlling subaerial exposure in the Orogrande Basin during the latest Pennsylvanian-earliest Permian. Specifically, Gzhelian-early Sakmarian variable δ\textsuperscript{13}C values from the Orogrande Basin likely represent
a local/regional signal (of subaerial exposure) that was ultimately controlled by global processes (i.e., glacioeustasy). The end of major subaerial exposure in the Orogrande Basin during the mid-Sakmarian corresponds temporally to the end of major glaciation across much of Gondwana (Fielding et al., 2008b), including the end of the P1 glacial epoch in eastern Australia (Fielding et al., 2008a). The inference of major glacioeustatic fluctuations ending in the mid-Sakmarian is also consistent with global compilations documenting eustatic trends during the Permian (Rygel et al., 2008).

Although major glaciation ended across much of Gondwana in the mid-Sakmarian, small-scale ice centers persisted locally in eastern Australia and Africa until the end of the Middle Permian (Isbell et al. 2003; Fielding et al. 2008a); small ice centers also persisted in Siberia during the Middle Permian (Raymond and Metz, 2004). These smaller-scale events from Gondwana may be used to explain some of the other $\delta^{13}C$ trends from the Orogrande Basin. For example, the interval of time that includes Excursion 1 roughly correlates with the P2 glacial epoch in eastern Australia (Fig. 5). The lowering of $\delta^{13}C$ values during Excursion 2 roughly correlates with glacial epochs in Western Australia, as well as the beginning of the P3 glacial epoch in eastern Australia (Excursion 2 on Fig. 5). These inferences suggest that the smaller-scale glaciations were not merely regional scale, but may have affected distal locations, such as the Orogrande Basin. Furthermore, these inferences from the Orogrande Basin are consistent with global isotopic studies (e.g., Isozaki et al., 2007; Montañez et al., 2007) that suggest the P2 and P3 glacial epochs in eastern Australia were of global significance.

The $\delta^{13}C$ values in the two negative excursions during the P2 and P3 glacial epochs do not suggest subaerial exposure in the Orogrande Basin, as compared to the
strata deposited during the P1 glacial epoch (Figs. 4 and 5). However, a ~1.5‰ drop in δ¹³C values during these small-scale events does suggest that conditions within the basin were indirectly responding to glacial fluctuations across Gondwana. Patterson and Walter (1994) demonstrated that δ¹³C values on interior portions of carbonate platforms can become depleted in ¹³C through exposure and/or an increase in salinity. They showed that even in slightly restricted conditions δ¹³C values can be lowered by as much as 4‰, relative to open-ocean conditions. The 1.5‰ drop in δ¹³C values during Excursions 1 and 2 is well within the range of possible δ¹³C values for restricted water described by Patterson and Walter (1994). Furthermore, the facies deposited during these excursions show a shallowing, from deeper ramp settings with phylloid algae bioherms to shallower, storm dominated sediments deposited higher up on the carbonate ramp (Jordan, 1975; Kottlowski, 1975; Milner, 1976; Lindsay and Reed, 1992; Koch, 2010). There is little evidence of extremely hypersaline depositional environments during Excursion 1. The 1.5‰ decrease in δ¹³C values for this time interval likely represent conditions that were slightly more restricted than the open marine environment, but not hypersaline. By contrast, lower San Andres Formation strata appear to have been deposited in much shallower conditions than the Cerro Alto Formation (i.e., intertidal-restricted environment of Facies Association A). The lowering of δ¹³C values during Excursion 2 likely resulted from hypersaline conditions in the basin during this time interval.

Restriction, which best explains both δ¹³C negative excursions was likely a result of a lowering of relative sea level in the basin, but not to the point where the carbonate platform was subaerially exposed for significant periods of time. As relative sea level
dropped, it is likely that the basin could have become partially disconnected from the open marine environment, leading to restricted conditions.

The carbon isotope interpretation for Excursions 1 and 2 suggests that the P2 and P3 glacial epochs in eastern Australia were less extensive than the P1 glacial epoch that occurred during the earliest Permian. Otherwise, evidence for significant subaerial exposure should be recognized, as is the case with uppermost Pennsylvanian-earliest Permian strata in the study area (Fig. 5). This is not consistent with stratigraphic and isotopic studies from eastern Australia that suggest the P2 glacial epoch was as extensive as the preceding P1 glacial epoch (Fielding et al., 2008a; Birgenheier et al., 2010). If an increase in subsidence occurred in the Orogrande Basin during the late Sakmarian-Kungurian, this could have worked to limit the amount of subaerial exposure (from glacioeustatic drop) experienced by carbonate deposits, which would make the P2 glacial epoch appear smaller than the P1. Yet, stratigraphic studies of the Orogrande Basin suggest that subsidence actually decreased during the late Sakmarian-Kungurian (Jordan, 1975; Candelaria, 1988). This apparent discrepancy between the Orogrande Basin and Gondwana during the P2 glacial epoch is one of the many questions that still remain unanswered, and will require additional research.

CONCLUSIONS

1.) Uppermost Pennsylvanian-Lower Permian carbonates from the Orogrande Basin show evidence of subaerial exposure, which is evident by significantly lower δ^{13}C values, compared to global trends (Korte et al., 2005; Frank et al., 2008; Grossman et al., 2008). Episodic subaerial exposure occurred primarily from the Gzhelian
through the mid-Sakmarian and corresponds to a period of increased glaciation across Gondwana (e.g., P1 glacial epoch in eastern Australia; see Fielding et al., 2008a). Because regional tectonic events were occurring during the Pennsylvanian-Permian transition, it is possible that some of the subaerial exposure could have resulted from tectonic uplift. However, previous Orogrande Basin studies have concluded that the majority of tectonic influence in the region occurred in the Late Pennsylvanian (Jordan, 1975; Candelaria, 1988), which suggests Early Permian subaerial exposure was primarily controlled by glacioeustatic fall. In strata younger than the early Sakmarian, δ¹³C values fall roughly within the range of global values that have been compiled by previous late Paleozoic stable isotope studies (Korte et al., 2005; Frank et al., 2008; Grossman et al., 2008).

2.) Two smaller-scale excursions of decreasing δ¹³C values are inferred for upper Sakmarian-middle Artinskian and Kungurian strata (Excursions 1 and 2). Both trends show a 1.5‰ decrease in δ¹³C values. These lower δ¹³C values are associated with a facies change to Shallower and slightly more restricted conditions. These negative excursions occurred during periods of renewed expansion of glacial ice across parts of Gondwana (i.e., P2 and P3 glacial epochs in eastern Australia). Unlike the major subaerial exposure inferred for Gzhelian-early Sakmarian strata, the δ¹³C values do not suggest major exposure occurred during these younger glacial epochs. Rather, the lower δ¹³C values suggest basin shallowing and a shift to slightly higher salinity levels from increased evaporation, which may have resulted from less connection with the open marine environment due to lower eustatic levels. Lower
inferred magnitudes of relative sea-level change during the Sakmarian through the Kungurian are consistent with global eustatic trends (Rygel et al., 2008).

3.) $\delta^{13}C$ values from the Orogrande Basin suggest glacioeustatic fluctuation was greatest during the latest Pennsylvanian-early Sakmarian, leading to episodic subaerial exposure. Conversely, large-scale subaerial exposure is not interpreted to have occurred in strata younger than the mid-Sakmarian. This suggests that younger glacial epochs on Gondwana (i.e., P2 and P3 in eastern Australia) did not create enough glacioeustatic change to expose carbonate strata of the Orogrande Basin for long intervals of time from the mid-Sakmarian through the Kungurian. This interpretation is not consistent with stratigraphic and isotopic evidence from eastern Australia, which suggests the P2 glacial epoch was nearly as extensive as the preceding P1 (Fielding et al., 2008a; Birgenheier et al., 2010).

4.) The data generated in this study contributes to much needed isotopic values for the Early Permian. However, additional isotopic studies of other Lower Permian carbonate environments are needed to determine if the $\delta^{13}C$ trends in the Orogrande Basin resulted primarily from yet undocumented local to regional tectonic effects (e.g., subsidence trends), or whether they represent a truly global signal.

ACKNOWLEDGEMENTS

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CHAPTER 4:
THE PENNSYLVANIAN-PERMIAN TRANSITION IN THE LOW-LATITUDE CARBONATE RECORD AND THE ONSET OF MAJOR GONDWANAN GLACIATION

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ABSTRACT

Recent studies suggest a marked expansion of glacial ice across much of Gondwana beginning in the earliest Permian. Because expansion of glacial ice results in a lowering of sea level, the imprint of ice expansion should be evident worldwide as significant exposure event, hiatuses, or other evidence for sea level drop at or near the Pennsylvanian-Permian boundary. This literature review investigates the signature of an Early Permian expansion of Gondwanan ice through examination of stratigraphic records from eight carbonate-dominated, paleotropical regions across Pangea. Tropical carbonate environments are used because most form in tectonically quiescent regions and are sensitive indicators of eustatic change. Correlation between stratigraphic sections is achieved using the most current biostratigraphic and absolute time constraints available.

All studied sections show a sequence boundary or basinward shift in facies at or near the Pennsylvanian-Permian boundary, supporting the hypothesis of a significant expansion of glacial ice and global eustatic lowstand beginning in the Early Permian. By contrast, a series of mid-Sakmarian-Kungurian transgression events in the paleotropics are interpreted to reflect the asynchronous deglaciation of Gondwana.

The stratigraphic framework developed herein will allow for better correlations among stratigraphic records from Gondwana and northern Pangea, which will ultimately improve the understanding of how carbonate systems respond to global icehouse conditions, such as the late Paleozoic.
INTRODUCTION

The late Paleozoic was a time of major climatic and environmental change that saw large-scale fluctuations in the size and distribution of Gondwanan ice sheets. It has been proposed that the Pennsylvanian-Permian transition (hereafter PPT) is a key interval of time during the late Paleozoic ice age (LPIA) because it represents a period when ice volumes increased dramatically across much of Gondwana (Isbell et al. 2003; Fielding et al. 2008a-c). If a significant increase in Gondwanan ice did occur during the earliest Permian, then sedimentary environments should have recorded the major expansion of global ice volume as a series of significant exposure events or hiatuses reflecting eustatic drawdown at or near the Pennsylvanian-Permian boundary. Tropical carbonate platforms (including traditional platforms, epeiric platforms, and ramps) are used in this study because they are especially sensitive indicators of eustatic change. Furthermore, these low-latitude environments are not subject to ice-proximal influences such as isostatic loading from glacial ice, and they often occur in tectonically quiescent areas. For these reasons, stacking patterns in tropical carbonates provide proxies for past eustatic events (i.e., transgressions, regressions, unconformities, etc.).

In this paper, the stratigraphic records from eight carbonate-dominated, paleotropical regions are examined: 1) U.S. Midcontinent; 2) Orogrande Basin, New Mexico; 3) Permian Basin, Texas; 4) Arctic North America; 5) Bolivia; 6) South China Platform; 7) Russian Platform; 8) Barents Shelf (Fig. 1). These regions are analyzed in terms of lithology/cyclicity changes across the Pennsylvanian-Permian boundary, as well as the stratigraphic position of major surfaces including unconformities and sequence boundaries. Results support an increase in global ice volume during the earliest Permian,
Figure 1: Map of locations used in this study (denoted by grey circles). Continental configuration based on reconstructions from Golonka and Ford (2000).
which clarifies how low-latitude carbonate deposition responded to a major shift in global climate conditions. An improved understanding of the paleotropics during the PPT allows for better global stratigraphic correlations across all of Pangea during the LPIA.

**BACKGROUND**

Until recently, the long-held view of the LPIA was one of a single widespread glaciation event that covered much of Gondwana from the mid-Late Mississippian through the Early Permian (e.g., Veevers and Powell 1987). This view, based mainly on information from Euramerican cyclothem deposits, suggested that glacial ice volumes reached a peak during the Late Pennsylvanian (Veevers and Powell 1987; Heckel 1994; 2008). However, Isbell et al. (2003) reviewed the stratigraphic data from across Gondwana and demonstrated that the LPIA could be divided into three distinct periods of glaciation (two in the Carboniferous and one in the latest Carboniferous-Sakmarian), separated by times of non-glacial conditions. The claim of discrete glacial intervals separated by warmer time periods during the LPIA was further resolved by work in eastern Australia (Fielding et al. 2008a), where the LPIA can be divided into eight distinct glacial epochs (four in the Carboniferous and four in the Permian), each lasting 1-8 million years, which are separated by non-glacial intervals of equal duration. Additionally, studies based on compilations of data from other Gondwanan paleofragments indicate that multiple ice centers expanded during the Early Permian in places like eastern Australia, Antarctica, India, southern Africa, South America, and the Middle East, which suggests an acme in the earliest Permian (Fielding et al. 2008b; Fielding et al. 2008c and references therein).
An Early Permian glacial expansion is also supported by geochemical proxy records of ice volume, $pCO_2$ (atmospheric carbon dioxide), and temperature. Late Pennsylvanian-Early Permian stable isotopic records from around the world are consistent with a drop in atmospheric $pCO_2$ in the earliest Permian, as indicated by carbon isotope records from marine carbonates, soil-forming minerals, and sedimentary organic matter (Ekart et al. 1999; Montañez et al. 2007; Birgenheier et al., 2010; see Fig. 2). During the PPT, a ~2‰ increase in $\delta^{18}O$ values from low-latitude marine carbonates is consistent with a significant expansion of global ice volume (Frank et al. 2008). As with oxygen isotope compositions, a ~1‰ to 2‰ increase in $\delta^{13}C$ values occurred during the PPT, which is consistent with a drop in $pCO_2$ and a possible increase in ocean productivity associated with glacial conditions (Frank et al. 2008; Grossman et al. 2008; see Fig. 2). Additionally, a recent compilation of late Paleozoic global sea-level change (Rygel et al. 2008) suggests large-scale fluctuations (in both magnitude and frequency) of eustatic change during the latest Pennsylvanian-earliest Permian, suggesting an increase in global ice volumes.

The notion of a major expansion of Gondwanan glacial ice during the Early Permian is not without contention. The Euramerican cyclothem framework has long been considered an accurate and reliable eustatic record of Gondwanan glacial fluctuations, especially for the Carboniferous (Heckel, 1986; 1994; Wright and Vanstone, 2001; Heckel, 2008). Disagreement has arisen resulting from the inconsistency among Gondwanan records suggesting a LPIA acme during the Early Permian (e.g., Isbell et al. 2003; Fielding et al. 2008a) and paleotropical cyclothem deposits, which have been interpreted to suggest a peak in glaciation during the Late Pennsylvanian (e.g., Veevers
and Powell, 1987; Heckel, 1994; Gonzalez-Bonorino and Eyles, 1995; Crowell, 1999; Heckel, 2008). However, the review by Isbell et al. (2003) concluded that the pre-Permian glacial epochs were not extensive enough to have been the primary mechanism that controlled the well documented base level changes in the Euramerican cyclothem record.

**METHODOLOGY AND TERMINOLOGY**

This study utilizes published stratigraphic data from across tropical and subtropical Pangea (Fig. 1; Table 1). Since hundreds of papers have been published concerning upper Paleozoic stratigraphic records, this study focused on sources that describe stratigraphically complete records that possess a high level of chronostratigraphic control. Specific stratigraphic information outlined by individual authors (paleogeographic location, lithology/facies, changes in cyclicity, and stratigraphic location of unconformities) was combined with the most robust biostratigraphic and absolute constraints available (e.g., Davydov et al. 1998; Ramezani et al. 2007; Figs. 2-3). The compiled records were calibrated to the Gradstein et al. (2004) time scale to produce a global stratigraphic framework. Stratigraphic data are plotted against the most recent reconstructions of Gondwanan glacial epochs (Fielding et al. 2008c and references therein), as well as other global climate change proxies such as oxygen and carbon stable isotopic trends (e.g., Frank et al. 2008; Grossman et al. 2008; Birgenheier et al., 2010), $p$CO$_2$ trends (Ekart et al. 1999; Montañez et al. 2007), and eustatic records (Rygel et al. 2008).
Table 1.— *Summary of stratigraphic data from locations examined in this study.*

<table>
<thead>
<tr>
<th>Region</th>
<th>Formations Studied</th>
<th>Paleolatitude</th>
<th>Nature of Pennsylvania-Permian Boundary</th>
<th>Primary Age Control</th>
<th>Other Stratigraphic Information (e.g., cycles, etc.)</th>
<th>Key References</th>
</tr>
</thead>
<tbody>
<tr>
<td>United States Midcontinent</td>
<td>Douglas, Shawnee, Wabaunsee, Admire, Council Grove, Chase, and Sumner Groups</td>
<td>L. Pennsylvaniaian: 0-5° N Golonka and Ford, 2000; E. Permian: 5-10° N (Scotese and Langford, 1995); Also, (Ross and Ross, 1990; Witzke, 1990)</td>
<td>Conformable boundary between the Glenrock Limestone and Bennett Shale members in the Red Eagle Limestone; overlying Asselian strata record a distinct basinward shift in facies.</td>
<td>Primarily fusulinids and conodonts</td>
<td>3rd-5th order cycles can be found in most of the strata (e.g., cyclothems). Major flooding events and sequence boundaries throughout the section.</td>
<td>Ritter (1995); West et al., 1997; Mazzullo (1998); Mazzullo (1999); Olszewski and Patzkowsky (2003); Sawin et al., 2006; Mazzullo et al. (2007); Heckel (2008)</td>
</tr>
<tr>
<td>Orogrande Basin</td>
<td>Madera and Hueco Groups, Abo, Yeso, Bursum, and San Andres Formations</td>
<td>L. Pennsylvaniaian: ± 5° N or 5° S (Golonka and Ford, 2000); E. Permian: -5° N (Scotese and Langford, 1995); Also, (Ross and Ross, 1990)</td>
<td>Regional unconformity at the &quot;Bursumian&quot;-Wolfcampian boundary (i.e., Penn-Permian boundary). Hiatus observed across the entire basin as exposure surfaces.</td>
<td>Primarily fusulinids; U-Pb ages</td>
<td>3rd 5th order cycles can be found in most of the strata. Major unconformity at Penn-Permian boundary.</td>
<td>Jordan (1975); Rankley et al. (1999); Candelaria (1988); Raatz (2002); Wahlman and King (2002); Mack et al. (2003)</td>
</tr>
<tr>
<td>Permian Basin (Glass Mountains, Midland Basin, Guadalupe Mountains)</td>
<td>Cisco Group, Gaptank, Neal Ranch, Lenox Hills, Wolfcamp, Hueco, Leonard, Victorio Peak, and Bone Spring Formations; Clear Fork Group</td>
<td>L. Pennsylvaniaian: ± 5° N or 5° S (Golonka and Ford, 2000); E. Permian: -5° N (Scotese and Langford, 1995); Also, (Ross and Ross, 1990)</td>
<td>Regional unconformity at the Penn-Permian boundary; latest Penn-earliest Permian strata appear to be missing in many places.</td>
<td>Primarily conodonts and fusulinids</td>
<td>3rd-5th order cycles can be found in most of the strata. Major hiatus at Penn-Permian boundary, and in upper Artinskian strata. Sakmarian hiatus in the Glass Mountains portion of the basin.</td>
<td>Ross (1963); Candelaria et al. (1992); Fitchen et al. (1995); Mazzullo (1995); Hill (1996); Yang et al. (1998); Saller et al. (1999); Yang and Komine (1999); Harrell and Lambert (2007); Tabor et al. (2008)</td>
</tr>
</tbody>
</table>
Table 1 cont’d. — Summary of stratigraphic data from locations examined in this study.

<table>
<thead>
<tr>
<th>Region</th>
<th>Formations Studied</th>
<th>Paleolatitude</th>
<th>Nature of Pennsylvania-Permian Boundary</th>
<th>Primary Age Control</th>
<th>Other Stratigraphic Information (e.g., cycles, etc.)</th>
<th>Key References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bolivia</td>
<td>Titicaca Group; Copacabana and Chutani Formations</td>
<td>L. Pennsylvanian: -30° S (Golonka and Ford, 2000); E. Permian: 20-25° S (Isaacson and Diaz-Martinez, 1995)</td>
<td>Major sequence boundary hiatus that coincides with the start of the P1 glaciation in eastern Australia (Fielding et al., 2008a),</td>
<td>Fusulinids, conodonts; also, corals, bryozoans</td>
<td>2nd-3rd order cycles throughout L. Penn., and E. Permian. A marked loss of cyclic deposits occurs in the Sakmarian. Major unconformities in Sakmarian, Artinskian strata; also at Penn-Permian boundary, major Sakmarian flooding event.</td>
<td>Dunbar and Newell (1946); Newell et al. (1953); Isaacson and Diaz-Martinez (1995); Sempere, (1995); Ottore et al. (1998); Grader et al. (2008)</td>
</tr>
<tr>
<td>South China</td>
<td>Chuanshan and Qixia Formations</td>
<td>L. Pennsylvanian: 0-15° S (Golonia and Ford, 2000); E. Permian: 0-15° S (Scotese and Langford, 1995); Also, (Nie et al., 1990)</td>
<td>Regional discontinuity or major facies change at Penn-Permian boundary. Evidence suggests platform was mostly exposed during much of Asselian-Sakmarian. Platform flooded in Artinskian.</td>
<td>Fusulinids, conodonts</td>
<td>3rd-5th order cycles can be found in most of the strata. Major sea level lowstand interpreted at Penn-Permian boundary based on facies changes.</td>
<td>Yang (1986); Meyerhoff et al. (1991); Enos (1995); Wang and Jin (2000); Shi and Chen (2006)</td>
</tr>
<tr>
<td>Russian Platform</td>
<td>Russian Platform (near the Moscow Basin)</td>
<td>L. Pennsylvanian: 15-25° N (Golonia and Ford, 2000); E. Permian: 20-30° N (Scotese and Langford, 1995); Also, (Scotese and McKerrow, 1990)</td>
<td>Penn-Permian boundary appears to be conformable in many places; significant Guhellen-Asselian hiatus/exposure events can be traced across the entire Russian Platform.</td>
<td>Conodonts, fusulinids, ammonoids; U-Ph ages</td>
<td>3rd-5th order cycles present throughout the section. Major unconformities in the Guhellen, Asselian, Sakmarian, and Artinskian strata.</td>
<td>Ross and Ross (1985); Churashov (1995); Davydov et al. (1998); Izart et al. (1990); Vennin et al. (2002); Davydov and Leven (2003); Izart et al. (2003); Ramezani et al. (2007); Vennin (2007)</td>
</tr>
<tr>
<td>Barents Shelf</td>
<td>Gipsdalien, Bjarmeland, and Tempelfjorden Groups</td>
<td>L. Pennsylvanian: 25-30° N (Golonia and Ford, 2000); E. Permian: 30-45° N (Scotese and Langford, 1995); Also, (Scotese and McKerrow, 1990)</td>
<td>Major regional unconformity across platforms. Only deep-water/high sediment influx areas do not show an unconformity. Significant emergence event hiatus at Penn-Permian boundary. L. Penn-earliest Permian strata missing in many areas.</td>
<td>Fusulinids, conodonts</td>
<td>3rd-5th order cycles from Late Penn. into E. Permian. A marked loss of cyclic deposits (flooding) occurs in late Sakmarian. Major late Artinskian hiatus; transition to cool-water carbonate deposition during the Artinskian.</td>
<td>Heafford (1988); Nakrem (1991); Stenersen and Worsley (1995); Harland and Geddes (1997); Stenersen (2000); Rafaelsen et al. (2008); Stenersen (2008)</td>
</tr>
</tbody>
</table>
Figure 2: Chronostratigraphic framework of the eight regions studied for this review. References for individual regions can be found in the text or on Table 1. Dark blue horizontal lines/shading represents lowstands in relative sea level (e.g. sequence boundaries/unconformities); horizontal red lines denote transgression events. Lithology symbols are included on the figure. References for Gondwanan glacial events, stable isotope/atmospheric carbon dioxide trends, and eustatic changes are included on the figure. Key conodont, fusulinid, and absolute ages used for correlation are denoted by green circles, stars, and squares, respectively. See the text and Figure 3 for more information on the biostratigraphy and original references for each section. Time scale based on Gradstein et al. (2004).
Figure 3: Biostratigraphic data used for age calibrations. Based on data presented in Gradstein et al. (2004 and references therein) and Ross and Ross (1988; 1995a).
In the body of this paper, the nature of the Pennsylvanian-Permian boundary for each region is described, along with a brief assessment of the stratigraphic data. Refer to the Appendix at the end of the paper for additional details regarding each region (e.g., paleogeography, systematic stratigraphy, biostratigraphic correlations, etc.).

Although the late Paleozoic represents a time of extensive global tectonic activity (Scotese and Langford 1995; Blakey 2008), most of the eight regions chosen for this study were tectonically quiescent during the PPT. One exception is the Permian Basin, parts of which underwent uplift and exposure during the PPT, related to the final assembly of Pangea (Ross 1986; Ross and Ross 1986). However, the Permian Basin is still included in this paper because it is one of the best understood in terms of age, depositional environment, and sequence stratigraphy.

The term sequence boundary, as used herein, refers to a regionally significant erosional hiatus or unconformity that separates distinct packages of strata (depositional sequences; Vail-Exxon model; see discussions in Vail et al. 1977; Catuneanu 2006). Sequence boundaries described in this paper are associated with subaerial exposure events, which are used to imply lowstands in relative sea level. It should be noted, however, that subaerial exposure is not necessarily a required element for the definition of a sequence boundary in the Vail-Exxon model (correlative conformities). For the literature utilized in this paper, the author’s original terminology was used to construct the global stratigraphic framework (Fig. 2). Regardless of which term is used by individual authors (sequence boundary, erosional unconformity, depositional hiatus, exposure surface), these events all refer to a significant drop in relative sea level for each
locality studied. Likewise, the term transgression is used to describe a regionally significant rise in relative sea level.

Many studies included in this paper describe different scales of cyclicity such as 3rd, 4th, and 5th-order cycles. However, a lack of consistency exists in the literature concerning the definitions for different scales of cyclicity. To be consistent, the definitions outlined in Catuneanu (2006), which follow Vail et al. (1991), are used herein: 1st-order- 50+ millions of years; 2nd-order- 3-50 million years; 3rd-order- 0.5-3 million years; 4th-order- 0.08-0.5 million years; 5th-order- 0.03-0.08 million years. Some authors (e.g., Heckel 2008; Stemmerik 2008) placed cycle durations within a Milankovitch framework; most commonly a 100,000 or 400,000-year eccentricity cycle; these frameworks were placed within the 4th or 5th-order cycle durations described above.

Biostatigraphic Correlations

The issue of biostratigraphic correlation between upper Paleozoic sections is complicated and resolution varies by region. Conodonts, fusulinids, and ammonoids, the most commonly used taxa for correlation, are used here (Fig. 3). To date, Gradstein et al. (2004 and references therein) provides the most robust global biostratigraphic framework for these taxa, with particular regard to their stratigraphic placement relative to official Stage boundaries for upper Paleozoic strata. The Pennsylvanian-Permian boundary discussed herein is based on the Global Stratotype Section and Point (GSSP) for the base of the Permian System, which was placed at a shallow marine shelf section, southeast of the Russian Platform in Kazakhstan (Davydov et al. 1998). This boundary corresponds
to the basal occurrence of the conodont *Streptognathodus isolatus*. Refer to the Appendix for more details about the biostratigraphic correlations used in this study.

**Global Correlations**

Although records can be correlated across northern Pangea with relative confidence due to the presence of cosmopolitan taxa such as conodonts (e.g., Menning et al. 2006), correlation between the northern Pangea and Gondwana remains difficult. In particular, the absence of conodonts in most Gondwanan sections hinders correlations between hemispheres. In some cases, such as Bolivia (part of Gondwana), fusulinids are present, which represent the primary biostratigraphic tool available to correlate these ice-proximal sections to northern Pangean records (Newell 1946; Newell et al. 1953; Grader et al. 2008). Absolute age control has solved some of these issues. For example, the Pennsylvanian-Permian boundary in Russia was dated using U-Pb ages from zircons (298.90 ±0.31 Ma; Ramezani et al. 2007). Additionally, recent isotope ages from Gondwanan sections in eastern Australia (Fielding et al. 2008a) has been used to constrain glacial epochs, which allows for more robust global correlations between ice-proximal and paleotropical successions.

**Data Limitations**

The data sets described in this paper all have potential biases and shortcomings. Age control is the most pertinent issue when performing global stratigraphic correlations. For this study, conodonts and fusulinids are used to determine the stratigraphic timing of individual events (unconformities, transgressions, etc.). However, even with conodont
data, there is a certain level of uncertainty in the stratigraphic records described above; specifically, within individual Stages (e.g., Asselian, Sakmarian, etc.). This study highlights the continuing need for increased biostratigraphic/radiometric resolution for all upper Paleozoic stratigraphic sections. Increased age control will allow for the construction of more highly resolved global stratigraphic frameworks. Until then, global stratigraphic research (i.e., this study) will contain a certain amount of uncertainty, which can impact interpretations.

Additionally, continuing sequence stratigraphic research is needed to determine the global sedimentological impacts resulting from eustatic change. For example, how fast (geologically) can global sequence boundaries form? Why does a eustatic drop create a major sequence boundary in one basin and a basinward shift in facies without a major unconformity in another? Further research could help answer these types of sequence stratigraphic questions, which can be used to help determine how glacioeustatic change is preserved in the stratigraphic record. Nonetheless, given the current state of sequence stratigraphy and biostratigraphic/radiogenic age control, the stratigraphic framework described below represents a robust synthesis of low-latitude stratigraphic data for the PPT.

REGIONAL STRATIGRAPHIC RECORDS

United States Midcontinent

The Pennsylvanian-Permian boundary in the U.S. Midcontinent occurs at the base of the Bennett Shale Member of the Red Eagle Limestone in the Red Eagle Composite Sequence, where black shale overlies shallow-water limestone (Ritter 1995; Olszewski
and Patzkowsky 2003; Fig. 2). This abrupt boundary is marked by a maximum transgressive surface within the Red Eagle Composite Sequence (Olszewski and Patzkowsky 2003; Sawin et al. 2006). The maximum transgression at the base of the Bennett Shale is one of the larger of such events in the Council Grove Group (Ritter 1995). Overlying the Red Eagle Limestone are stacked paleosols of the Roca Shale, which record maximum regression during deposition of the Red Eagle Composite Sequence (Olszewski and Patzkowsky 2003). As with the Red Eagle, the overlying Grenola Composite Sequence contains deeper water deposits within the Grenola Limestone, which are overlain by stacked paleosols (maximum regression) of the Eskridge Shale (Olszewski and Patzkowsky 2003; Sawin et al. 2006). Although the Pennsylvanian-Permian boundary in the Midcontinent is associated with a significant transgression event, several studies have documented a prolonged basinward shift in facies from the Red Eagle Limestone to the Beattie Limestone, which overlies the Eskridge Shale (West et al. 1997; Olszewski and Patzkowsky 2003; Sawin et al. 2006). This claim has also been supported by studies of brachiopod assemblages, which suggest a dramatic loss in taxa due to shallowing of the basin (Cycle 29 of Olszewski and Patzkowsky 2001). In general, there is an overall shallowing of facies that occurs throughout the Wolfcampian section of the Midcontinent. For example, typical deep water black shale facies of Pennsylvanian strata are rare to absent in Permian strata, and stacked paleosols increase throughout the Wolfcampian (West et al. 1997). It has been postulated that this Permian shallowing could have resulted either from increased glacial ice causing a eustatic drop, small-scale tectonically induced thermal uplift associated with
the assembly of Pangea, or some combination of the two (Ross and Ross 1987; West et al. 1997; Sawin et al. 2006).

Because of the robust biostratigraphic control (Thompson 1954; Ross and Ross 1988; 1995; Ritter 1995; Chernykh and Ritter 1997) and minimal tectonic influence (West et al. 1997; Mazzullo 1999), the U.S. Midcontinent section is ideal for studying PPT sea level change. In many of the other areas described below (except the Russian Platform), the Pennsylvanian-Permian boundary is an easily identifiable unconformity. However, the Midcontinent section, which is dominated by high-frequency (3rd and 5th-order) cyclic deposits, appears to record a gradual lowering of relative sea level during the Asselian (West et al. 1997; Olszewski and Patzkowsky 2003). Gradual sea-level fall is consistent with the slow build-up of glacial ice across Gondwana during the Early Permian.

**Orogrande Basin, New Mexico, USA**

The Pennsylvanian-Permian boundary in south-central New Mexico is expressed as a regional erosional unconformity, suggesting a major withdrawal of the sea (Candelaria 1988; Wahlman and King 2002; Fig. 2). An erosional unconformity at the Pennsylvanian-Permian boundary is also present across much of the southwestern U.S. (Ross 1986). It should be noted that Rasbury et al. (1998) reported several U-Pb ages, utilizing paleosol calcite, from the Sacramento Mountains (Beeman, Holder, and Laborcita Formations). Using biostratigraphic positions within cyclothem deposits, they documented the age of the Pennsylvanian-Permian boundary as 302 ±2.4 Ma, which was within the Laborcita Formation. Reevaluation of the original U-Pb ages, utilizing current
GSSP absolute ages and biostratigraphic horizons (Davydov et al. 1998; Ramezani et al. 2007), would place the System boundary at the top of the Laborcita Formation in the Orogrande Basin. In the Sacramento Mountains an angular unconformity occurs between the Holder and Laborcita Formations (i.e., below the Pennsylvanian-Permian boundary). This angular unconformity is present only in the Sacramento Mountains, which implies a local tectonic influence on sedimentation in the eastern part of the basin (uplift of the Pedernal Landmass) during the Late Pennsylvanian (Candelaria 1988; Rankey et al. 1999).

The use of fusulinids and U-Pb ages results in a fairly robust chronostratigraphic framework for the Orogrande Basin (Williams 1966; Steiner and Williams 1968; Rasbury et al. 1998; Wahlman and King 2002). The erosional unconformity at the Pennsylvanian-Permian boundary distinctly separates Pennsylvanian (i.e., “Bursumian”) fusulinid faunas from Permian (Nealian) assemblages (Wahlman and King 2002). Because large-scale tectonic uplift in the region had largely ended in the Late Pennsylvanian (Candelaria 1988), the formation of this erosional unconformity is consistent with a eustatic drop in the earliest Permian.

Permian Basin, Texas, USA

An erosional unconformity marks the boundary between Pennsylvanian and Permian strata in the Permian Basin (Candelaria et al. 1992; Fitchen et al. 1995; Mazzullo 1995; Hill 1996; Harrell and Lambert 2007; Fig. 2). In places, the boundary is expressed as an angular unconformity, which has been attributed to tectonic activity related to the final phase of thrusting within the Marathon orogenic belt (Ross 1986; Ross and Ross 2007).
1986; Hill 1996). A middle Wolfcampian unconformity also occurs in some areas such as the Glass Mountains, which is thought to be related to continuing regional tectonic activity during the Early Permian (Ross 1986; Fitchen et al. 1995; Hill 1996).

The stratigraphic record for this region is well-constrained due to conodont and fusulinid studies (Ross 1963; 1995; Wardlaw and Davydov 2000), which allows for accurate correlations with the U.S. Midcontinent (Wardlaw and Davydov 2000). As such, these correlations make it possible to infer an overall drop in relative sea level during the Asselian. However, an angular unconformity at the Pennsylvanian-Permian boundary across much of the basin would suggest an overall tectonic cause related to Pangean assembly (Ross 1986; Ross and Ross 1986). Therefore, an earliest Permian drop in relative sea level likely accompanied tectonic influences to create a major regional unconformity across the PPT.

*Arctic North America*

The Pennsylvanian-Permian boundary in the North Alaska Platform is an erosional unconformity (Sub-Sadlerochit unconformity) that separates Wahoo Formation carbonates from Echooka Formation clastics (Bamber and Waterhouse 1971; Crowder 1990; Beauchamp 1995; Hanks et al. 2006; Fig. 2). It has been estimated that tens of meters of Wahoo Formation are missing as a result of widespread emergence and erosion of the platform (Crowder 1990).

Although not as robust as conodonts, fusulinids are the primary biostratigraphic indicator used for correlations. In this region, they are useful for distinguishing Pennsylvanian from Permian strata (Ross 1995). The unconformity that separates
Pennsylvanian-Permian strata is ubiquitous across the region and does not appear to be angular in nature. Rather, it is erosional, which suggests a relative fall in sea level, either from tectonic emergence or a eustatic drop. It has been suggested that the Early Permian was a time of overall subsidence in the region (Beauchamp 1995). Overall subsidence in the basin during the PPT would support a eustatic drop as the causal mechanism for a widespread subaerial/erosional unconformity.

Bolivia

A major sequence boundary occurs at the Pennsylvanian-Permian boundary within the carbonate dominated Copacabana Formation (at 299 Ma; see Fig. 2). Grader et al. (2008) correlated this erosional boundary with the beginning of the P1 glacial epoch in eastern Australia, of Fielding et al. (2008a). Higher up in the succession, a regionally extensive latest Sakmarian-early Artinskian flooding surface represents a period of high relative sea level throughout the region (Grader et al. 2008).

Both conodont and fusulinid studies have allowed for the construction of a robust stratigraphic framework (Dunbar and Newell 1946; Newell et al. 1953; Suárez-Riglos et al. 1987Ottone et al. 1998). The formation of the erosional unconformity across the PPT is consistent with a eustatic drop in the earliest Permian. Additionally, the formation of this sequence boundary does not appear to be related to regional tectonic uplift, which further supports a glacioeustatic control (Grader et al. 2008).
Southern China

The boundary between Pennsylvanian and Permian strata in southern China is marked in many areas by a disconformity and a discernable facies shift from open marine bioclastic limestone to a widespread (~500,000 km\(^2\)) large oncolite-dominated lithofacies up to 12 m thick (Yang et al. 1986; Meyerhoff et al. 1991; Shi and Chen 2006). This major shift in lithofacies has been attributed to a relative sea level lowstand across the South China Platform during the earliest Permian (Shi and Chen 2006). Other stratigraphic and sedimentologic data from across southern China also suggest that the stable cratonic South China Platform remained largely emergent during large portions of the Asselian-Sakmarian, and was then regionally flooded in the early Artinskian (Enos 1995). It should be noted that the PPT exposure event documented in southern China also has been observed in an oceanic atoll succession in Japan (Sano and Kanmera 1991).

Because conodonts and fusulinids are used (Yang 1986; Meyerhoff et al. 1991; Enos 1995; Wang 2000; Shi and Chen 2006), the base of the Permian system in China can be confidently correlated with the GSSP in Kazakhstan. The tectonic stability of the South China Platform during the PPT (Enos 1995; Shi and Chen 2006) suggests that base-level change was largely controlled by eustatic mechanisms. The disconformity and unique oncolite facies present in the in the lowermost Permian strata suggests a significant lowering of relative sea level compared to the uppermost Pennsylvanian open marine bioclastic limestones (Shi and Chen 2006).
**Russian Platform**

As with the U.S. Midcontinent, the Pennsylvanian-Permian boundary in the Russian section is not associated with a major unconformity, but exhibits a general shallowing of facies through the Asselian Series. Lowermost Asselian strata consist of bioclastic wackestones deposited below fair-weather wave base; upper Asselian strata consist of bioclastic packstones deposited in wave agitated conditions (Vennin et al. 2002). Additionally, Asselian strata contain several widespread exposure events that do not appear to be related to a tectonic mechanism (Vennin et al. 2002).

The excellent chronostratigraphic control make this section the most well constrained record used for this study (Davydov et al. 1998; Izart et al. 1999; Vennin et al. 2002; Davydov and Leven 2003; Ramezani et al. 2007). As with the U.S. Midcontinent, the Russian Platform succession is herein interpreted as recording a more gradual eustatic drop during the PPT, culminating in a series of Asselian unconformities.

**Barents Shelf (Finnmark Platform); Svalbard**

A major exposure surface occurs at the Pennsylvanian-Permian boundary, which has been well-documented in Svalbard (Stemmerik and Worsley 1995; Fig. 2). The uppermost Pennsylvanian strata are missing in most parts of the Barents Shelf region, which suggests a large-scale drop in relative sea level during the PPT (Stemmerik and Worsley 1995). Additionally, much of the upper Wordiekammen (Kapp Duner Formation) shows evidence of karsting, suggesting repeated exposure during the Asselian (Nakrem 1991).
Because conodonts and fusulinids are used (Nakrem 1991; Harland and Geddes 1997; Nilsson and Davydov 1997; Anisimov et al. 1998; Stemmerik 2000) the chronostratigraphic control within this succession is fairly robust. During the PPT, this region was tectonically quiescent (Stemmerik and Worsley 1995). As such, the major unconformity during the PPT (Stemmerik and Worsley 1995) was most likely caused by a glacioeustatic drop.

**DISCUSSION**

*Synthesis of Global Stratigraphic Patterns*

With the exception of the Russian Platform and the U.S. Midcontinent, the Pennsylvanian-Permian boundary is everywhere represented by a major exposure surface, erosional unconformity, disconformity, or hiatus (Fig. 2). The apparent discordance of the U.S. Midcontinent and Russian stratigraphic records requires further evaluation. For the Russian Platform, Vennin et al. (2002) determined that the Asselian-Artinskian interval was dominated by largely regressive deposits associated with a drawdown in sea level. This Asselian-Artinskian “common regression” is evident as widespread exposure surfaces and a general transition from carbonate reef facies to very shallow, wave-influenced paleoaplysind boundstones (Vennin et al. 2002). Izart et al. (2003) suggested that high-frequency cyclic sedimentation patterns and subaerial unconformities in these strata most likely resulted from glacioeustatic controls. Although a major unconformity is not present at the Pennsylvanian-Permian boundary in the Russian section, the relative tectonic stability, basinward shift in facies in units overlying the boundary, and karsting
events in Asselian strata are interpreted as a protracted drawdown in relative sea level, which is consistent with Gondwanan ice build-up during the Early Permian.

In the U.S. Midcontinent, Ritter (1995) noted that the Pennsylvanian-Permian boundary lies within a major cycle of relative sea-level change, and is associated with a large transgression event. However, multiple studies suggest that facies above the boundary record a significant basinward shift (West et al. 1997; Olszewski and Patzkowsky 2001; Olszewski and Patzkowsky 2003; Sawin et al. 2006). Although there is no subaerial unconformity at the Pennsylvanian-Permian boundary, Asselian strata show an overall shallowing, which records a gradual drawdown in relative sea level, likely from a glacioeustatic mechanism.

For the Russian and U.S. Midcontinent successions, the gradual shallowing of facies during the Early Permian may have been largely influenced by thermal uplift and regional tectonic warping of the cratonic surface associated with the final assembly of Pangea (West et al. 1997; Watney et al. 2006). As such, high-frequency glacioeustatic cycles (Olszewski and Patzkowsky 2003) are likely superimposed upon a 2nd or 3rd–order tectonic emergence signal, which may help explain the gradual drawdown of relative sea level during the Asselian.

Although not globally synchronous, major transgression events are recorded in many of the Lower Permian stratigraphic records (see Fig. 2). Many of these regionally significant transgressions occurred during or after the mid-Sakmarian and are often attributed by authors to eustatic rise following the major deglaciation of Gondwana (Grader et al. 2008; Stemmerik 2008). In three of the studied successions, the Barents Shelf, North Alaska, and Bolivia, transgressions correspond with a loss in well-defined
depositional cyclicity (Beauchamp and Baud 2002; Grader et al. 2008; Stemmerik 2008). A lack of global synchronicity implies that these post-Sakmarian transgression events may reflect local/regional changes in sedimentation and tectonics, rather than a single eustatic rise following deglaciation. Additionally, stratigraphic evidence from Gondwana indicates that small-scale glaciation did continue in places like eastern Australia and Africa through the Middle Permian (Isbell et al. 2003; Fielding et al. 2008a), suggesting an asynchronous deglaciation at the end of the LPIA. It is difficult to ascertain why there is not global synchronicity with respect to these post-Sakmarian transgressions (i.e., deglaciations).

Implications

Despite remaining difficulties in absolute age control, a clear temporal pattern has emerged from this work. Carbonate platforms and ramps across Pangea record subaerial exposure and/or a basinward shift in facies during the PPT, which is interpreted as a sequence boundary. Eustatic patterns ultimately reflect a combination of global tectonic influences and fluctuations in continental ice volumes. In the late Paleozoic, assembly of Pangea forced changes in mid-ocean ridge volume, subduction zone mechanics, regional subsidence rates, and continental collisions (see additional discussions in Golonka and Ford 2000 and Blakey 2008). However, eustatic changes related to these mechanisms operate on 1st-3rd-order time scales (millions of years), whereas glacioeustatic mechanisms work on much shorter time scales (4th and 5th-order cycles; see Coe 2003). Both tectonic and glacioeustatic mechanisms can lead to subaerial exposure and the development of unconformities. However, only large-scale increases in continental ice
volume work at a scale that would result in the creation of an abrupt and globally synchronous unconformity. If the mechanisms work in concert, rapid eustatic fall (glacioeustasy) combined with slow continental emergence (tectonic) would act together to create a globally synchronous sequence boundary.

The Late Pennsylvanian-Early Permian collision between Gondwana and Euramerica is well documented (e.g., Scotese and Langford 1995; Golonka and Ford 2000; Blakey 2008). As such, it is likely that some of the eustatic drop during the PPT can be explained by tectonic uplift across portions of Pangea, albeit on much slower time scales than glacioeustasy. This is the case for the Permian Basin where regional uplift within the Marathon orogenic belt created a major hiatus and an angular unconformity in uppermost Pennsylvanian strata (Ross 1986; Ross and Ross 1986; Hill 1996). Additionally, slow continental emergence related to thermal tectonic uplift and cratonic warping may also help to explain the gradual shallowing of facies inferred for the U.S. Midcontinent and Russian sections (West et al. 1997; Watney et al. 2006).

Perhaps the most tectonically stable region studied is the South China Platform. During the PPT, the South China Platform was not attached to Pangea; it lay on the tectonically passive side of the Yangtze Craton (Enos 1995; Scotese and Langford 1995; Golonka and Ford 2000; Wang and Jin 2000). Even in this tectonically stable environment, the stratigraphy records a significant drop in relative sea level during the PPT (Yang et al. 1986; Meyerhoff et al. 1991; Shi and Chen 2006). This evidence implies that the global stratigraphic pattern documented herein cannot be explained by a tectonic mechanism alone.
The global stratigraphic pattern of a eustatic drop documented in this study most likely resulted from a eustatic fall caused by the interplay between higher frequency (3rd-5th-order) glacioeustatic signals superimposed upon long-term global tectonic influences (1st-3rd-order cycles) coinciding with the beginning of increased ice accumulations across much of Gondwana (Isbell et al. 2003; Fielding et al. 2008b). The combination of major tectonic events (Pangea assembly) and glacioeustatic fluctuations (LPIA) helps explain why the lowest sea levels during the Paleozoic were during the Permian. Since areas of relative tectonic stability (i.e., South China Platform) were also influenced by a major drop in relative sea level during the PPT, it is likely that the ice sheets on Gondwana had a more significant role in controlling eustasy during the PPT.

Given all of the available data, it is hypothesized that variation in Gondwanan ice volume was the dominant control over eustatic change during the latest Pennsylvanian-Sakmarian. Stratigraphic data from carbonate environments across the paleotropics and subtropics is consistent with a peak of the LPIA occurring in the very latest Pennsylvanian-Early Permian. Low-latitude carbonate platforms and ramps also record the deglaciation of Gondwana as a series of large-scale asynchronous transgressions beginning in the mid-Sakmarian.

**CONCLUSIONS**

Stratigraphic data from carbonate platforms and ramps spanning equatorial Pangea contain a sequence boundary or a distinct basinward shift in facies that suggest a major eustatic drop in the latest Pennsylvanian-earliest Permian. This event is present in some form in all eight regions studied for this review, regardless of degree of regional
tectonic activity (Stemmerik and Worsley 1995; Hill 1996; West et al. 1997; Vennin et al.
2002; Wahlman and King 2002; Hanks et al. 2006; Shi and Chen 2006; Grader et al.
2008). Results from this study suggest a common overall cause for the formation of this
global stratigraphic pattern, which is consistent with a glacioeustatic drawdown
beginning in the earliest Permian.

This study supports emerging evidence that global ice volume reached an acme in
the Early Permian (Isbell et al. 2003; Fielding et al. 2008b-c). A significant increase in
ice volume across Gondwana during the PPT resulted in widespread exposure events or a
significant basinward shift in facies in low-latitude carbonate environments. The eustatic
drop during the PPT was likely enhanced by the tectonically driven eustatic fall
associated with the final assembly of Pangea, which operated on 1st-3rd-order time scales
(West et al. 1997; Watney et al. 2006).

Mid-Sakmarian-Kungurian transgression events documented here suggest an end
to major Gondwanan glaciation by the mid-Sakmarian. However, the persistence of
small ice centers through the Middle Permian suggests an asynchronous deglaciation of
Gondwana after the Sakmarian (Isbell et al. 2003; Fielding et al. 2008a-c). Further
research is needed to document the global nature of these complex post-Sakmarian
transgressions and how they relate to the overall demise of the LPIA.

It is intriguing that the inferred eustatic drop during the PPT manifests itself
differently in stratigraphic successions across the globe. In many of the records, the
eustatic lowstand is represented by a large unconformity. Other records show only a
basinward shift in facies during the Asselian, instead of a major exposure surface at the
Pennsylvanian-Permian boundary. This study highlights the complicated stratigraphic
patterns that can form during periods of increased glaciation, combined with large-scale
tectonic processes associated with supercontinent assembly.

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APPENDIX

Biostratigraphic Framework

Conodonts.—Perhaps the best index fossils for correlation of upper Paleozoic
strata are conodonts due to their cosmopolitan nature and rapid evolutionary rates during
the late Paleozoic (Wardlaw, 1995). The Pennsylvanian-Permian boundary discussed
herein is based on the Global Stratotype Section and Point (GSSP) for the base of the
Permian System, which was placed at a shallow marine shelf section, southeast of the
Russian Platform in Kazakhstan (Davydov et al. 1998). This boundary corresponds to
the basal occurrence of the conodont Streptognathodus isolatus, which is a morphotype
of Streptognathodus “wabaunsensis” (see Fig. 3, Davydov et al. 1998). In the U.S.
Midcontinent, this horizon occurs in the basal portion of the Bennett Shale Member of
the Red Eagle Limestone (Ritter, 1995). This conodont boundary also has been identified
in the Glass Mountains portion of the Permian Basin (Wardlaw and Davydov 2000), and in southern China (Wang 2000).

Utilizing conodont zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al. 2004 and references therein; see Fig. 3): the base of the Sakmarian is defined by the basal occurrence of *Sweetognathus merrilli*; the base of the Artinskian is defined by the first appearance of *Sweetognathus whitei*; and the base of the Kungurian is within the *Neostreptognathodus* zone (*N. pnevi, N. exculptus, N. pequopensis*)..

**Fusulinids.**—Fusulinids represent a widely used biostratigraphic tool for upper Paleozoic sections across the paleotropical and mid-latitude localities (Ross 1995; Ross and Ross 1995), but their biostratigraphic usefulness is limited by provincialism and homeomorphy with respect to evolution (Davydov 1996). Despite these issues, fusulinids continue to be utilized as biostratigraphic indicators for upper Paleozoic studies around the world. At the GSSP, the *Sphaeroschwagerina vulgaris*-*Sphaeroschwagerina fusiformis* zone occurs 6 m above the conodont marker, *Streptognathodus isolatus*. In many regions such as the Orogrande Basin, Barents Shelf, Arctic North America, and Bolivia, conodont data are unavailable and fusulinids are relied upon for biostratigraphic correlations (Dunbar and Newell 1946; Newell et al. 1953; Ross and Ross 1987; Ross 1995; Stemmerik 2000; Wahlman and King 2002). In the Boreal and Tethyan regions, the *Sphaeroschwagerina vulgaris*-*Sphaeroschwagerina fusiformis* zones provides the closest approximation for the base of the Permian (Fig. 3; Davydov 1996). In North America, the base of the Permian is defined by first occurrence
of the Nealian fusulinid fauna, which is represented by: *Pseudoschwagerina uddeni*, *P. texana*, *P. beedei*, *Paraschwagerina gigantea*, larger *Leptotriticites* and advanced *Schwagerina* (Fig. 3; Ross 1963; Wahlman and King 2002).

Utilizing fusulinid zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al. 2004 and references therein; see Fig. 3): the base of the Sakmarian is defined by the basal occurrence of *Schwagerina moelleri*; the base of the Artinskian is defined by the first appearance of the *Pseudofusulina* genus; and the base of the Kungurian is at the base of the *Brevaxina* zone.

**Ammonoids.**—Ammonoids have also been utilized as biostratigraphic indicators for upper Paleozoic strata, but as with fusulinids, they show a significant amount of provincialism, thus limiting their use for high-resolution biostratigraphy (Gradstein et al. 2004 and references therein). At the GSSP, the *Svetlanoceras* ammonoid zone appears ~27 m above the first occurrence of the conodont marker, *Streptognathodus isolatus* (Davydov et al. 1998). Although less reliable, the boundary between the *Svetlanoceras* and *Shumardites* ammonoid zones marks the Carboniferous-Permian boundary (Fig. 3; Gradstein et al. 2004 and references therein). Utilizing ammonoid zones, the remaining Lower Permian strata are divided as follows (after Gradstein et al. 2004 and references therein; see Fig. 3): the base of the Sakmarian is within the upper part of the *Svetlanoceras* genus; the base of the Artinskian is defined by the first appearance of the *Neopronorites-Metaperrinites* zone; and the base of the Kungurian is within the *Aktubinska-Artinska-Neopronorites* zone.
**Additional Stratigraphic Information**

**United States Midcontinent.---**The Permian System of the Midcontinent extends from Oklahoma to Iowa (including Kansas, Missouri, and Nebraska; it encompasses several basins including the Salina, Forest City, and Sedgwick (Heckel 1986; Mazzullo 1998; 1999; Fig. 1; Table 1). Paleomagnetic data suggest these basins were located between 0-10° N during the PPT (Ross and Ross 1990; Witzke 1990; Scotese and Langford 1995; Golonka and Ford 2000). The tectonic stability during the late Paleozoic and lateral continuity of strata across this broad epeiric platform have allowed for the construction of a robust regional stratigraphic framework (Mazzullo 1999; Olszewski and Patzkowsky 2003; Heckel 2008). Age control for the upper Paleozoic strata of the Midcontinent is achieved through the use of conodont and fusulinid biostratigraphy (Thompson 1954; Ross and Ross 1988; 1995; Ritter 1995; Fig. 3). The GSSP conodont marker, *Streptognathodus isolatus*, has been observed in the base of the Bennett Shale (Ritter 1995; Chernykh and Ritter 1997) and the Nealian fusulinid fauna (described above) occurs in the Neva Limestone (Thompson 1954).

Upper Paleozoic strata of the Midcontinent are known for containing cycles of mixed clastic/carbonate deposits. In general, sea level highstands are represented by offshore dark gray/black shales, and sea level lowstands are recorded by shallow marine carbonates, delta deposits, and/or paleosols (Heckel 1986; 2008). Individual cycles are traceable across broad regions with relative ease due to a robust biostratigraphic framework and limited tectonic modification.

Several studies have suggested that 3rd and 4th-order cycles dominate in Upper Pennsylvanian/Lower Permian strata (Mazzullo 1998; Mazzullo 1999; Olszewski and
Patzkowsky 2003; Heckel 2008). For PPT strata, high resolution sequence stratigraphic analysis of outcrop data has suggested 5th-order cycles that are superimposed on lower order cycles (Olszewski and Patzkowsky 2003). However, the style of cyclicity across the Pennsylvanian-Permian boundary remains constant. Although cyclicity appears to have not changed significantly during the PPT, a shift from humid conditions in the Late Pennsylvanian to drier conditions in the Permian has been inferred based on stable isotope compositions of marine carbonates and a shift to more evaporite rich lithofacies (West et al. 1997; Mazzullo et al. 2007). This shift to drier conditions during the Permian is consistent with other reports demonstrating a shift to more arid conditions across western Pangea in the Permian (Tabor and Montañez 2002; Tabor et al. 2008).

**Orogrande Basin, New Mexico, USA.**---The Orogrande Basin of south-central New Mexico contains some of the most continuous Pennsylvanian-Permian sections of marine carbonates in western Euramerica (Robledo Mountains; Wahlman and King 2002). The basin was located within 5° of the equator during the PPT (Ross and Ross 1990; Scotese and Langford 1995; Golonka and Ford 2000; Fig. 1; Table 1). During Pennsylvanian-Permian time, the Orogrande Basin was a westward extension of the Permian Basin in Texas; it was relatively shallow and elongate, trending generally north to south (Jordan 1975; Candelaria 1988). Deposits range from neritic marine carbonate strata in the south to mixed siliciclastic-carbonate strata in the northern portion of the basin (Jordan 1975; Candelaria 1988; Rankey et al. 1999). The nomenclature of Upper Pennsylvanian mixed clastic-carbonate strata in the basin is complex and varies with geographic location. In the northern San Andres and Franklin Mountains (northern and
southern part of the basin, respectively), Upper Pennsylvanian strata are the Panther Seep and Bursum Formations; coeval strata in the Sacramento Mountains (eastern part of the basin) are the Beeman, Holder, and Laborcita Formations (Kues 2001; Raatz 2002). The term Madera Group has been proposed for Upper Pennsylvanian strata in the western part of the basin, such as the Robledo Mountains (Kues 2001).

Lower Permian strata consist of open marine carbonate-dominated Hueco Group in the southern part of the basin and the terrestrial to marginal marine siliciclastic-dominated Abo Formation in northern areas (Jordan 1975; Candelaria 1988; Raatz 2002; Wahlman and King 2002; Mack et al. 2003). A transitional zone occurs in the central portion of the basin where Hueco and Abo strata interfinger. In this part of the basin, the marine carbonates of the Hueco Group show characteristics consistent with a more restricted environment (impoverished faunas, etc.) and the siliciclastic strata of the Abo Formation show a more marine influence (e.g., wave ripples, tidal flat facies, etc.) than do lithofacies further to the north (Jordan 1975; Wahlman and King 2002; Mack et al. 2003). Since conodont data are unavailable for the Orogrande Basin, PPT deposits have been dated through the use of fusulinid biostratigraphy; the Nealian fusulinid fauna, which represents lowermost Permian, has been found throughout the basin (Williams 1966; Steiner and Williams 1968; Wahlman and King 2002).

Depositional style changed in the Orogrande Basin across the PPT, partially due to a decrease in basin subsidence related to regional tectonic activity (Candelaria 1988). The Pennsylvanian strata tend to be dominated by high-frequency (3rd and 4th-order) cyclic deposits, especially in the mixed carbonate-clastic successions in the Sacramento Mountains (Holder and Laborcita Formations). Individual cycles (1-3 meters thick)
consist of open marine limestone overlain by terrestrial to marginal marine clastic deposits (Rankey et al. 1999; Raatz 2002). While 3rd-5th-order cyclic deposits occur in Permian strata, they tend to be less uniform in thickness and do not form classic cyclothems as observed in the underlying Pennsylvanian units (Jordan 1975; Rankey et al. 1999; Mack et al. 2003). In the southern part of the basin they consist of shallowing upward sequences in open marine carbonate strata. In the central and northern part of the basin these cycles consist of open to restricted marine carbonates overlain by terrestrial to marginal marine clastic deposits (Jordan 1975; Mack et al. 2003).

**Permian Basin, Texas, USA.**---The Permian Basin in western Texas and southeast New Mexico comprises several subbasins (e.g., Delaware, Midland, etc.) that existed as a segmented foreland basin during the late Paleozoic. These relatively deep basins contain some of the most complete Permian sections in northern Pangea (Fitchen et al. 1995; Mazzullo 1995; Hill 1996). During the PPT, the region was located within 5° of the paleoequator (Ross and Ross 1990; Scotese and Langford 1995; Golonka and Ford 2000; Fig. 1; Table 1). Strata contain mixed clastic-carbonate facies, which resulted from the collision of the North and South American plates during the late Paleozoic (Ross 1986; Fitchen et al. 1995). The best outcrops of PPT strata occur in the Glass Mountains region of the Delaware Basin, where the Upper Pennsylvanian Gaptank Formation consists of open marine carbonates and is unconformably overlain by Permian (Wolfcampian) strata (Hill 1996). Lower Permian strata in the Glass Mountains consist of open marine carbonate shelf facies that are divided into the Neal Ranch Formation, which is unconformably overlain by the Lenox Hills Formation (Ross 1963; Hill 1996).
Outside of the Glass Mountains the uppermost Pennsylvanian mixed clastic-carbonate strata are called the Cisco Formation (Group), and the lowermost Permian mixed clastic-carbonate strata are assigned to either the Wolfcamp Formation or Hueco Group (Fitchen et al. 1995; Mazzullo 1995; Hill 1996; Yang et al. 1998). In the Glass Mountains, the conodont genus *Streptognathodus* has been found in the Grey Limestone Member, which places the Pennsylvanian-Permian boundary near the top of the Gaptank Formation (Wardlaw and Davydov 2000). Other key conodont markers such as, *Sweetognathus merrilli*, *Sweetognathus whitei*, and *Neostreptognathodus pequopensis*, define Sakmarian, Artinskian, and Kungurian deposits, respectively (Fig. 3). The fusulinid biostratigraphy also supports the conodont data, with the Nealian fusulinid fauna present in Lower Permian strata (Ross 1963; 1995; Fig. 3). However, the biostratigraphic resolution of the fusulinid data in the Glass Mountains is not as highly resolved as the conodont data (Wardlaw and Davydov 2000).

There is evidence of multiple scales of depositional cyclicity throughout the basin (3rd and 4th-order, etc.) in both Pennsylvanian and Permian strata, which are recorded as alternating beds of shallow-water carbonates and terrestrial to basinal clastics (Mazzullo 1995; Yang et al. 1998; Saller et al. 1999; Yang and Kominz 1999). Although subtle changes in facies and cycle thickness are evident, significant changes in depositional style do not appear to be present across the PPT.

**Arctic North America.**---Northern Alaska moved from 30-35° N during the Pennsylvanian to 30-40° N during the Permian (Scotese and McKerrow 1990; Scotese and Langford 1995; Golonka and Ford 2000; Bensing et al. 2008; Fig. 1; Table 1).
Rifting dissected Arctic Alaska and the North Alaska Platform developed on horst-structures in a shallow-water marine environment (Beauchamp 1995). The North Alaska Platform was chosen over the adjacent Sverdrup Basin because tectonic rifting was less developed in this region (Beauchamp 1995). Upper Pennsylvanian strata consist of the Wahoo Limestone (upper Lisburne Group), which is overlain by mixed carbonate-clastic strata of the Permian-age Echooka Formation/lower Sadlerochit Group (Bamber and Waterhouse 1971; Crowder 1990; Beauchamp 1995; Hanks et al. 2006). The uppermost Mississippian-Upper Pennsylvanian Wahoo Limestone was deposited under open marine conditions on a shallow carbonate ramp and contains a diverse fauna of brachiopods, fusulinids, conodonts, foraminifers, and corals (Bamber and Waterhouse 1971; Crowder 1990; Hanks et al. 2006). The overlying Echooka Formation (informal basal conglomerate member and the Joe Creek Member) contains a mix of conglomerate, restricted carbonates, and nearshore clastic strata (Bamber and Waterhouse 1971; Crowder 1990; Beauchamp 1995; Hanks et al. 2006). Upper Paleozoic strata from the North Alaska Platform are dated primarily through the use of fusulinids (Ross 1995; Fig. 3). Only Carboniferous-Early Permian fusulinids are known from northern Alaska, which is attributed to the tectonic drift into more northerly latitudes during the mid-Permian (Beauchamp 1995; Ross 1995). Although Permian strata are predominantly clastics (Echooka Formation), coeval open marine carbonate strata in the adjacent Sverdrup Basin contain key Permian fusulinid markers such as: *Sphaeroschwagerina vulgaris-fusiformis* (Pennsylvanian-Permian boundary); *Schwagerina moelleri* for the base of the Sakmarian; *Pseudofusulina* genus for the base of the Artinskian; and the *Parafusulina* genus for upper Artinskian-lower Kungurian strata (Ross 1995; see Fig. 3).
Cyclic deposits (3rd and 4th-order) are known, consisting of typical shoaling upward cycles that dominate the Wahoo Formation (Bamber and Waterhouse 1971). Fining upward cycles, interpreted as tempestites, are common in the Echooka Formation (Crowder 1990). Even though cyclic deposits have been documented, there has been only a broad attempt to relate these changes to global paleoclimatic and eustatic variations. Beauchamp and Baud (2002) determined that high-frequency depositional cyclicity ended in the latest Sakmarian, which coincides with a regional maximum flooding event.

**Bolivia.---** The Bolivia/Peru region of South America was situated at ~30° S during the Mississippian and Early Pennsylvanian (Golonka and Ford 2000; Fig. 1; Table 1). Northward movement into more tropical latitudes (20-25° S) during the Late Pennsylvanian-Early Permian allowed for the development of extensive warm-water carbonate platforms (Isaacson and Díaz-Martínez 1995; Sempere 1995). This region is different from the other localities discussed in this paper because Bolivia was part of Gondwana, and more proximal to the extensive glacial activity that occurred further to the south. Upper Pennsylvanian-Lower Permian deposits in Bolivia are represented by the Copacabana Formation, which consists of mostly shallow-water carbonates interbedded with terrigenous clastic units (Grader et al. 2008). The Copacabana Formation occurs in a series of basins and subbasins throughout Bolivia and parts of central and southern Peru (Newell et al. 1953; Grader et al. 2008). The mixed clastic-carbonate-volcanic Chutani Formation unconformably overlies the Copacabana Formation throughout most of the region.
Although conodont studies exist (e.g., Suárez-Riglos et al. 1987), fusulinid biostratigraphy remains the most widespread taxa used for age determinations of Bolivian PPT strata (Dunbar and Newell 1946; Newell et al. 1953; Ottone et al. 1998; Fig. 3). The fusulinid faunas described from Bolivia (Nealian fusulinid fauna) can be correlated with those from the southwestern United States (Ross 1963; Wahlman and King 2002) because they contain similar taxa (Dunbar and Newell 1946; Newell et al. 1953). Corals and bryozoans also have been used for age control in the region (Wilson 1990; Sakagami 1995).

A major sequence boundary occurs at the Pennsylvanian-Permian boundary within the Copacabana Formation (299 Ma; see Fig. 2). Grader et al. (2008) correlated this boundary with the beginning of the P1 glacial epoch in eastern Australia, of Fielding et al. (2008a). Higher up in the succession, a regionally extensive latest Sakmarian-early Artinskian flooding surface represents a period of high relative sea level (Grader et al. 2008).

The Copacabana Formation contains high-frequency cycles throughout, which have been interpreted as 2nd, 3rd, and 4th-order sequences (Grader et al. 2008). Individual cycles consist of lowstand and transgressive clastic deposits that are capped by highstand open marine carbonate strata. The tops of cycles often show signs of exposure (Isaacson and Díaz-Martínez 1995). As in the Barents Shelf and North Alaska regions, the higher-frequency cycles disappear from the record after the late Sakmarian (Beauchamp and Baud 2002; Grader et al. 2008; Stemmerik 2008).

The formation of the erosional unconformity across the PPT is consistent with a eustatic drop in the earliest Permian. Additionally, the formation of this sequence
boundary does not appear to be related to regional tectonic uplift, which further supports a eustatic control (Grader et al. 2008).

**Southern China.---**During the late Paleozoic, the Yangtze Craton (i.e., South China block) was situated between 0-15° S in the Paleo-Tethys Ocean (Nie et al. 1990; Scotese and Langford 1995; Golonka and Ford 2000; Fig. 1; Table 1). Southern China, a broad and relatively flat terrain during the late Paleozoic, did not collide with northern China (Sino-Korean Craton) until the Mesozoic (Enos 1995; Wang and Jin 2000). This paleogeographic configuration allowed epicontinental seas to periodically cover southern China, leading to the development of epeiric platforms in the Dian-Qian-Gui Basin during portions of the Pennsylvanian-Permian (Wang and Jin 2000). PPT strata in the region consist of the Chuanshan Formation, an oncoid-bearing carbonate unit which is disconformably overlain by the late Cisuralian carbonate-dominated Qixia Formation (Wang and Jin 2000; Shi and Chen 2006).

Upper Paleozoic strata from the Paleo-Tethys region are dated using conodont and fusulinid biostratigraphy (Yang 1986; Meyerhoff et al. 1991; Enos 1995; Wang 2000; Shi and Chen 2006; Fig. 3). The key conodont species, *Streptognathodus isolatus*, which defines the base of the Permian System, is widely distributed throughout southern China (Wang 2000). This conodont marker coincides with a key fusulinid zone (*Sphaeroschwagerina vulgaris-Sphaeroschwagerina fusiformis*), which also aids in Pennsylvanian-Permian boundary correlations (Ross 1995). Other faunas such as, brachiopods, ammonoids, and corals provide useful age constraints in many places (Meyerhoff et al. 1991).
Whereas depositional cyclicity has been observed in many sections across the South China Platform, there have been few attempts to correlate individual cycles and sequences to other global stratigraphic frameworks (e.g., Midcontinent, Russian Platform, etc.). Shi and Chen (2006) recognized high-frequency (3rd and 4th-order?) cycles in sedimentary stacking patterns of the Chuanshan Formation and related these cycles to glacioeustatic fluctuations.

**Russian Platform.**—During the Late Pennsylvanian, most of eastern Russia was situated between 15-25° N in a tropical to subtropical environment (Golonka and Ford 2000; Fig. 1; Table 1). The region moved north during the Early Permian to 20-30° N (Scotese and McKerrow 1990; Scotese and Langford 1995). The Late Pennsylvanian-Early Permian Russian Platform was dominated by warm-water carbonate deposition; phylloid algal mound development reached a peak in the Asselian-Sakmarian (Vennin et al. 2002; Vennin 2007). The Russian nomenclature system utilizes a set of horizons instead of formations to differentiate lithostratigraphic units. Horizon determinations are based on lithostratigraphic changes as well as fusulinid/conodont assemblages. Gzhelian deposits across the Russian Platform are generally separated into the Rusavkiany, Pavlovoposadian, Noginian, and Melekhovian Horizons (Ross and Ross 1985; Davydov and Leven 2003). Asselian strata are separated into the Sjuranian, Uskalykian, and Shikhanian Horizons, and Sakmarian deposits are divided into the Tastubian and Sterlitamakian Horizons (Chuvashov 1995; Vennin et al. 2002). Artinskian strata are divided into the Burtsevian, Irginian, and Sarginian Horizons, and Kungurian strata defined by the Saranian, Fillipovian, and Irenian Horizons (Chuvashov 1995).
The chronostratigraphic framework is extremely well constrained across the Russian Platform. Although conodonts are used to identify the official Pennsylvanian-Permian boundary (i.e., first appearance of *Streptognathodus isolatus*), fusulinid and ammonoid biostratigraphy remain extremely useful (Davydov et al. 1998; Izart et al. 1999; Vennin et al. 2002; Davydov and Leven 2003; Fig. 3). Additionally, volcanic tuffs interbedded with conodont-rich facies have allowed for U-Pb dating of GSSP-equivalent strata, which yielded an absolute age of 298.90 ±0.31 Ma (Ramezani et al. 2007).

Cyclic deposits are very common in carbonate strata of the Russian Platform. Many of these cycles have been interpreted as 3rd and 4th-order (high-frequency) in nature (Izart et al. 1999). Many are bounded by subaerial unconformities, including a widespread exposure event at the Sakmarian-Artinskian boundary (Ross and Ross 1985; Vennin et al. 2002; Vennin 2007; Fig. 2). Additionally, documented deeper water marl deposits overlying shallow water carbonate reef facies in upper Artinskian strata have been interpreted as a regional transgression and platform drowning (Vennin et al. 2002).

**Barents Shelf (Finnmark Platform); Svalbard.** The present-day Barents Shelf region was situated 25-30° N during most of the Pennsylvanian (Golonka and Ford 2000; Stemmerik 2000; Fig. 1; Table 1). The region moved progressively northward to 35-45° N during the Late Pennsylvanian and Early Permian (Scotese and McKerrow 1990; Scotese and Langford 1995; Golonka and Ford 2000; Stemmerik 2000). This northward shift led to dramatic climate change in the region, from subtropical warm and dry during the Gzhelian-early Sakmarian to cool temperate conditions during the late Sakmarian (Stemmerik 2000; Rafaelsen et al. 2008; Stemmerik 2008). On the Finnmark Platform,
photozoan-dominated warm-water carbonates of the upper Gipsdalen Group (Ørn Formation) were deposited during the Gzhelian through mid-Sakmarian (Stemmerik 2000; Rafaelsen et al. 2008; Stemmerik 2008). The Gipsdalen Group is overlain by heterozoan-dominated cool-water carbonates of the Bjarmeland Group (Isbjørn and Polarrev Formations; see Rafaelsen et al. 2008). Kungurian through Wuchiapingian strata are deeper water carbonates, including the chert-dominated Tempelfjorden Group (Røye Formation; see Rafaelsen et al. 2008).

The upper Paleozoic stratigraphy on Svalbard is similar to that of the Finnmark Platform. The Upper Pennsylvanian-Lower Permian Ørn Formation on the Finnmark Platform is coeval with the Wordiekammen Formation and Finnlayfjellet Members of the Gipsdalen Group (Harland and Geddes 1997; Rafaelsen et al. 2008; Stemmerik 2008). The Bjarmeland Group from the Finnmark Platform is not recognized, partially due to an erosional unconformity that removed much of the Artinskian strata (Stemmerik 1995; Rafaelsen et al. 2008). Lower Bjarmeland Group equivalent strata on Svalbard are represented by the Gipshuken Formation of the upper Gipsdalen Group (Rafaelsen et al. 2008; Stemmerik 2008). On Svalbard the Gipsdalen Group is unconformably overlain by the Kapp Starostin Formation of the Tempelfjorden Group (Heafford, 1988; Stemmerik and Worsley 1995; Rafaelsen et al. 2008).

Age constraints for Pennsylvanian-Artinskian strata of the Barents Shelf region come primarily from fusulinid and conodont biostratigraphy (Nakrem 1991; Harland and Geddes 1997; Nilsson and Davydov 1997; Anisimov et al. 1998; Stemmerik 2000 and references therein; Fig. 3). Cooler conditions during the late Early Permian were not conducive to fusulinid communities, and the record disappears by the late Artinskian. As
such, ages for upper Artinskian and younger strata are based on available conodont biostratigraphy (Nakrem 1991). Lowermost Permian strata contain the fusulinid, *Sphaeroschwagerina vulgaris*, and the conodont, *Streptognathodus constrictus*, which suggests an early Asselian age (Nakrem 1991; Nilsson and Davydov 1998; Stemmerik 2000 and references therein; Figs. 2 and 3). The fusulinids *Schwagerina moelleri* and *Eoparafusulina paralinearis* have been found in the upper Kapp Duner Formation (upper Wordiekammen), suggesting a Sakmarian age (Nilsson and Davydov 1998; Stemmerik 2000). The fusulinid, *Parafusulina jenkinsi*, and the conodonts *Neostreptognathodus pequopensis* and *Sweetognathus whitei*, are known from the Hambergfjellet Formation (Gipshuken Formation), which suggests an Artinskian age (Nakrem 1991; Stemmerik 2000 and references therein; Figs. 2 and 3). The Tempelfjorden Group (Fig. 2) contains the conodont, *Neostreptognathodus idahoensis*, suggesting a Kungurian age (Nakrem 1991).

Cyclic deposits are common and 3rd, 4th, and 5th-order cycles have been inferred in both Upper Pennsylvanian and Lower Permian strata (Stemmerik 2008). Shoaling upward cycles of warm-water carbonates are often capped by exposure surfaces. There does not appear to be a major change in depositional style across the Pennsylvanian-Permian boundary. However, a major change in sedimentation occurs in the mid-Sakmarian, where a significant flooding event and loss of depositional cyclicity occurs (Stemmerik and Worsley 1995; Stemmerik 2008; Fig. 2).
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CHAPTER 5: CONCLUSIONS

Large-scale climatic change during the late Paleozoic left a stratigraphic imprint in the Orogrande Basin. Yet, it is clear from this work that documenting the signal of Gondwanan glaciation in the paleotropics is not as simple as constructing a relative sea level curve. The complex interplay between glacioeustatic variations and regional tectonic change can often blend together in the stratigraphic record, which makes it difficult to distinguish between these controlling mechanisms. However, the combination of a sequence stratigraphic and stable isotopic approach utilized in this study has allowed for large-scale comparisons between the Orogrande Basin and the Gondwanan glacial record during the Early Permian.

Sequence stratigraphic interpretations from the Orogrande Basin suggest relative sea-level changes were largely controlled by glacioeustatic mechanisms because depositional patterns are not consistent with the tectonic history of the region (Jordan, 1975; Candelaria, 1988; Raatz, 2002). For example, two transgression events in the mid-Sakmarian and mid-Artinskian occurred during intervals of decreased subsidence in the region (Jordan, 1975; Candelaria, 1988; see Chapter 2, Fig. 14). Because the timing of these transgression events is roughly coeval with the end of glacial epochs P1 and P2 in eastern Australia (Fielding et al., 2008a), these transgressions are interpreted to represent eustatic rise resulting from large-scale deglaciation events.

The distribution of $\delta^{13}C$ and $\delta^{18}O$ values, petrographic, and stratigraphic data for Gzhelian through the lower Sakmarian strata suggest episodic subaerial exposure resulting in low $\delta^{13}C$ values, which likely reflects the incorporation of soil-derived CO$_2$ into the carbonate system (Lohmann, 1988; Patterson and Walter, 1994; Theiling et al., 2007). These multiple subaerial exposure events during the Gzhelian through early
Sakmarian are interpreted to have been controlled by high-frequency sea-level change associated with glacioeustatic fluctuation (Chapter 3). This interpretation is consistent with recently compiled records of late Paleozoic eustasy, which suggest the occurrence of major glacioeustatic fluctuations (up to 120 m) until the mid-Sakmarian (Rygel et al., 2008).

By contrast, the distribution of $\delta^{13}$C and $\delta^{18}$O values, petrographic, and stratigraphic data for mid-Sakmarian through Kungurian strata suggest large-scale subaerial exposure did not occur during this time interval. Additionally, $\delta^{13}$C values return to the range of global values compiled by Korte et al. (2005), Frank et al. (2008), and Grossman et al. (2008). The lack of subaerial exposure events during this time is interpreted to represent smaller magnitudes of eustatic change. Smaller eustatic change after the mid-Sakmarian is consistent with trends documented in global eustatic studies (Rygel et al., 2008) and Gondwanan stratigraphic records suggesting an end to major glaciation during the mid-Sakmarian (Fielding et al., 2008b-c).

Smaller eustatic fluctuations after the mid-Sakmarian, documented globally (Rygel et al., 2008) and inferred for the Orogrande Basin (Chapter 3), suggest that younger glacial epochs on Gondwana (P2 and P3 in eastern Australia) did not create enough glacioeustatic change to expose carbonate strata of the Orogrande Basin for long intervals of time from the mid-Sakmarian through the Kungurian. This interpretation is not consistent with stratigraphic and isotopic evidence from eastern Australia, which suggests the P2 glacial epoch was nearly as extensive as the preceding P1 (Fielding et al., 2008a; Birgenheier et al., 2010). One potential explanation of this discrepancy is that the P2 glacial epoch was as extensive as the P1 in eastern Australia, but overall ice volume
across Gondwana was lower after the mid-Sakmarian. Lower overall ice volumes after
the mid-Sakmarian appears to be consistent with Gondwanan stratigraphic records
compiled by Fielding et al. (2008b).

Considering a global perspective, an examination of other well constrained
stratigraphic records spanning the Pennsylvanian-Permian boundary suggest that low-
latitude carbonate platforms underwent significant exposure during the Asselian (see
Chapter 4; Stemmerik and Worsley 1995; Hill 1996; West et al. 1997; Vennin et al. 2002;
Wahlman and King 2002; Hanks et al. 2006; Shi and Chen 2006; Grader et al. 2008).
This stratigraphic pattern suggests a major eustatic drop, which is consistent with a
glacioeustatic drawdown beginning in the earliest Permian. These stratigraphic
interpretations support the notion of a glacial acme for the late Paleozoic ice age during
the Early Permian (Isbell et al. 2003; Fielding et al. 2008b-c). In some of the regions
(Permian Basin, Texas), the eustatic drop across the Pennsylvanian-Permian boundary
was likely enhanced by tectonically driven eustatic fall associated with the final assembly
of Pangea (West et al. 1997; Watney et al. 2006). Additionally, analysis of large-scale
transgression events in the low-latitudes suggests an asynchronous deglaciation of
Gondwana after the Sakmarian (see Chapter 4, Fig. 2). These transgression events need
to be studied in further detail to determine their global significance and relationship to
deglaciation events across Gondwana.

Sequence stratigraphic and stable isotopic analyses from the Orogrande Basin
(Chapters 2 and 3) are consistent with other stratigraphic successions (Chapter 4),
suggesting the preservation of a global signal. It is clear that glaciation during the late
Paleozoic left a distinct imprint within the carbonate strata of the Orogrande Basin in the
form of facies changes related to sea-level variations, exposure events, and isotopic excursions throughout the succession. These stratigraphic and isotopic changes were driven primarily by eustatic fluctuation, which was likely enhanced by large-scale climatic variations (e.g., tropical cyclones, etc.) associated with deglaciation of Gondwana and the final assembly of Pangea (Tabor and Montañez, 2002; Tabor, 2007; Tabor et al., 2008).

The results of this study demonstrate that globally significant glacial events, such as those during the Early Permian, can leave a stratigraphic and isotopic imprint in paleotropical successions. However, this glacial imprint is often muddled by local to regional-scale tectonic events. The results from the Orogrande Basin represent only a single paleotropical locality during the Early Permian. It is difficult to determine the global significance of isotopic and stratigraphic changes in the Orogrande Basin without additional Lower Permian paleotropical data sets. Paleotropical data (i.e., this study) only provide an indirect record of glaciation that only allows for general inferences to be made about the ice-proximal environment. Low-latitude data do not record information related to ice center locations or movements. More research on upper Paleozoic strata from high and low-latitudes is needed to enhance chronostratigraphic, biostratigraphic, isotopic, and sequence stratigraphic relationships, which will ultimately result in higher-resolution correlations between ice-proximal and equatorial regions. A global stratigraphic framework is necessary to help explain the complex climatological relationships between the tropics and poles during icehouse periods, such as the late Paleozoic. Future research will focus on constructing high-resolution isotopic and
sequence stratigraphic records for other upper Paleozoic successions in an effort to help resolve global stratigraphic patterns associated with late Paleozoic climate change.
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APPENDIX A: GRAPHIC COLUMNS

OF OUTCROPS AND CORES
Figure 1: List of symbols used for graphic logs throughout Appendix A.
<table>
<thead>
<tr>
<th>Thickness ft</th>
<th>Graphic Log (Lithologies, Structures, Fossils)</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-340</td>
<td></td>
<td>1a; 1b</td>
<td>94.0-103.04 ft: Blocky red mudstone as below, laminated, thin (&lt; 0.5 cm thick) sandstone units (as below) interbedded with mudstone, top 3 m is bedded sandstone as below. Continues to next page.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5b</td>
<td>92-94.04 ft: Wackestone to packstone, massive, fusulinids, brachiopods, conchostraca (sample SN 53).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1a; 1b</td>
<td>89.14-92.84 ft: Blocky red mudstone as below, same interbedded sandstone units as above.</td>
</tr>
<tr>
<td>300-400</td>
<td></td>
<td>5a; 5b</td>
<td>83.14-86.14 ft: Wackstone, massive with some winnowed horizons i.e., concentrated grains, micritic in some areas, fusulinids concentrated in some areas (samples SN 85 m and SN 85 m).</td>
</tr>
<tr>
<td>300-400</td>
<td></td>
<td>1a; 1b; 1c</td>
<td>82.84-83.14 ft: Wackstone to packstone, conglom. unit at base, grades into a fusulinid packstone with small algal biocerm present (samples SN 83 m, SN 84 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5a; 5a; 5b</td>
<td>77.64-82.84 ft: Red mudstone as below, small channel feature (77 cm thick) that has a chalk conglom. with gastropods, samples SN 81 m.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5a; 1a; 1c</td>
<td>59.71-77.64 ft: Wackestone, massive, brachiopods, conchostraca (samples SN 77 m, SN 78 m).</td>
</tr>
<tr>
<td>200-300</td>
<td></td>
<td>6a; 6b; 1c</td>
<td>74.00-78.37 ft: Nodular fusulinid packstone, some brachiopods and rugose algae, a small brachiopod bed at 75 m, sample 75 m.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>6a; 6b; 3a</td>
<td>59.71-74.00 ft: Wackstone, wavy beds, nodules, black banding, conchostraca, gastropods, bryozoans, sample SN 73 and SN 73 m.</td>
</tr>
<tr>
<td>100-200</td>
<td></td>
<td>1a; 1b</td>
<td>70.47-72.37 ft: Blocky red mudstones, laminated, interbedded sandstone 68.47-70.47 ft: Fine-grained sandstone beds (10-15 cm thick) with interbedded shale (&lt; 5 um), laminated, some sandy limestone beds within the unit containing a few scattered fossils (sample SN 68 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5a; 5a; 1d</td>
<td>68.97-66.97 ft: Wackstone, coquina pebble packstone within the unit (&lt; 1 cm thick), some shale intervals (&lt; 5 cm thick), unit grades into a massive bedded fusulinid packstone (top 1.2 m) with stylolitized contacts, bryozoans, echinoids, brachiopods, fusulinids, coarsens up-sequence, coated grains present (samples SN 65 m, SN 65 m, 67 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4</td>
<td>56.97-56.97 ft: Mostly covered with some exposures of wackestone to grainstone to buffstone, foraminifera, fusulinids, brachiopods, some silt and mud-dominated horizons, intercalation, some interbedded shale units (&lt; 5 cm thick), algal coatings present on many grains. (sample SN 24 m and SN 46.47 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5b; 4</td>
<td>11.07-18.48 ft: Calcareous siltstone with some shale interbedded, wavy beds, some lamination present, some are massive, grades into massive very silty limestone, some granule (fine-sand) present, sort of a non-descript unit.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1d</td>
<td>9.55-11.07 ft: Wackstone, some areas of packstone, possible cross-beds, less fossil material than previous unit (sample SN 6.60 m, 11.0 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5b; 4</td>
<td>6.15-8.53 ft: Granule, brachiopods, echinoids, mollusks, some marine intervals, wavy convolute beds (sample SN 6.50 m).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1d</td>
<td>7.00-11.5 m: Black shale, not tested, laminated.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1d</td>
<td>0.88-7.73 ft: Calcareous shale to siltstone, planar-laminated, fluid escape structures, unit becomes coarser up-section into a coquina-rich grainstone, laminated structure disappears up-section, few fossils (samples SN 3.75, 4.0, 5.5 m).</td>
</tr>
</tbody>
</table>

Figure 2: Graphic log from the lower Sacramento Mountains (0-100 m).
Figure 3: Graphic log from the lower Sacramento Mountains (100-200 m).
Figure 4: Graphic log from the lower Sacramento Mountains (200-290 m).
**Figure 5:** Graphic log from the Getty #5 Williard Beatty Core (0-67 m).
Figure 6: Graphic log from the Nicor DDH-2 Core (0-39 m).
<table>
<thead>
<tr>
<th>Thickness</th>
<th>Graphic Log</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ft</td>
<td>m</td>
<td></td>
<td>16.0-16.5m- Wackestone with crinoids, brachiopods, rare fusulinds; very organic-rich, black, deeper water?</td>
</tr>
<tr>
<td></td>
<td></td>
<td>15.0-16.0m- Black to dark gray fissile shale; non calcareous and no visible fossils; deeper water</td>
<td>14.52-15.15m- Wackestone to packstone; fusulinds, brachiopods, forams, bioturbation; unit dark in color.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14.1-14.65m- Brown elastic mudstone and fissile shale; non calcareous; limestone clasts; exposure?</td>
<td>13.1-14.1m- Sandy siltstone, limpet debriss, red in color; non-calcereous, exposure?</td>
</tr>
<tr>
<td></td>
<td></td>
<td>12.0-13.1m- Sandy green mudstone; non-calcereous</td>
<td>11.0-12.0m- GAP</td>
</tr>
<tr>
<td></td>
<td></td>
<td>9.71-11.53m- Wackestone/packstone bafflestone; fusulinds, Tubiphytes, coated grains; styloites, bioturbation.</td>
<td>7.71-9.11m- Wackestone/packstone; crinooids, forams fusulinds, phylloid algae; bioturbation; styloites.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5.25-7.91m- Phylloid algae wackestone-bafflestone; forams, brachiopods; Tubiphytes, rare fusulinds and crinoids; styloites, bioturbation and mottled porosity.</td>
<td>2.45-5.25m- Wackestone-packstone, abundant fusulinds, phylloid algae, crinoids, brachiopods, Tubiphytes. Styloites, bioturbation; mottled porosity.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.75-2.45m- Black shale; non calcareous; bioturbation</td>
<td>0.75-1.75m- Wackestone to packstone with abundant coral, forams, brachiopods; fusulinds appear in top 20cm of the section; styloites and shale partings.</td>
</tr>
</tbody>
</table>

Figure 7: Graphic log from the Tres Papalotes Core (0-16.5 m).
Figure 8: Graphic log from the Unocal State 1-33 Core (0-33.5 m).

<table>
<thead>
<tr>
<th>Thickness</th>
<th>Graphic Log</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ft</td>
<td>m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>0</td>
<td></td>
<td>5a; 5b; 6a; 6b; 4</td>
<td>27.6-33.5m. Wackestone to packstone with abundant fusulinids, crinoids, forams, phylloid algae, some Tubiphytes, some intrasets in lower part of unit; some calcite-filled fractures; styloites; bioturbation; capped by a grainstone with abundant gastropods, forams, brachiopods, and crinoids.</td>
</tr>
<tr>
<td>30</td>
<td></td>
<td>5a; 5b; 1a</td>
<td>26.52-27.8m. Wackestone with large intrasets (i.e., limestone breccia), fusulinids and crinoids, siltstone deposit/deepening? Unit base is top of Bough B.</td>
</tr>
<tr>
<td>60</td>
<td></td>
<td>5a; 5b; 4; 6b</td>
<td>23.7-26.52m. Phylloid algal boundstone to fusulinid packstone, fusulinids, rare corals, shell clasts, forams, crinoids; some calcite-filled fractures; algal coatings; still organo-rich (dark) bioturbation.</td>
</tr>
<tr>
<td>100</td>
<td></td>
<td>5a; 5b; 1a</td>
<td>21.58-23.7m. Wackestone to packstone (some grainstone); shale partings; crinoids, forams, phylloid algae, brachiopods, coral, fusulinids; no styloites or modic porosity; algal coatings; rock is very dark and organo-rich (oil migration?).</td>
</tr>
<tr>
<td>120</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>18.7-21.58m. GAP</td>
</tr>
<tr>
<td>140</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>13.93-16.7m. Packstone to grainstone grading into a wackestone/packstone (some grainstone); forams, phylloid algae, crinoids, brachiopods, Tubiphytes (increasing upsection); some dolomite; styloites; bioturbation; modic porosity.</td>
</tr>
<tr>
<td>160</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>10.26-13.93m. Phylloid algae bafflestone; forams, Tubiphytes common; modic porosity (decreasing upsection); styloites; some fine-grained dolomite.</td>
</tr>
<tr>
<td>180</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>8.01-10.28m. GAP</td>
</tr>
<tr>
<td>200</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>2.47-6.01m. Probable phylloid algae bafflestone; Tubiphytes and forams common. Some fine-grained dolomite; styloites; bioturbation; modic porosity.</td>
</tr>
<tr>
<td>220</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>1.72-2.42m. Packstone to grainstone capped with a wackestone/packstone; forams, Tubiphytes, forams, fusulinids; some phylloid algae, bioturbation.</td>
</tr>
<tr>
<td>240</td>
<td></td>
<td>6a; 6b; 5b</td>
<td>0.1-1.72m. Foram packstone to grainstone grading into an algal boundstone; some dolomite; phylloid algae, Tubiphytes, and crinoids common; modic porosity; styloites; bioturbation; top of Bough C is at 0.3m.</td>
</tr>
<tr>
<td>260</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>280</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>300</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>320</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>340</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Figure 8: Graphic log from the Unocal State 1-33 Core (0-33.5 m).*
**Figure 9**: Graphic log from the lower San Andres Mountains (0-90 m).
<table>
<thead>
<tr>
<th>Thickness ft</th>
<th>Graphic Log (Lithologies, Structures, Fossils)</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>340</td>
<td></td>
<td>5b; 4</td>
<td>95.88-114.88 m (see next page): Wackestone (similar to 90.89-95.89 m, but no packstone), new strike/dip N50E, 37NW, planar laminated beds, some zones of intraclasts (samples FM 98 and 103).</td>
</tr>
<tr>
<td>320</td>
<td></td>
<td>4; 5b; rare 3b</td>
<td>89.38-95.88 m: Wackestone, areas of packstone, buff to gray color, bivalves, ostracods, brachiopods, forams, some lamninated beds (&lt; 1 cm in thickness for each laminations), bioturbation (samples FM 90 and 93).</td>
</tr>
<tr>
<td>300</td>
<td></td>
<td>5b</td>
<td>85.88-89.38 m: Packstone, some wackestone, (similar to 79.88-85.88 m), fewer fossils (sample FM 87).</td>
</tr>
<tr>
<td>280</td>
<td></td>
<td>5b</td>
<td>79.88-85.88 m: Wackestone, some areas of micrite, chert nodules present (3-10 cm in diameter), bioturbation, bivalves, brachiopods, crinoids.</td>
</tr>
<tr>
<td>260</td>
<td></td>
<td>6a; 6b</td>
<td>43.88-79.88 m: Wackestone to bafflestone, some areas of micrite, but no chert present, gray color, bioturbated, brachiopods, bivalves, phylloid algae, crinoids, foram, bryozoans, ostracods, trilobites. Tubiphytes, shell debris becomes more abundant up-section, some areas are partially covered, bioherm that grades into shelf limestone (samples FM 54.38 and 62.38).</td>
</tr>
<tr>
<td>240</td>
<td></td>
<td>3a</td>
<td>40.88-43.88 m: Micrite, gray color, chert nodules present (3-5cm in diameter).</td>
</tr>
<tr>
<td>220</td>
<td></td>
<td>6a; 6b</td>
<td>10.88-40.88 m: Wackestone-bafflestone, occasional zones of micrite, gray color, shell and fossil debris, phylloid algae, Tubiphytes, crinoids, bivalves, forams, ostracods, partially covered in places, bioturbation present, appears to be a bioherm (samples FM 18.0 and 36.0).</td>
</tr>
<tr>
<td>200</td>
<td></td>
<td>5b; 3b; 1d; 2</td>
<td>1.88-10.88 m: Fissile micritic limestone with shale partings (similar to 0-1.88 m), laminations and wavy beds present, buff orange color, no fossils (sample FM 1.8).</td>
</tr>
<tr>
<td>180</td>
<td></td>
<td>5b; 3b; 1d; 2</td>
<td>0-1.88 m: Fissile, micritic limestone with shale partings (&lt; 1cm thick), some well-cemented areas lack shale, buff color, laminations present, brachiopods/bivalves, green algae, strike/dip N50E, 23NW, (sample FM 1.0) BASE OF MEASURED SECTION.</td>
</tr>
</tbody>
</table>

Figure 10: Graphic log from the Franklin Mountains (0-100 m).
Figure 11: Graphic log from the Franklin Mountains (100-200 m).
**Figure 12:** Graphic log from the Franklin Mountains (200-300 m).

<table>
<thead>
<tr>
<th>Thickness (ft/m)</th>
<th>Graphic Log (Lithologies, Structures, Fossils)</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1000</td>
<td>continues</td>
<td>5b; 1d</td>
<td>Continues on next page.</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>5a; 1d; rare 4</td>
<td>265.5-296.5 m: Wackestone to packstone as below, fusulinids, some areas are fusulinid grainstone, shale/siltstone interbedded as below, wavy bedding, phylloid algae, Tubiphytes, ostracods, bryozoans also present, (sample FM 280).</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>5a; 1d; rare 4</td>
<td>271.5-285.5 m: Wackestone to packstone (new formation?), light gray, crinoids, bivalves, fusulinids, forams, bryozoans, large Ophioceras gastropods, wavy beds with siltite and shale interbeds (1-3 cm thick), top 3 m partially covered (sample FM 272).</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>5a; 1d; rare 4</td>
<td>Covered section.</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>6a; 5b; 4</td>
<td>209.0-249.0 m: Wackestone to packstone, some areas of fusulinid/crinoid grainstone, areas of bioherm with abundant phylloid algae, no bedding features present, scattered chert nodules (e.g. 210 m) that decrease up-section, fusulinids, crinoids, forams, brachiopods, bryozoans, ostracods, Tubiphytes, gastropods, (samples FM 210, 222, 225, 237, 241.5, 244.5, and 240).</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>5b; 4</td>
<td>201.0-209.0 m: Wackestone, similar to previous unit, fusulinids, forams, crinoids, bryozoans, chert nodules as below in some areas, top 2 m forms a slope (sample FM 202).</td>
</tr>
<tr>
<td>900</td>
<td></td>
<td>5b; 4</td>
<td>Covered section.</td>
</tr>
</tbody>
</table>
Figure 13: Graphic log from the Franklin Mountains (300-400 m).
**Figure 14:** Graphic log from the Franklin Mountains (400-500 m).
### Franklin Mountains Section (Cerro Alto Fm.)

**Location (lat/long):** 31°55.69’N/106°31.35’W  
**Total Thickness Measured:** 783 m  
**Date Measured:** June 15-18, 2008  
**Measured By:** Jesse Koch and Drew Nelson

<table>
<thead>
<tr>
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<th>Graphic Log (Lithologies, Structures, Fossils)</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>ft</strong></td>
<td><strong>m</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1900</td>
<td><strong>600</strong></td>
<td></td>
<td>Continues on next page.</td>
</tr>
<tr>
<td>1920</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1940</td>
<td><strong>590</strong></td>
<td></td>
<td>5b; rare 4</td>
</tr>
<tr>
<td>1960</td>
<td></td>
<td></td>
<td>542.4-586.0 m: Covered Section, exposed float is similar to previous unit, small exposed packstone layer with crinoids, brachiopods, gastropods, forams, bivalves, and ostracods, (&lt;1 m thick) at 554 m (sample FM 554).</td>
</tr>
<tr>
<td>1980</td>
<td></td>
<td></td>
<td>5b; 4</td>
</tr>
<tr>
<td>2000</td>
<td></td>
<td></td>
<td>5a; 5b; 1d</td>
</tr>
<tr>
<td>2020</td>
<td></td>
<td></td>
<td>531.3-542.4 m: Wackestone to micrite, light gray, forms a prominent ledge, blocky appearance, occasional shale/siltstone units as below (&lt;1 cm thick), fusulinids, forams, brachiopods, gastropods, bryozoans, ostracods, (sample FM 554).</td>
</tr>
<tr>
<td>2040</td>
<td></td>
<td></td>
<td>494.3-531.3 m: Wackestone as below, some areas of micrite, brachiopods, fusulinids, encrusting forams, gastropods, some laminated beds, some areas are partially to mostly covered (sample FM 518).</td>
</tr>
</tbody>
</table>

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**Figure 15:** Graphic log from the Franklin Mountains (500-600 m).
Figure 16: Graphic log from the Franklin Mountains (600-700 m).

### Franklin Mountains Section (Alacran Mountain Fm.)

**Location (lat/long):** 31°56.45'N/106°31.46'W

**Total Thickness Measured:** 783 m

**Date Measured:** June 15-18, 2008

**Measured By:** Jesse Koch and Drew Nelson

<table>
<thead>
<tr>
<th>Thickness</th>
<th>Graphic Log</th>
<th>Facies Code</th>
<th>Lithologic Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ft</td>
<td></td>
<td>6a; 6b</td>
<td>653.5-656.0 m: Mostly covered section, small bafflestone ledge at 657 m. Lots of phylloid algae, forams, bryozoans, ostracods, brachiopods, Tubiphytes. (sample FM 867).</td>
</tr>
<tr>
<td>m</td>
<td></td>
<td></td>
<td>601.0-653.5 m: Micrite to packstone, light gray. Lots of chert nodules at base of unit, new formation; decreases up-section, siltstone and shale units interbedded with limestone (3-5 cm thick), wavy beds in limestone, fusulinids, brachiopods, lots of gastropods at 610 m, bryozoans, ostracods, Tubiphytes, forams, fusulinids disappear up-section and other debris becomes more abundant (i.e. brachiopods), siltstone/mudstone layers disappear up-section, areas of bioherm (samples FM 603 and 629).</td>
</tr>
<tr>
<td>650</td>
<td></td>
<td>6a; 6b</td>
<td></td>
</tr>
<tr>
<td>600</td>
<td></td>
<td>5a; 5b; 1d</td>
<td></td>
</tr>
<tr>
<td>5a</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>400</td>
<td></td>
<td>6a; 6b; 1d</td>
<td></td>
</tr>
</tbody>
</table>
Figure 17: Graphic log from the Franklin Mountains (700-783 m).
Figure 18: Graphic log from the upper San Andres Mountains (0-101 m).
Figure 19: Graphic log from the upper Sacramento Mountains (0-100 m).
Figure 20: Graphic log from the upper Sacramento Mountains (100-186 m).
APPENDIX B:

ADDITIONAL LITHOFACIES PHOTOGRAPHS
Figure 1: High-frequency sequence from the Sacramento Mountains (Holder Formation-94-103 m). A) Uninterpreted high-frequency sequence from the Holder Formation. Note backpack for scale. B) Interpreted sequence stratigraphy of the outcrop above.
Figure 2: Small phylloid algae bioherm in the Holder Formation (Sacramento Mountains-84m). An example of a “patch” bioherm, likely from the middle to upper ramp environment.
Figure 3: Shallowing upward cycles in the Franklin Mountains (Hueco Canyon Formation-20-200 m). A) Uninterpreted photo. B) Interpreted shallowing upward cycles; a basal wackestone-packstone (~5 m; Facies Association 5), overlain by 1-2 m of shallow carbonate shoal facies (Facies Association 4).
Figure 4: Large phylloid algae bioherm at the top of the Alacran Mountain Formation (Franklin Mountains-753-783 m). Black scale bar is 10 m.
Figure 5: Large phylloid algae bioherms in the lower Alacran Mountain Formation (Franklin Mountains- upper 150 m). Each of the large cliff-forming units is a large phylloid algae-*Tubiphytes* bioherm. Black scale bar is 5 m.
Figure 6: Close-up view of phylloid algae in the Franklin Mountains (Alacran Mountain Formation-667 m).
APPENDIX C:
MISCELLANEOUS FIGURES NOT INCLUDED IN PUBLICATIONS
Figure 1: Composite $\delta^{13}C$ values plotted against the Gradstein et al. (2004) time scale.
Figure 2: Composite $\delta^{18}$O values plotted against the Gradstein et al. (2004) time scale.
Figure 3: GIS map of study area (#1) showing lithostratigraphic units and location of urban areas. GIS data from the New Mexico Bureau of Geology and Mining Resources; Texas data from the Bureau of Economic Geology (accessed online, October-2007).
Figure 4: GIS map of study area (#2) showing lithostratigraphic units and location of mountain ranges. GIS data from the New Mexico Bureau of Geology and Mining Resources; Texas data from the Bureau of Economic Geology (accessed online, October 2007).
APPENDIX D: MISCELLANEOUS FUSULINID IDENTIFICATIONS

All identifications by Dr. Greg Wahlman (Wahlman Geological Services, Houston, TX).

FRANKLIN MOUNTAINS SECTION

210m (Hueco Canyon Formation)
Fusulinid biostratigraphy: Latest Pennsylvanian, latest Virgilian (“Bursumian”); Schwagerina aff. longissimoidea (tangential specimen), Triticites cf. creekensis (partial axial specimen).

286m (Cerro Alto Formation)
Fusulinid biostratigraphy: Early Permian, Wolfcampian (Nealian); Pseudoschwagerina beedei (one axial specimen, and several tangential to oblique specimens), Schwagerina sp. (oblique specimens only), Schubertella sp.

348m (Cerro Alto Formation)
Fusulinid biostratigraphy: Early Permian, Wolfcampian (Nealian); Fusulinids are all tangential to oblique specimens of Pseudoschwagerina aff. beedei, Eoparafusulina sp., Schwagerina sp., Schubertella sp., and Staffella sp.

554m (Cerro Alto Formation)
Fusulinid biostratigraphy: Early Permian, Wolfcampian; Pseudoschwagerina cf. texana (one tangential specimen); rare Schubertella sp. juvenaria.

LOWER SACRAMENTO MOUNTAINS SECTION

67m (Holder Formation)
Fusulinid biostratigraphy: Late Pennsylvanian, Virgilian; Triticites cf. beedei.

73.3m (Holder Formation)
Fusulinid biostratigraphy: Late Pennsylvanian, Virgilian; Triticites cf. beedei.

114m (Holder Formation)
Fusulinid biostratigraphy: Late Pennsylvanian, Virgilian; Triticites sp.

177m (Holder Formation)
Fusulinid biostratigraphy: Late Pennsylvanian, Virgilian; Dunbarinella cf. ervinensis.

235m (Lower Abo Formation)
Fusulinid biostratigraphy: Late Pennsylvanian, Virgilian; Triticites sp.
GETTY #5 WILLIARD BEATTY CORE

9.65m (Upper Bough D member)
Fusulind biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites creekensis*, and possibly *T. cellamagnus*. One good fusulinid axial section has quite a large proloculus, and could be *T. cellamagnus*. Large *Triticites* having thick walls with coarse keriothecal structure appear in the upper Virgilian (Waubansee) and continue through the overlying “Bursumian”. Garner Wilde claimed in a paper that in such large triticitids, the proloculus increased in size markedly from the Virgilian to the “Bursumian”, and so the large thick-walled *Triticites* with large proloculi in this sample are typical of the “Bursumian” interval.

66.9m (Upper Bough B member)
Fusulind biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites creekensis* (very common), *Schwagerina* cf. *S. grandensis* (sparse), and *Staffella* sp. (sparse, which is a typical “Bursumian” assemblage.

TRES PAPALOTES CORE

11.3m (“Saunders” unit overlying Bough A member)

NICOR DDH-2 CORE

17.9m (Bursum Formation)
Fusulind biostratigraphy: Virgilian to (“Bursumian”); *Triticites* sp. indeterminant. The small oblique fragments and juvenile specimens in this thin-section are only adequate to allow a broad Virgilian to “Bursumian” age assignment.

UNOCAL STATE #33-1 CORE

21.05m (Bough B member)
Fusulind biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites* sp. indeterminant, and *Schwagerina emaciata*.

21.9m (Bough B member)
Fusulind biostratigraphy: Latest Pennsylvanian (“Bursumian”); One oblique specimen of a very large inflated *Triticites* that has abundant septal pores is of the *T. ventricosus* group. Several oblique specimens of medium-sized elongate primitive forms are *Schwagerina* cf. *S. grandensis*. 
24.4m (Bough B member)

24.7m (Bough B member)

29.9m (Bough A member)
Stratigraphic unit: Bough A
Fusulinid biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites creekensis*.

30.8m (Bough A member)
Fusulinid biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites creekensis*. Two specimens with moderately large proloculi, which are typical of many “Bursumian” *Triticites*.

32.0m (Bough A member)
Fusulinid biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites ventricosus* group. There are no well-oriented specimens, but the very large inflated triticitids with relatively plane septa and moderately large proloculi are most similar to *Triticites ventricosus*, a common “Bursumian” species. It is noteworthy that in the North American Midcontinent, *T. ventricosus* is particularly abundant in the Hughes Creek Shale of the Foraker Limestone, which is the cyclothem below the Red Eagle cyclothem that contains the new Pennsylvanian-Permian boundary at the Glenrock limestone-Bennett shale contact.

32.3m (Bough A member)
Stratigraphic unit: Bough A
Fusulinid biostratigraphy: Latest Pennsylvanian (“Bursumian”); *Triticites ventricosus* group, *Triticites* cf. *T. creekensis*, and *Schwagerina campa*. It should be noted that *Schwagerina campa* was described from the Glenrock limestone in Kansas, the unit that underlies the new Pennsylvanian-Permian boundary.
REFERENCES


