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BASIN-MARGIN SEDIMENTATION: EOCENE
LA JOLLA GROUP, SAN DIEGO COUNTY, CALIFORNIA

by

JEFFREY ALLYN MAY

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

DOCTOR OF PHILOSOPHY

APPROVED, THESIS COMMITTEE:

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HOUSTON, TEXAS

APRIL 1982
ABSTRACT

Continental to deep-marine facies transitions, eustatic versus tectonic controls on basin-margin stratigraphy, shelf-edge unconformities, and depositional mechanisms along basin margins were investigated for Middle Eocene strata, San Diego County, California. Coeval fan delta, nearshore, offshore, shelf, slope, submarine canyon, and proximal submarine fan facies indicate steep paleobathymetric gradients.

Mass-transport processes dominated the canyon-fan system: sandy and muddy debris flows, fluidized and liquefied flows, grain flows, high- and low-density turbidity currents, slumps, and rockfalls. The submarine canyon fill is tripartite and fining-upward, representing progressive detachment from a nearshore source. Planar- to convolute-laminated sandstone overlies a basal amalgamated pebbly sandstone. Lithologically variegated cross-cutting channels to 100's of meters wide cap the sequence.

A qualitative sand budget indicates the pebbly sandstone bypassed the wave zone, directly tapping an unsorted fluvial source. Residual lag deposition predominated. The coarsest fraction (0 to 3 φ) was also trapped and deposited by traction in the paralic zone, whereas intermittent suspension removed the 3 to 4 φ component onto the shelf. Size-sorting occurred downcanyon. Traction and intermittent suspension characterized inner-fan channel deposition. Lag plus traction and suspension constituents distinguish mid-fan channels.

Eustacy primarily controlled stratigraphic development. A depositional "hemicycle" of 9-10 m.y. corresponds to Vail et al.'s (1977) supercycle Tb. Punctuation by marine progradation was concurrent with
an intervening eustatic fluctuation. Subaerial notching of the shelf edge coincided with the Late Penutian sea level drop. During the subsequent rise, canyons eroded headward and a thin, retrogradational sequence was deposited. Coarse-grained, nearshore accumulations of the Early Ulatisian highstand were flushed basinward, responding to a slight sea level fall; submarine fan progradation resulted. After minor retrogradation, a Late Ulatisian to Early Nairizian highstand induced thick, progradational development.

Similar stratigraphic sequences developed simultaneously in other Pacific margin coastal basins. This suggests primary eustatic control on sedimentation and/or simultaneous continental-margin uplift and subsidence. Variations in rates of and absolute paleodepth changes indicate local tectonics. Combining global sea level fluctuations and resultant depositional patterns can provide a powerful tool in frontier exploration.
ACKNOWLEDGEMENTS

Dr. John E. Warme conceived and initiated this research project, and was instrumental in obtaining funding. I would like to express my gratitude for the guidance and discussions he provided during the course of this investigation, along with his critical review and advice concerning this manuscript. John M. Lohmar carried out the preliminary research upon which much of my study is based. I sincerely appreciate the insights he provided.

This dissertation was aided by the many helpful suggestions of Dr. John B. Anderson. Dr. Anderson also directed me to pertinent literature on grain-size analysis and allowed me free rein in his sedimentologic laboratory. The other members of my thesis committee were likewise of invaluable assistance. Dr. Richard E. Casey gave me a sound micropaleontologic background, and was a ready and willing reference person. Dr. Robert S. Cartwright, with little forewarning, gave much appreciated time and effort.

Field work would have been difficult, and certainly much more expensive, without the equipment loaned to me by Roy D. Adams and Douglas A. Steed. And because of the field area's location in a metropolitan resort region, nothing could have been accomplished without inexpensive living arrangements; Thea Schultze and Bob Butler went out of their way to help two complete strangers, and Edna Crawford aided further in these endeavors. Dr. Donald G. McCubbin, Marathon Oil Company, visited me in the field and gave freely of his vast knowledge and experience. Finally, access to key outcrops would have been prohibited if not for Mr. Jack P. Welch,
California Department of Parks and Recreation, Manager of the San Diego Coast Area. He and the staff of Torrey Pines State Reserve were extremely helpful in allowing for mapping and collecting within the Reserve.

Funding is gratefully acknowledged from a variety of sources: Marathon Oil Company, Union Oil Company of California, Shell Development Company, Amoco Production Company, the American Association of Petroleum Geologists, and the Geological Society of America. Stipend and tuition provided by a Nettie S. Autrey Fellowship during my last year at Rice University also made the financial scimping a little less severe. Dr. McCubbin made available results of nannoplankton and foraminiferal analyses from beach-cliff samples. Dr. Alvin A. Almgren of Union also examined foraminifera from submitted samples and furnished his interpretations; unfortunately, these were provided too late to be included in the final dissertation copy, but will be incorporated into a later publication. Dr. David G. Nussmann of Shell Development and Dr. Norman L. McIver of Shell Oil were especially helpful in aiding with photographic reproduction; plates of field and microscope photos and of the colored beach-cliff cross sections are due solely to their efforts. Connie Pedde and Anne Elsweiler helped tremendously with the typing of various drafts.

It would be impossible to mention all my colleagues who provided stimulating discussions and welcome support during the course of this research. Ross Yeo supplied much insight, as did Robyn Wright, who also shared her work on modern mass-flow deposits. Jill Singer and Ana María Pérez-Guzman kept my spirits high, and gave up some of
their valuable time to help hand-color beach-cliff cross sections at almost the last hour before my defense. William Miller, III directed me to useful references on Eocene mollusks.

My manager at Marathon Research, Dr. Richard J. Ebens, has been exceptionally encouraging and a great aid to my finishing this final draft. My family has been constantly supportive of my work, but as my graduate student career comes to an end, I want to express my deepest gratitude to my wife, Karen A. Crossen. At times, she was the only one who kept me going; she acted as my field and laboratory assistant, darkroom technician, draftsperson, typist, editor, proofreader, shoulder, cook, and maid. Karen sacrificed much time and sleep, and this Ph.D. should be at least half hers. So, to her this thesis is dedicated.

And to anyone else I may have forgotten to mention, it's not because of a lack of appreciation but rather a lack of memory!
# TABLE OF CONTENTS

**ABSTRACT**

***

**ACKNOWLEDGEMENTS**

v

**LIST OF FIGURES**

xii

**LIST OF TABLES**

xv

**INTRODUCTION**

1

**PURPOSE AND SCOPE**

1

**STUDY AREA**

15

- **Location**
  - 15
- **Accessibility**
  - 15
- **Climate**
  - 20
- **Physiography**
  - 23

**PREVIOUS WORK**

25

- **San Diego Geology**
  - 25
- **Submarine Canyon-Fan Models and Sediment Gravity Flows**
  - 26

**REGIONAL GEOLOGIC HISTORY**

30

- **Mesozoic Geology**
  - 30
- **Cenozoic Geology**
  - 31

**METHODS**

36

- **Field Methods**
  - 36
- **Laboratory Methods**
  - 39

**RESULTS**

46

**MEASURED SECTIONS**

46

- **Facies Descriptions**
  - 46

- **Lithofacies Exposed Along the Beach Cliffs**
  - 47

  - **Clast-Supported Conglomerate**
    - 47
  - **Matrix-Rich Conglomerate**
    - 53
  - **Amalgamated Pebbly Sandstone**
    - 53
  - **Massive to Laminated Sandstone**
    - 54
  - **Cross-bedded Sandstone**
    - 57
  - **Channelized Sandstone**
    - 62
  - **Turbidite Sandstone**
    - 62
  - **Burrowed Silty Sandstone**
    - 65
Interbedded Sandstone and Siltstone 68
Interlaminated Sandstone, Siltstone, and Mudstone 68
Thinly-Interbedded/Lenticular Bedded Sandstone and Mudstone 71
Fossiliferous Interbedded Sandstone and Mudstone 72
Laminated Siltstone and Silt-Shale 73
Slurried Unit 77
Pebbly Mudstone 77

Lithofacies Exposed Inland 81
Clast-Supported Conglomerate 81
Channelized Sandstone 84
Interlaminated Sandstone, Siltstone, and Mudstone 89
Laminated to Burrowed Sandy Siltstone 89
Pinch-and-Swell Sandstone 90
Turbidite Sandstone 93
Conglomerate-Based Sandstone 93
Planar-Laminated Sandstone with "Perched" Cobbles 95
Cross-bedded Sandstone 95

Vertical Facies Transitions 97

Lithofacies Exposed Along the Beach Cliffs 97
Lithofacies Exposed Inland 100

Lateral Facies Changes 102

Torrey Pines-Scripps Beach-Cliff Transect 102
Tourmaline Surfing Park Beach-Cliff Transect 107
Regional Correlations 108

TEXTURAL ANALYSIS 110

PETROGRAPHIC ANALYSIS 150

Thin-section Study 150
Mineralogy 150
Fabric Analysis 156
Cementation 160
Diagenesis 164

Heavy-mineral Study 170

SEDIMENTARY FABRIC ANALYSIS 175
Conglomeratic Fabrics 175
Sandstone Fabrics 177
Other Paleocurrent Indicators 178

PALEONTOLOGY 181
Previous Work 181
Timing of Deposition 186
Environments of Deposition 187

INTERPRETATION AND DISCUSSION 193

FACIES ANALYSIS 193

Beach-cliff Transects 193

Summary 193
Shallow-marine Deposits 196
Shelf Deposits 198
Submarine Canyon Deposits 201
Slope Deposits 204
Inner-fan Deposits 205
Middle-fan Deposits 210

Inland Area 217

Summary 217
Fan-delta Deposits 218
Nearshore Deposits 219
Offshore Deposits 223
Continental Shelf Deposits 231

BASIN RECONSTRUCTION 233

Shallow- to Deep-marine Facies Transitions 233
Depositional Timing Along the Torrey Pines-Scripps Transect 239
Provenance Variations in Time 243
Regional Paleogeography and Geologic History 248
Diagenetic History 270

DEPOSITIONAL PROCESSES 275
Depositional System and Sand Budget 275
Channelized Sediment Gravity Flows 280

Introduction 280
<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Submarine Canyon Fill</td>
<td>283</td>
</tr>
<tr>
<td>Conglomerate Fills</td>
<td>292</td>
</tr>
<tr>
<td>Laminated to Massive Sandstone Fills</td>
<td>297</td>
</tr>
<tr>
<td>CONCLUSIONS</td>
<td>331</td>
</tr>
<tr>
<td>REFERENCES CITED</td>
<td>341</td>
</tr>
<tr>
<td>APPENDIXES</td>
<td>365</td>
</tr>
<tr>
<td>APPENDIX I. Stratigraphic sections and sample locations</td>
<td>365</td>
</tr>
<tr>
<td>APPENDIX II. Sedimentary fabric analysis</td>
<td>391</td>
</tr>
<tr>
<td>APPENDIX III. Beach-cliff cross-sections</td>
<td>402</td>
</tr>
</tbody>
</table>
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Index map of study area</td>
<td>3</td>
</tr>
<tr>
<td>2</td>
<td>Regional facies relationships</td>
<td>5</td>
</tr>
<tr>
<td>3</td>
<td>Composite columnar stratigraphic section</td>
<td>7</td>
</tr>
<tr>
<td>4</td>
<td>Structures in sediment-gravity flow deposits</td>
<td>11</td>
</tr>
<tr>
<td>5</td>
<td>Processes in sediment-gravity flows</td>
<td>13</td>
</tr>
<tr>
<td>6</td>
<td>Locations of measured outcrops</td>
<td>19</td>
</tr>
<tr>
<td>7</td>
<td>A &quot;hazard&quot; of field work</td>
<td>22</td>
</tr>
<tr>
<td>8</td>
<td>Flow chart of laboratory methods</td>
<td>41</td>
</tr>
<tr>
<td>9</td>
<td>Types of resedimented conglomerates</td>
<td>49</td>
</tr>
<tr>
<td>10</td>
<td>Beach-cliff conglomerate lithofacies</td>
<td>52</td>
</tr>
<tr>
<td>11</td>
<td>Beach-cliff amalgamated pebbly sandstone lithofacies</td>
<td>56</td>
</tr>
<tr>
<td>12</td>
<td>Beach-cliff massive to laminated sandstone lithofacies</td>
<td>59</td>
</tr>
<tr>
<td>13</td>
<td>Beach-cliff sandstone lithofacies</td>
<td>61</td>
</tr>
<tr>
<td>14</td>
<td>Beach-cliff channelized sandstone lithofacies</td>
<td>64</td>
</tr>
<tr>
<td>15</td>
<td>Beach-cliff sandstone lithofacies</td>
<td>67</td>
</tr>
<tr>
<td>16</td>
<td>Beach-cliff interlaminated and interbedded lithofacies</td>
<td>70</td>
</tr>
<tr>
<td>17</td>
<td>Beach-cliff fine-grained lithofacies</td>
<td>75</td>
</tr>
<tr>
<td>18</td>
<td>Beach-cliff deformational lithofacies</td>
<td>79</td>
</tr>
<tr>
<td>19</td>
<td>Inland coarse-grained lithofacies</td>
<td>83</td>
</tr>
<tr>
<td>20</td>
<td>Inland lithofacies</td>
<td>88</td>
</tr>
<tr>
<td>21</td>
<td>Inland sandstone lithofacies</td>
<td>92</td>
</tr>
<tr>
<td>22</td>
<td>Correlation of beach-cliff sections</td>
<td>106</td>
</tr>
</tbody>
</table>
Figure 46. Submarine fan stratigraphic sequences
Figure 47. Channel and interchannel facies
Figure 48. Nearshore and offshore facies
Figure 49. Basin-margin dip sections
Figure 50. Basin-margin fence diagram
Figure 51. Beach-cliff age relationships
Figure 52. Environments of the San Diego Embayment
Figure 53. Correlation of eustacy and stratigraphic development
Figure 54. Late Early Eocene paleogeography
Figure 55. Early Middle Eocene paleogeography
Figure 56. Sequential fill of the submarine canyon
Figure 57. Medial Middle Eocene paleogeography
Figure 58. Late Middle Eocene paleogeography
Figure 59. Late Eocene paleogeography
Figure 60. Paragenetic sequences
Figure 61. Paleotransport through the San Diego Embayment
Figure 62. Classification, mechanical behavior, transport mechanisms, and deposits of mass transport
Figure 63. Variegated channel-fill deposits
Figure 64. Basin formation
Figure 65. Flextural loading model
Figure 66. Eustatic control on deposition
Figure 67. Effects of spreading rates
Figure 68. Middle Eocene stratigraphic correlations, West Coast basins
Figure 69. West Coast Middle Eocene paleogeography
<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Table 1. Locations of Measured Sections</td>
<td>16</td>
</tr>
<tr>
<td>Table 2. Textural Analysis</td>
<td>115</td>
</tr>
<tr>
<td>Table 3. Mineralogic Analysis</td>
<td>151</td>
</tr>
<tr>
<td>Table 4. Fabric Analysis</td>
<td>157</td>
</tr>
<tr>
<td>Table 5. Heavy Mineral Analysis</td>
<td>171</td>
</tr>
<tr>
<td>Table 6. Nannofossils from the Beach-cliff Transect</td>
<td>188</td>
</tr>
<tr>
<td>Table 7. Distribution of Extant Macrofossils</td>
<td>191</td>
</tr>
</tbody>
</table>
BASIN-MARGIN SEDIMENTATION: EOCENE
LA JOLLA GROUP, SAN DIEGO COUNTY, CALIFORNIA
INTRODUCTION

PURPOSE AND SCOPE

Eocene strata of San Diego County, California, present an exceptional opportunity for the analysis of shelf-edge unconformity development, of depositional mechanisms operating along basin margins, of three-dimensional relationships of continental to deep-marine facies transitions, and of eustatic vs. tectonic controls on the evolution of basin margin sedimentary sequences. Unlike many ancient convergent basin margins, the units deposited within the San Diego Embayment (Figure 1) are both widely exposed and relatively tectonically undeformed. These rocks consist of intertonguing, eastward-thinning strata (Figure 2) deposited during two major transgressive-regressive cycles and with a maximum thickness of approximately 700 meters (Kennedy and Moore, 1971a; Kennedy, 1975). Fluvial Eocene conglomerates and sandstones rest on a deeply eroded igneous and metamorphic basement complex to the east. Within a distance of 10 km these grade westward to lagoonal and nearshore bar deposits, and to a continental shelf and slope sequence which is exposed along the present shoreline, where they are underlain by Upper Cretaceous clastics (Kennedy and Moore, 1971; Kennedy, 1975). Miocene, Pliocene, and Pleistocene strata unconformably overlie portions of the Eocene sequence (Figure 3).

One major problem addressed by this study is the poorly-understood lithostratigraphic and time-stratigraphic relationships of the Eocene formations (Figure 2). Although Kennedy (1973a, 1975) correlated
Figure 1. Index map showing area of study in San Diego County relative to the Middle Eocene shoreline. This eastern limit to the San Diego Embayment is represented by the dashed line (from Kennedy, 1975).
Figure 2. Regional facies relationships of the Eocene units along a southwest to northeast transect (redrawn from Kennedy and Moore, 1971a).
EOCENE FACIES RELATIONS ACROSS THE DEPOSITIONAL STRIKE (KENNEDY & MOORE)

<table>
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<tr>
<th>SW</th>
<th>NE</th>
</tr>
</thead>
<tbody>
<tr>
<td>MISSION VALLEY FM. (NEARSHORE SHELF)</td>
<td>POMERADO CGL. (NON MARINE)</td>
</tr>
<tr>
<td>SCRIPPS FM. (NEARSHORE SHELF)</td>
<td>STADIUM CONGLOMERATE (DELTA)</td>
</tr>
<tr>
<td>ARDATH SHALE (OUTER SHELF)</td>
<td>FRIARS FM. (LAGOON)</td>
</tr>
<tr>
<td>MOUNT SOLEDAD FM. (TRANSGRESSIVE CGL.)</td>
<td>TORREY SS. (BARRIER BEACH)</td>
</tr>
<tr>
<td></td>
<td>DELMAR FM. (LAGOON)</td>
</tr>
<tr>
<td></td>
<td>UPPER CRETACEOUS &amp; BASEMENT</td>
</tr>
</tbody>
</table>

POWAY GRP. LA JOLLA GRP.
Figure 3. Composite columnar stratigraphic section of the San Diego area.
calcareous nanoplankton and planktonic foraminiferal zones of the continental shelf and slope deposits with paralic mollusk stages and non-marine vertebrate ages, the benthonic faunas used in regional correlation are facies-controlled (Lohmar, 1978). Therefore, detailed facies mapping within the study area (Figure 1), with a concentration on the Mount Soledad Formation, Torrey Sandstone, Ardath Shale, Scripps Formation, and Stadium Conglomerate, is used to reconstruct the paleogeographic relations and depositional history. Rapid paleobathymetric and paleoenvironmental gradients across coeval alluvial fan, paralic, continental shelf, and submarine canyon-slope-fan deposits aid the development of a model for basin-margin sedimentation. This model in turn may be used to test recent syntheses of shallow- to deep-marine facies patterns (Reading, 1978; Walker, 1979) and the areal development of their resultant interrelated sedimentary sequences (Vail and Hardenbohl, 1979).

Of particular importance is a major erosional surface which cuts across the lagoonal Delmar Formation and the Torrey Sandstone barrier system (Lohmar and Warme, 1978). Warme and Boyer (1975) defined the depositional environments of the Delmar Formation and Torrey Sandstone exposed along beach outcrops to the north. However, an irregular erosional surface abruptly separates the shallow-marine sequence from overlying submarine canyon and fan deposits. Therefore, another aspect of this project is to refine Lohmar's (1978) interpretation of this deep-water sequence based on detailed stratigraphic analyses and lateral facies relationships. This is done following methods established for interpreting submarine fan sequences (e.g., Walker and
Mutti, 1973; Mutti, 1974; Mutti et al., 1975; Ricci Lucchi, 1975; Walker, 1978). Of special note is that while much traditional work on submarine fan stratigraphic sequences has been directed toward the more distal flysch environments, the deep-marine San Diego Eocene units are channelized, coarse-grained, proximal deposits, a system rarely preserved and reported in the literature.

The proximal nature of the deep-water rocks also provides further data for helping resolve a point of contention in recent literature (Nilsen, 1980; Normark, 1980; Walker, 1980). Detailed depositional and morphological transitions from an inner submarine fan environment to the mid-fan and outer-fan complex is poorly known (see discussion section). This debate in part is due to the variability in clastic size fractions funneled to different fan systems, and the tectonic setting of each. The three-dimensional relationships exposed within the Eocene fan sequence provide detailed field evidence for the critical evaluation of these submarine fan models.

As knowledge and the ability to collect data about mass flow deposition has increased, both in the deep sea and on land, so has the recognition that mechanisms other than turbulent flow are operative in shelf-to-basin transport. Numerous theoretical models to explain the dynamics of sediment gravity movement by a transitional range of submarine hydrodynamic processes, and their resultant sedimentological variations (Figures 4 and 5) have been proposed (Dott, 1963; Sanders, 1965; Middleton and Hampton, 1973, 1976; Carter, 1975; Walker, 1978; Lowe, 1979; Nardin et al., 1979). Even though geologists have continued to refine their approaches to distinguishing ancient submarine
Figure 4. Sequences of sedimentary structures in hypothetical single-mechanism sediment gravity flow deposits (from Middleton and Hampton, 1973).
Turbidity Current

- Rippled or flat top
- Ripple drift (micro-
  laminae)
- Laminated
  good grading
  ("distribution
  grading")
- Fluxes, tool
  marks on base

Fluidized Flow

- Sand volcanoes or flat top
  Convolute lam.
- Fluid escape
  "pipes"
- Dish
  Structure?
- Poor grading
  ("coarse tail"
  grading")
- Grooves,
  striations
  on base?

Grain Flow.

- Flat top
  no grading?
- Mosaic
grain orientation
  parallel to flow
- Reverse grading?
near base
  scour, injection
  structures?

Debris Flow.

- Irregular top
  (large grains
  projecting)
- Massive
  poor sorting
  random fabric
- Poor grading, if any.
  ("coarse tail")
- Basal zone of
  "shearing"
  bread "crust?"
  striations at
  base
Figure 5. Conceptual model of the processes involved in the initiation, transport, and deposition of various sediment gravity deposits. Framework is one of time and/or space, and concentration of flows. Grain-support mechanisms (insert, lower right) include: 1) fluid turbulence, 2) liquifaction, 3) collision between individual grains (grain dispersive pressure), and 4) matrix strength (from Walker, 1978).
canyon and fan processes, there still remains a paucity of field support for these depositional models. Thus another aspect of this study is to define the products of a variety of mass flow mechanisms based on both field and laboratory criteria, and develop a practical model for shelf-to-basin sedimentological transitions.

The dependency of facies distributions on eustatic vs. tectonic controls is also examined for the Eocene of San Diego. The types, timing, and variation of materials deposited along the basin margin are defined, and related to local vs. worldwide sealevel changes. Associated with this, the geometry and timing of shelfedge unconformity development are investigated. All of these results are then related to similar sequences worldwide in an effort to develop a predictive model for submarine canyon and basin-margin fill related to global sealevel variations. This may be especially important for application to hydrocarbon exploration within similar depositional packages, providing a better understanding of porosity and permeability trends and predicting sandstone-body geometry development through time in submarine canyon and fan sequences.
STUDY AREA

Location

The field area comprises approximately 90 km² located north of San Diego, California (Figure 1). Though not truly rectangular, because the boundaries are dictated by accessible rock exposures, the area lies entirely within the Del Mar and La Jolla Quadrangles. Approximate borders are the Pacific Ocean coastline on the west to 117° 10' 30" west longitude eastward, and 32° 55' 30" and 32° 48' 15" north latitudes respectively on the north and south. Structurally the area is a tectonic coastline with Cretaceous and Cenozoic sedimentary deposition having occurred on a narrow, steep, coastal plain and continental shelves system. The eastern limit, and ultimate source terrane of much of the detrital material shed into this complex, is the northwest-trending pre-batholithic Jurassic volcanics and Cretaceous batholith system of the Peninsular Ranges.

Locations of measured stratigraphic sections referred to within the text are given in Table 1 and shown on Figure 6.

Accessibility

San Diego proper is the 11th largest city in the United States, and has a county-wide population of about 1.8 million (Lane, 1980). Its freeway system is California's second largest, and a sprawling system of primary paved roads allow for easy travel throughout the county. Many unmaintained roads passable to 4-wheel drive vehicles are present on private, undeveloped areas, although a 2-wheel drive vehicle was adequate to provide access to Los Peñasquitos Canyon.
<table>
<thead>
<tr>
<th>SITE NUMBER*</th>
<th>NAME OF LOCATION</th>
<th>LONGITUDE AND LATITUDE</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Torrey Pines Reserve Canyon</td>
<td>32°55'10&quot;N 117°15'26&quot;W</td>
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<tr>
<td>2</td>
<td>Bathtub Rock Trail</td>
<td>32°54'53&quot;N 117°15'23&quot;W</td>
</tr>
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<td>Canyon #1</td>
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</tr>
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<td>4</td>
<td>Canyon #2</td>
<td>32°54'15&quot;N 117°15'9&quot;W</td>
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<td>Rose Canyon</td>
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<td>21</td>
<td>San Clemente Canyon</td>
<td>32°50'31&quot;N 117°11'51&quot;W</td>
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*Refer to Figure 6
Figure 6. Index map of field area showing outcrops of Eocene strata and locations of measured stratigraphic sections (numbers refer to specific sections listed in Table 1).
The main disadvantage of this easy accessibility is that urban construction and man-induced vegetation has quickly covered many rock outcrops (Figure 7). Still others are "off-limits," as law enforcement personnel and private landowners can prove to be stubborn in allowing close-up examination of road and construction cuts.

Climate

The climate of San Diego is distinguished by uniformity of temperature and regularity of sunshine. The climatic zone is defined as a Mediterranean Scrub Woodland by Rumney (1968). More classical schemes base the climatic zone on temperature and rainfall, and hence San Diego falls into the Cs class of Köppen, (mesothermal forest climate with a dry summer), or the Subhumid (dry) Mesothermal grouping of Thornthwaite (Rumney, 1968). These reflect the dry summer and relatively wet winter, with 80% of the annual rainfall occurring between November of one year and March of the next. The average yearly precipitation is 10.1 inches (25.5 cm) (Carpenter, 1913). Only 14 days per year average a quarter of an inch (0.65 cm) or more rainfall, and there is an average of less than 9 days per year without one hour or more of sunshine (Carpenter, 1913).

Relative humidity is greatest in spring and summer, averaging 80% or under, and lower in fall and winter, averaging 70%. During "Santa Ana" conditions, in which a high pressure cell develops over the interior of southern California, northeasterly or easterly föhn winds push over the Peninsular Ranges, raising temperatures and dropping relative humidity to 5% or lower (Carpenter, 1913; Rumney, 1968). The tempera-
Figure 7. One of the "hazards" of field work in the San Diego area, aptly portrayed by J. C. Holden (from Abbott, 1979).
"There! Right between the two conduits — that's the type area for the San Diego Formation. It was more accessible two weeks ago before they built the subdivision here..."
ture along the coast averages less than one hour per year above 90°F, and the lowest reading ever recorded was 32°F. Mean annual temperature is 61°F, with a mean January temperature of approximately 55°F and mean July temperature of approximately 70°F (Carpenter, 1913; Rumney, 1968).

Physiography

San Diego County is larger than the states of Vermont, New Hampshire, New Jersey, or Maryland, and nearly equals the area of Massachusetts, Connecticut, Rhode Island, and Delaware combined. The main physiographic provinces of this area are: 1) a double-crested mountain range, the Laguna Mountains, formed by Jurassic volcanics and Cretaceous granitic batholiths which trend northwest and rise to 3000 m; 2) the Lindavista Terrace to the west, composed of relatively flat-lying Cretaceous and Cenozoic sedimentary rocks, formed as a Pleistocene wave-cut platform and having undergone later uplift; 3) the Poway Terrace developed on the Pomerado Conglomerate, lying at an elevation of 275 to 325 meters within the Laguna Mountains, also possibly formed as a result of Pleistocene wave planation; 4) a desert region to the east of the Laguna Mountains (Kennedy and Peterson, 1975; Gastil and Higley, 1977).

The study area lies wholly within the coastal Lindavista Terrace belt. The terrace gradually rises from nearly vertical cliffs at an elevation of ≤ 100 m along the coastline to 165 m in the east. It is dissected by numerous ephemeral streams which trend northeast, draining the mountains. Some major valleys also trend northwest; in many cases these trends appear to be related to regional fault systems, having
preferentially eroded along the weakened zones. This intersecting pattern accounts for elongate trends of rolling hills and isolated flat-topped mesas. Many of the major highways and secondary roads are built within the eroded valleys.

The major structural element in the study area is the Mount Soledad fault block. To the south of Mount Soledad at Mission Bay, at least 100 m of vertical separation has taken place along the Rose Canyon Fault, with the west side down. In the vicinity of La Jolla, approximately 135 m of vertical movement took place, with the west side up (Kennedy, 1975). Thus the Mount Soledad block appears to have rotated along an axis normal to the fault, elevating Mount Soledad on the northwest side and depressing Mission Bay along the southwest. Folds associated with northeast Mount Soledad are probably due to compression associated with right-lateral strike-slip movement along the Rose Canyon and Mount Soledad Faults (Kennedy, 1975).
PREVIOUS WORK

San Diego Geology

Reconnaissance geologic mapping was first carried out in the mid-to late 1800's (Blake, 1856; Fairbanks, 1893). Ellis (1919) made the first attempt at establishing Eocene stratigraphic nomenclature in a groundwater investigation, by defining the Late Eocene Poway Conglomerate based on exposures near Poway, within the Poway Quadrangle. Clark (1926) proposed the name La Jolla Formation for Middle Eocene strata. Hanna (1926) carried out the first detailed regional stratigraphic and paleontologic study, and extended the Poway Conglomerate to overlying non-conglomeratic rocks. He also divided the La Jolla Formation into the Delmar Sandstone, Torrey Sandstone, and Rose Canyon Shale Members. Later field and biostratigraphic studies (Hertlein, 1929; Hertlein and Grant, 1939, 1944, 1954; Strand, 1961; Weber, 1963; Golz, 1973) included only minor revisions of early work.

Kennedy and Moore (1971a) made major changes in both Cretaceous and Eocene stratigraphy (Figure 2). They elevated the La Jolla Formation to group status, and redefined the Delmar and Torrey Sandstones as formations. They abandoned the Rose Canyon Shale Member of Hanna (1926), instead dividing it into four new formations - the Mount Soledad Formation, Ardath Shale, Scripps Formation, and Friars Formation. Finally, they raised the Poway Conglomerate to a group with three formations - the Stadium Conglomerate, Mission Valley Formation, and an unnamed conglomerate later named Pomerado Conglomerate by Kennedy and Peterson (1974).
Useful guidebooks to the area are those of Shephard et al. (1957), Milow and Ennis (1961), Kennedy and Moore (1971b), and Kennedy (1973b). The most up-to-date geologic maps of the San Diego area are those of Kennedy (1975) and Kennedy and Peterson (1975), largely based on Kennedy's earlier (1969, 1973a) studies. Much of the recent work has focused on the problem of possible source areas for the red to purple to gray metarhyolite clasts unique to the Eocene strata (De Lisle et al., 1965; Merriam, 1968; Woodford et al., 1968, 1972; Minch, 1972, 1979; Abbott and Peterson, 1978; Abbott and Smith, 1978). Exceptions to this are detailed facies analyses of the shallow marine Delmar Formation and Torrey Sandstone (Boyer and Warme, 1975; Clifton, 1979), preliminary facies analysis of coastal exposures (Lohmar and Warme, 1978, 1979; Lohmar et al., 1979), and a cursory basin analysis based on conglomerate sedimentology largely south and offshore of my study area (Howell and Link, 1979; Howell, 1980).

Submarine Canyon-Fan Models and Sediment Gravity Flows

A complete discussion of either of these two subjects would fill a large volume; the following provides only a highlight. Many important articles are omitted here, but included in later pertinent discussion sections.

Since the late 1930's the study of shelf-to-basin transport and deposition of terrigenous material has witnessed an explosive growth. The majority of initial work was concerned with the origin of submerged valleys - whether through subaerial and/or submarine processes (Daly, 1936; Johnson, 1938; Veatch and Smith, 1939; Shepard and Emery, 1941). Keunen (1937, 1965) established much of the groundwork on which
interpretations of submarine high-density, turbulent current flow and sedimentation are based. Successive, systematic submarine cable breaks downslope from submarine canyon heads then provided indirect evidence for the turbidity-current process in deep water (Heezen and Ewing, 1952; Houtz and Wellman, 1962; Krause et al., 1970; Heezen and Drake, 1974). Following this came the recognition of deep-marine coarse clastic deposits in the ancient and recent, with characteristic sedimentary structures and sequences due to submarine hydrodynamic processes associated with these turbidity flows (Kuenen and Migliorini, 1950; Bouma, 1962; Kuenen, 1964; Rusnak and Nesteroff, 1964; Dzulynski and Walton, 1965; Harms and Fahnestock, 1965; Sestini and Pranzini, 1965; Walker, 1965; Horn et al., 1971).

As knowledge and the ability to collect data about submarine canyon-fan deposition has increased, both in the deep sea and on land, so has the recognition that mechanisms other than turbulent flow are operative in deep-marine transport. Numerous theoretical models to explain the dynamics of downslope hydrodynamic processes, and the resultant plexus of sedimentary deposits, were proposed (Crowell, 1957; Dott, 1963; Sanders, 1965; Walker, 1967; Griggs et al., 1970; Carter, 1975). During the past decade, emphasis has been placed on refining this work. Genetic schemes which deal with the mechanics of transport and internal flow support for sediment gravity movement have been developed (Hampton, 1972, 1975; Middleton and Hampton, 1973, 1976; Lowe, 1976a, 1976b, 1979; Enos, 1977; Nardin et al., 1979). Oftentimes using somewhat circular reasoning, these genetic schemes have been applied in an attempt to explain various ancient deep-water depositional sequences and lithofacies (Lowe, 1972, 1976a, 1976b; Stanley and Unrug, 1972; Walker and

Concomittant with this, and indeed almost inseparable, has been a refinement in defining the wide variety of deep-water depositional environments, especially continental slope and submarine canyon-fan facies. Some of the impetus for this has come from studies of modern marginal settings (Gorsline and Emery, 1959; Moore, 1969; Nelson et al., 1970; Normark, 1970, 1974; Haner, 1971; Normark and Piper, 1972; Nelson and Kulm, 1973). However, the main advances were based on studies of ancient analogs. Early approaches to submarine canyon-fan facies analysis (Sullwold, 1960; Stanley, 1961; Walker, 1966, 1970; Jacka et al., 1968; Hubert et al., 1970; Piper, 1970) evolved into complex models accounting for three-dimensional morphology, vertical and lateral fan growth through time, and depositional mechanisms responsible for the canyon-fan system (Mutti, 1974; Mutti and Ricci Lucchi, 1972; Ricci Lucchi, 1975; Walker, 1978; Walker and Mutti, 1973).

One result of this rapid expansion in interpretative work is that hypothetical models of canyon-fan system deposition and deposits have outpaced factual bases. New technologies are developing to permit both direct and indirect observations of recent mass-flow processes and their actual deposits (Kelling and Stanley, 1968). However, the necessary experimental data are still lacking. Further investigation of well-exposed ancient systems will continue to refine the various models, especially if direct comparisons can be made to modern systems (von Rad, 1968; Normark and Piper, 1969; Nelson and Nilsen, 1974; Doyle and Pilkey, 1979; Nilsen, 1980; Normark, 1980; Walker, 1980). The
ultimate goal will be to develop predictive patterns for worldwide basin-margin sedimentation through time and in space.
REGIONAL GEOLOGIC HISTORY

For a detailed review of the San Diego area stratigraphy, the interested reader is referred to Kennedy (1975), Kennedy and Peterson (1975), and Gastil and Higley (1977). This section outlines the general geologic history, concentrating on the Eocene. A diagrammatic stratigraphic section is shown for reference (Figure 3).

Mesozoic Geology

The basement complex of the San Diego area includes: 1) the Upper Jurassic Santiago Peak Volcanics, an elongate belt of deformed and mildly metamorphosed volcanic and marine and nonmarine volcanioclastic and sedimentary rocks exposed from the southern edge of the Los Angeles Basin (210 km to the north) southward into Mexico, and 2) mid-Cretaceous plutonic rocks of the southern California batholith, which intrude the Santiago Peak Volcanics (Kennedy, 1975). Together these compose the core of the 2,500 km-long Peninsular Ranges of southern and Baja California.

Interpreting these sequences in light of plate tectonics suggests that San Diego was part of a Late Jurassic island arc lying just west of present-day central Mexico, after accounting for later right-lateral strike-slip faulting along the San Andreas Fault. This arc developed following a Middle Jurassic break-up of an Andean-type margin, with the opening of a back-arc basin underlain by oceanic crust whose remains lie east of the Peninsular Ranges (Nilsen, 1977b). Remnants of the Late Jurassic subduction complex lying west of the island arc are rare in the San Diego area, with only a few probable ophiolite fragments
present (Gastil and Higley, 1977).

Cretaceous geologic history was characterized again by an Andean-type margin after closing of the marginal sea. K-Ar dates in the Peninsular Ranges become younger eastward, from 112 to 75 m.y., reflecting continued subduction of the Farallon Plate (Nilsen, 1977a). Thus by Late Cretaceous time volcanic activity had ceased along the western shoreline, and short, steep drainage patterns developed along a deeply-dissected terrain of recently-unroofed granodiorite and metamorphic rocks (Gastil and Higley, 1977). The oldest post-batholith unit is the locally-derived, nonmarine, boulder-conglomerate Lusardi Formation. The oldest marine strata are the Point Loma and Cabrillo Formations, described in detail by Yeo (unpublished Ph.D. dissertation, in preparation, Rice University) for the San Diego and northern Baja regions.

Cenozoic Geology

The Tertiary history of San Diego is that of a narrow coastal plain, continental shelf, and continental slope which, in contrast to the diastrophism of plate convergence, arc-related igneous activity, and transform faulting in most of coastal California, has remained relatively stable (Gastil and Higley, 1977; Nilsen, 1977b). Relatively flat-lying strata represent a succession of coastal plains which were alternatively deposited and eroded along a continental margin with transportation of detrital material to adjacent restricted basins, without significant tectonic alterations.

No Paleocene rocks are known in the San Diego area. Instead, a striking erosional surface and laterite paleosol is developed on
crystalline rocks of the Peninsular Ranges and Cabrillo Formation (Gastil and Higley, 1977). This weathering surface is thus post-Lower Maestrichtian, and in most places underlies Eocene units. However, along the beach cliffs north of La Jolla, a similar paleosol is developed on possible Early Eocene units (Lohmar, 1978; further analysis in this study). Therefore, a warm, humid, tropical climate similar to that of the modern equatorial belt persisted into the Eocene (Peterson et al., 1975).

The Eocene San Diego Embayment witnessed two major transgressive-regressive cycles imprinted upon regional uplift of the Peninsular Ranges and subsidence of the adjacent coastal area (Kennedy, 1973a). Early to Middle Eocene nearshore and lagoonal deposits of the Delmar Formation and Torrey Sandstone transgressed over the basal Mt. Soledad Formation, which in turn unconformably overlies the Upper Cretaceous Cabrillo Formation. These Eocene units mark the appearance of red to purple to gray metarhyolite cobbles, the "Poway" clasts. While the Cretaceous conglomerates are: 1) very poorly sorted, with angular to subangular boulders, 2) derived from the local pre-batholithic and batholithic basement, and 3) severely weathered in situ, the Eocene conglomerates are: 1) moderately well-sorted of rounded clasts, 2) dominated by a metarhyolite to metadacite suite, with minor quartzite, and 3) derived from no known local source area (Peterson, 1970). This indicates that the Peninsular Ranges by this time were fairly low lying, contributing little coarse terrigenous material. These gravels were brought westward to San Diego via the Ballenas River (and possibly other fluvial systems) from a source in Sonora, Mexico (Abbott and
After a major Middle Eocene regression, with deposition of the Ardath Shale and Scripps and Friars Formations, a Middle to Late Eocene transgression occurred. Lohmar (1978) portrayed a model of deltaic development, with the Stadium Conglomerate funneled directly off the front of the delta into a submarine canyon-fan system. A second regressive period with deposition of the Mission Valley Formation and Pomerado Conglomerate took place. By this period the Eocene climate had shifted from a humid tropical to semi-arid climate (Peterson, et al., 1975). Middle and Upper Eocene nonmarine units contain beds, lenses, and nodules enriched in calcium carbonate interpreted as caliche paleosols. Thus the climate was warmer with a similar seasonal rainfall (though slightly higher than at present).

In the Oligocene, regional tectonism began to change. Far to the east, the andesitic volcanism was replaced by ignimbrites, the continental highland began to fragment into basins and ridges, and rivers previously draining to the Pacific were diverted elsewhere (Dibblee, 1977; Gastil and Higley, 1977). No known Oligocene strata are exposed in the San Diego region, with redbeds deposited only west of the Peninsular Ranges. This activity was in response possibly to the spreading center between the Farallon and Pacific Plates and then the Pacific Plate itself coming into contact with North America in southern California, thereby initiating by Late Oligocene the right-lateral strike-slip system operating today (Nilsen, 1977a).

No Miocene deposits are present in west-central San Diego County, though they are present in northern and southern portions and on the
continental borderland to the west. The Middle Miocene San Onofre Breccia extends as far south as Oceanside and is present on the Los Coronado Islands. Clasts within this unit are primarily blueschist, greenschist, quartz schist, amphibolite, serpentinite, and mafic plutonic rocks (Stuart, 1975). Rocks of similar age and provenance compose the Rosarito Beach Formation, which extends from south of Tijuana into southwestern San Diego County. These units were derived from sources to the west in the borderland, which became emergent as andesitic and basaltic volcanoes shedding detritus eastward, as well as south from the Los Angeles basin and north from Baja California, into a series of small basins. However, while this activity and further uplift, compression, and faulting took place in response to continued Pacific Plate subduction, San Diego remained a comparatively quiet highland (Gastil and Higley, 1977).

In the Pliocene, the Borderland as well as the San Diego coastal plain began to subside. Eocene rhyolites and quartzite and Miocene breccia were reworked with newly-derived metaandesite and granodiorite from the east. Nearshore conglomerates and sandstones of the San Diego Formation rest unconformably on pre-Pliocene rocks. In places this formation grades upward into the Plio-Pleistocene Lindavista Formation, whereas the contact is unconformable elsewhere, representing wave-cut terrace deposition. Although having little local effect southwestern California and northwestern Mexico became coupled to the Pacific Plate in the Pliocene and have experienced perhaps up to 300 km of right-lateral offset (Nilsen, 1977a; Page, 1977).

The Lindavista Formation, deeply stained red by iron oxide, in
turn is overlain by marine and nonmarine deposits of the Bay Point Formation. Thus, the Pleistocene was a time of regressive beach ridge and wave-cut platform development, with deeply eroding river valleys stranding fluvial terraces along channel margins. Lateritic soil formation indicates a warm, high-rainfall regime (Gastil and Higley, 1977). Surficial deposits of Pleistocene and Holocene age include stream terrace, landslide, alluvium, and slope wash material.

Major faults in the San Diego area lie subparallel to the San Andreas system, striking northwest. The main fault zone is approximately 20 km wide, extending from La Jolla on the north southward into Mexico, and is composed of three main systems, the Rose Canyon, Point Loma, and La Nacion faults. Although Quaternary strike-slip movement has predominated, dip-slip faults are present. The maximum age for the most recent movement is Late Pleistocene, as Holocene deposits are rarely displaced (Kennedy, et al., 1975). Since Late Pliocene, an average rate of 1 - 2 meters per thousand years strike-slip and of 10 - 15 centimeters per thousand years dip-slip has occurred, with the latter probably due to differential compaction along the margin of the San Diego Embayment (Kennedy et al., 1975; Gastil and Higley, 1977). A secondary fault system trends northeast (Kennedy et al., 1975).
METHODS

As previously outlined, this work investigates the scale and geometry, mode of emplacement, detailed petrology, relationships to underlying and laterally-equivalent rocks, transitions from shallow-to-deep water facies, precise ages, paleobathymetry, and genesis within the tectonic and eustatic framework of Eocene continental to deep-marine strata exposed in the San Diego area. In short, this is accomplished by: 1) defining and mapping the facies exposed along the coastal cliffs within the study area, 2) tracing facies inland and analyzing their transitions to shallow water, 3) collecting microfossil and nannofossil samples, 4) textural and petrographic analyses of collected samples, and 5) relating the whole sequence to similar outcrop and subsurface units onshore, offshore, and worldwide, and to recent developments in seismic stratigraphy concerning sealevel changes relative to basin-margin sedimentation and unconformity development.

Field Methods

Field work was accomplished during approximately 22 weeks, from early June to late July, 1978, mid-June to mid-August, 1979, one week in November, 1979, and late January to late February, 1981. In order to establish a framework into which sedimentological data could be placed, 21 stratigraphic sections were measured in the field (Table 1, Figure 6). Sufficient composite vertical exposure, generally greater than 40 m in height, and adequate regional coverage to display lateral facies variations were the bases in selecting outcrops. In this densely-populated metropolitan and prime resort area, most exposures
are seeded over and rapidly covered or built upon. Therefore, measured sections cluster along the beach cliffs and railroad track, with few road cuts.

Preliminary examination of individual sections involved picking out and describing major units. Measuring was then carried out bed-by-bed, with beds distinguished on the basis of erosional contacts or abrupt changes in grain size. Measuring was accomplished using a 1 1/2-meter Jacob's staff or 10-meter steel tape; composite sections were constructed with the aid of a hand-held level. For each bed, the following were noted: topographic expression, texture (grain size, sorting, shape of grains), type of rock, composition, color (based on the G.S.A. Rock-Color Chart), induration and presence of carbonate cement, fossils, primary and secondary structures, nature of contacts, geometry and lateral continuity, and any unusual features.

Samples for paleontologic, petrographic, and textural studies were selected on the basis of their "representativeness" for that type of bed or interval; i.e., samples were chosen for specific purposes and in a nonrandom fashion to characterize certain depositional modes or sequences. Samples were not collected from every bed or at specific intervals (say, one meter) for every section. Monotonously thick or repetitive sections were sampled at very large intervals, while closely-spaced specimens were gathered from some single beds. Although most units are extremely friable, oriented samples were collected from carbonate-cemented units to be used for grain orientation study. Intervals at which samples were taken are included with the detailed columnar stratigraphic sections (Appendix I).
Paleocurrent measurements were obtained from those rare beds which display traction-current features in three dimensions, such as ripple cross-lamination or dune and sand-wave cross-bedding. However, conglomeratic units yielded the majority of paleocurrent trends. Attitudes were measured only for elongate, flat cobbles - those with an a-axis to b-axis length ratio $> 1.5$ and with an a-axis $> 1$ cm. All of the conglomerate fabric data were acquired directly in the field, not from photographs as done by some previous workers (e.g. Davies and Walker, 1974). Although the photographic method is less time-consuming and less tedious, it is much less accurate owing to the projection of three-dimensional clast orientations onto a two-dimensional surface (Hein, 1979, Appendix 2). The friable nature of the rocks studied also makes them more ammenable to direct attitude measurement.

Because of the relative absence of suitable clasts, 50 closely-spaced cobbles meeting the size and shape requirements were analyzed for each bed after randomly selecting the first clast for measurement; i.e., no grid pattern or traverse lines were used because of the great variations in sorting and size of clasts. Strike and dip measurements were made using a Brunton compass on cobbles carefully dug out of the surrounding matrix, then reinserted after determining the a-b plane. The very low regional dip - less than $10^\circ$ - facilitated in most cases the direct measurement of preferred fabric without the need for later rotating the data back to the original depositional attitude. Replicate measurements on the same bed during two different field seasons yielded vector means within $15^\circ$ of each other.

The in-depth investigation of the facies exposed along coastal
cliffs entailed the construction of large-scale cross-sections of the units, mapping major channel boundaries and lithologies. Oblique aerial photographs were first taken during helicopter flights along the cliffs. Inaccessible portions of the exposures were then mapped on the photos with the use of binoculars from beach level, and correlated with laterally-equivalent units described and measured within the stream valleys dissecting these cliffs. Accessible areas were field checked after defining the units from beach level, and samples for laboratory processing were collected. The final mosaics include an 8-km transect from north of La Jolla between Torrey Pines State Reserve and Scripps Institution of Oceanography, and a 0.4 km transect from Tourmaline Surfing Park, Pacific Beach.

Aside from determining and mapping the variations in channel-fill lithologies and individual channel dimensions, the three-dimensional relationships were also examined. Strikes of channel walls were obtained by digging away material truncated by the channel. A notebook was placed against the overhanging channel margin, and the dip and strike of the notebook was recorded (see Walker, 1975). Channels were also traced inland where possible in order to test lateral continuity and variations in strike, and to analyze their transitions into shallow water units.

Laboratory Methods

Two different schemes were utilized to analyze texture and petrography (Figure 8). Because of the friable nature of the majority of samples, they were directly amenable to textural characterization using the automated settling tube system in the Department of Geology at
Figure 8: Flow chart of laboratory methodology in analyzing texture and petrography of collected samples.
Sample Friable?

Yes

Complete Disaggregation in Calgon?

Yes

Wet Sieve & Retain Sand-sized Fraction

Grain-size (Settling Tube) Analysis

No

Further Disaggregation in Oxalic Acid

Heavy-mineral Separation

No

Well-cemented?

Yes

Thin Sectioning

Petrographic Study

No

Vacuum Impregnation

Grain Orientation Measurement
Rice University. This apparatus determines grain size distributions of sand- and silt-sized fractions (-1.0 to 6.0 \( \Phi \)) at 0.25 \( \Phi \) intervals, based on settling velocity of the grains (Anderson and Kurtz, 1979). Phi (\( \Phi \)) is the negative logarithm to base 2 of the grain diameter. Textural analyses made in this manner respond to real hydraulic characteristics of the sediment, and thus should reflect mechanics of transport and deposition (Reed et al., 1975). Dry weight percents of sand, silt, and clay fractions are given in a later section, as are graphic displays of frequency and cumulative size distributions, and mean grain size, standard deviation, skewness, and kurtosis for these samples.

After dry weighing, most samples readily disaggregated overnight in a mixture of deionized water and Calgon (sodium hexametaphosphate). Those specimens with grain aggregates caused by iron oxide cementation were subjected to heated oxalic acid with a piece of aluminum for 10 minutes (Ingram, 1971). Two samples lacking aggregates were processed in Calgon plus water and in oxalic acid, to test whether the acid treatment affected grain-size distributions. The silt and clay fractions were separated from the sand-sized material for each by wet sieving. The sand fraction (-1.0 to 4.0 \( \Phi \)) was measured on the large settling tube (140 cm long and 15 cm inside diameter). The coarse silt (4.0 to 6.0 \( \Phi \)) for numerous specimens was processed on the small tube (25 cm long and 6 cm inside diameter), but because an unknown percentage of each is of authigenic rather than primary origin, these results were judged not texturally significant.

Three hundred mineralogic grain counts were conducted on heavy-mineral splits for the 2.0 to 4.0 \( \Phi \) (sieved) fraction for each of the
friable samples. For general purposes, 300 grains is an adequate number to count in order to get maximum accuracy at the 50 confidence level from a minimum investment of time. Below 300 grains, the error increases rapidly, whereas above 300 it decreases slowly (Galehouse, 1971). "Lights" and "heavies" were separated using tetrabromoethane (specific gravity 2.97) following the method of Bates and Bates (1960). The heavy fraction was simply mounted on a glass slide using a thin layer of Buehler epoxy, then counted under reflected light magnification using the ribbon method (Galehouse, 1971). Heavy mineral suites are sensitive indicators of provenance, and are used herein to define variations in source-area input and hence to correlate regional depositional events.

Indurated samples were analyzed in a different manner. Well-cemented samples were simply thin-sectioned, while lightly-cemented samples were vacuum impregnated with blue epoxy prior to thin sectioning. Etching in hydrofluoric acid for 2-3 minutes followed by a 30-second emersion in a sodium cobaltinitrite solution stained potassium feldspar. Three hundred point counts per sample were made using a mechanical stage on a Zeiss Photomicroscope II, following the Glagolev-Chayes (grid) method (Galehouse, 1971). These data are provided as percent pore vs. cement vs. matrix vs. grains, as well as types of grains. Following the scheme of Dickinson (1970), the grains are then normalized to 100% for quartz, feldspar, and lithic fragments, and for quartz, potassium feldspar, and plagioclase. Petrographic descriptions of each sample include cement types and relationships, sorting, packing, types of grain contacts, sphericity, roundness, microstructures, and diagenetic features. These data aided in defining facies variations and deposi-
tional mechanisms. A generalized paragenetic diagram was also generated
in order to define post-depositional history, and is used to provide
insight as to the reservoir potential of this and similar systems.

Dimensional orientation of quartz and feldspar grains were made on
15 well-cemented samples collected from a variety of locations along
the Torrey Pines-Scripps beach-cliff transect. Fabric analysis from a
sedimentation unit may yield data on both the flow direction of its
depositing current and the depositional mechanisms of that current (see
Hiscott and Middleton, 1980). Combining grain orientations of sand-
stones with fabric data from conglomeratic units, regional paleocurrent
directions, types of depositional environments, and variations in depo-
sitional style through time are defined. Measurements relative to north
of either 100 or 50 quartz sand grains, with an a-to-b axis length ratio
of \( \geq 1.5 \), were taken per petrographic section from a Zeiss Projection
Attachment mounted on the photomicroscope. A mechanical point-counting
stage facilitated selection of appropriate grains which intersected the
center of the eyepiece cross hairs. The line of each grain's orienta-
tion was based on its least projection elongation (Dapples and Rominger,
1945), or if bilaterally symmetrical, the trace of the symmetry plane
was used (Hiscott and Middleton, 1980). Taking into account operator
errors of sampling, thin sectioning, and orientation measurement, a
rough estimate of standard deviation of the vector mean for each sample
is 20°.

Fabric data (Appendix II), both cobble imbrication measurements
and grain orientations, were plotted on rose diagrams where numerous
measurements were taken per bed, and analyzed using a von Mises or cir-
cular normal distribution computer program (Curray, 1956; Till, 1974). Statistical values obtained were: 1) mean vector direction, 2) mean vector magnitude (1.0 if all the vectors are in the mean direction, and 0.0 if paleocurrent vectors are randomly distributed over 360°), and 3) angular standard deviation. Mean vectors and single paleocurrent trends (e.g., channel axes) are plotted along the stratigraphic columns at appropriate intervals (Appendix I) and on the paleogeographic map.

Paleontologic information was obtained from previous researchers' published work, macrofossils identified herein, nannofossils processed by C. H. Ellis while with Marathon Oil Company, Denver Research Center, and further foraminiferal studies (D. G. McCubbin, oral communication). The timing and environments of deposition obtained from these data aided in constructing a paleogeographic map and a model for the evolution of the Eocene basin margin of San Diego.
RESULTS

MEASURED SECTIONS

Stratigraphic sections were measured in detail at 21 localities. Each locality was given a geographic name and a reference number (Table 1 and Figure 6). Recent stream valleys that dissect the beach cliffs along the Torrey Pines-Scripps transect were numbered sequentially, 1 through 6, proceeding from Bathtub Rock south.

Detailed bed-by-bed lithologic descriptions are not presented here, but are available upon request. Graphic representations of the sections are given in Appendix I, along with orientations of current structures and interpretations of fining- and coarsening-upward sequences and of depositional environments.

Facies Descriptions

The term 'facies' is defined as the "general appearance or nature of one part of a rock body as contrasted with other parts" (American Geological Institute, 1957). For sedimentary rocks, 'facies' refers to a unit of rock distinguishable in the field from adjacent rock units based upon lithologic characteristics and sedimentary and biogenic structures (de Raaf et al., 1965). The vertical and lateral sequences of sedimentary facies are not random, but instead result from physical, chemical, and biological processes in the depositional subenvironments (Pettijohn, 1975). The facies motif (three-dimensional pattern) may thus be used to interpret the regional system of deposition.

In this chapter, a basic description is given of each lithofacies,
i.e. facies based primarily on lithologic character. This information is later used to construct a depositional model.

**Lithofacies Exposed Along The Beach Cliffs**

Descriptions for facies of the Delmar Formation and Torrey Sandstone discussed in detail by Boyer and Warme (1975) and Clifton (1979) will not be repeated here. However, some of the Torrey Sandstone facies have not been delineated elsewhere, and are included below. Twenty facies from the Torrey Sandstone, Ardath Shale, Scripps Formation, and Mt. Soledad Formation are defined from the Torrey Pines-Scripps transect and from Tourmaline Surfing Park; each is described below.

**Clast-Supported Conglomerate:** Conglomerates containing cobble-sized and rarer pebble- and boulder-sized clasts are prevalently clast-supported with a coarse- to medium-grained sand-sized matrix. These conglomerates display a variety of grading, stratification, and fabric patterns which fit into Walker's (1975) models for resedimented conglomerates, plus two other types recognized by Hein (1979). These are: 1) disorganized (ungraded), 2) graded, 3) graded-stratified, 4) inverse-to-normally graded, 5) inverse graded-to-disorganized, and 6) inverse graded (Figure 9). The most common type present along the beach-cliff transects is disorganized (Figure 10a); poorly graded, graded-stratified, inverse-to-normally graded and inverse-to-disorganized are subordinate. Grading is present most often near the top of individual conglomerate packets. Stratification is poorly developed, although crude graded-stratified examples are present, with either horizontal or inclined stratification. Fabric of these conglomerates is predominantly a-axis parallel to flow and imbricated with the a-axis dipping in the presumed
Figure 9. Sketch of different types of resedimented conglomerates (from Hein, 1979).
1. DISORGANIZED

2. GRADED

3. GRADED — STRATIFIED

4. INVERSE — TO-NORMALLY GRADED

5. INVERSE GRADED — TO-DISORGANIZED

6. INVERSE GRADED
upcurrent direction, although a-axis transverse to flow with the b-axis imbricated, as well as unordered fabrics, are present. Matrix-supported conglomerate is found only at the tops of sedimentation units. Clasts within these conglomerates are typically crystalline metavolcanics and quartzites, although fossiliferous siltstones and mudstones and reworked shell debris are locally abundant. Clasts range in size from pebbles (0.4 to 6.4 cm) to 25 cm, although most are less than 10 cm in their maximum dimension.

Two subfacies of coarse conglomerate are defined. One is completely amalgamated, as evidenced by abrupt changes in cobble size and scoured lensoid remnants of massive to laminated sandstone. Lenses of mudstone rip-ups (many bent and therefore deposited before complete consolidation) as well as huge mudstone blocks meters in length are encompassed by the amalgamated crystalline-clast conglomerate. These provide further evidence of the cross-cutting amalgamated nature. Individual sedimentation units range from about 0.2 to 3 meters in thickness, and form complete amalgamated sequences up to 11 meters thick. While most of the individual conglomerate packets in this subfacies display basal scour, some are flat-based, generally thinner, and display the best stratification (Figure 10a).

The other subfacies is isolated channel-form conglomerate (Figure 10b). This subfacies, like the amalgamated conglomerate, is cross-cutting and may be amalgamated. However, individual sedimentation units are thinner, with a 1.5 meter maximum thickness, and pinch out laterally within meters or 10's of meters (Figure 10c). Sandstone- and mudstone-clast lenses are subequal in total thickness to crystalline-
Figure 10. Conglomerate lithofacies exposed along the beach cliffs.

A) Clast-supported conglomerate, amalgamated subfacies, composed of cross-cutting crystalline-clast and intraclast conglomerates. The crystalline-clast units are predominantly disorganized, with some poorly-graded to graded-stratified intervals.

B) Clast-supported conglomerate, isolated channel-form subfacies, with lateral pinching out (arrow) of thick sedimentation units.

C) Close-up of above. This subfacies is better organized than the amalgamated conglomerate subfacies.

D) Matrix-rich conglomerate. Note the stratified nature and sandstone lenses with floating clasts.
clast conglomerate where amalgamation does occur. Otherwise, conglomeratic channels are isolated, predominantly eroding into mudstone and gradationally passing upward into massive to laminated, or amalgamated channel-form, sandstone. This channel-form subfacies is typically better organized, with graded-stratified and graded types more abundant than disorganized.

**Matrix-Rich Conglomerate:** Only one major exposure of this facies is present along the beach-cliff transects. It is interbedded clast-supported to matrix-supported by pinkish-gray, medium- to fine-grained sandstone (Figure 10d). Some lenses of sandstone have horizontal planar to wavy laminations. The conglomerate is poorly horizontally stratified with some possible inverse grading. Fabric appears to be a-axis transverse to flow with b-axis imbricated. Locally, the sheet-like conglomeratic layers become disorganized and clast-supported. Bases of individual layers do not appear erosive. The sandstone matrix, which is always > 40%, increases near the top of the sequence.

A minor subfacies is matrix-supported cobble- to pebble-conglomerate which occurs at the very top of the clast-supported conglomerate sequences described above. The thickness of this subfacies is always < 20 cm, and typically one clast-layer in thickness. Oftentimes this matrix-supported layer represents the culmination of a grading-upward interval, or it may be disorganized. Fabric is typically unordered or a-axis imbricated.

**Amalgamated Pebby Sandstone:** These rocks are composed of cross-cutting and annealed, yellowish-gray to iron-stained orange, pebbly, granule to medium-grained sandstone packets. The sandstone is very
poorly sorted with abundant lithoclasts, feldspar, and quartz. Each cut-and-fill unit may be massive and structureless or faintly laminated due to grain size segregation (Figure 11a). The basal portion of many individual packets contains mudstone intraclasts which have either very irregular margins and are often matrix-supported (Figure 11b), or are very rounded and in point contact with one another (Figure 11c). Isolated clasts and very large mudstone blocks ("outrsize" clasts) to 5 meters or more in length are present "floating" within the sandstone at all levels (Figure 13a). Large crystalline clasts are rare.

Evidence of amalgamation is displayed in three ways: 1) change in grain size, i.e. granule or coarse-grained sands cutting across a medium-grained sandstone unit, 2) high to low angle of discordance between the laminations of separate sedimentation units (Figure 11a), and 3) intraclast-rich basal layers (Figures 11a, 11b, and 11c). The lowermost contact of this facies is erosive, having formed overhanging ledges (Figure 11d), erosive remnants projecting up into the sandstone (Figure 11a), and injection features within the underlying unit. Huge mudstone and siltstone blocks up to the size of train boxcars rest on this contact (Figure 13b). Bioturbation structures and fossiliferous debris are absent within this facies as are sedimentary structures formed by traction, such as cross-beds or ripple cross-lamination.

**Massive to Laminated Sandstone:** Somewhat gradational with the amalgamated pebbly sandstone, this rock type is yellowish gray, coarse- to fine-grained, micaceous sandstone with rare thin mudstone laminae. While appearing massive and structureless from a distance, size segregation of grains often produces faint laminations. These are
Figure 11. Amalgamated pebbly sandstone lithofacies exposed along the beach cliffs.

A) Amalgamated pebbly sandstone unconformably overlying cross-bedded sandstone (erosional surface marked by arrows and rock hammer). Note the individual cross-cutting sedimentation units, marked by basal intraclast-rich intervals and laterally-truncated planar-laminated fills.

B) Basal intraclast-rich intervals of the amalgamated pebbly sandstone units are disorganized or display inverse-to-normal grading. The mudstone clasts typically have very irregular margins, and are in point contact or "float" within the sandstone matrix.

C) Less predominant are organized basal intraclast-rich layers with rounded mudstone clasts, imbricated with a-axes parallel to flow.

D) The pebbly amalgamated sandstone facies cuts the underlying strata, forming a stepped and undercut erosional unconformity (arrows).
best brought out by organic- or biotite-rich layers (Figure 12a). The
laminae are predominantly horizontal and planar, although intervals of
box-like, arched, or "flamed" (overturned-in-a-downstream-direction)
convolutions are present (Figure 12b). These laminae may be broken
through due to syndepositional dewatering.

Stringers of small mudstone clasts, mudstone laminae with flame
structures (Figure 12d), dish-and-pillar structures (Figure 12c), and
clastic dikes are present in upper portions of this facies. Rare
climbing ripples with only the lee-side laminations preserved (type 2
of Jopling and Walker, 1968) also occur in this uppermost interval.
Layers of small iron-stained concretions are present throughout. Iron
staining also highlights the laminae, although some obviously cuts
across bedding.

Massive to laminated sandstone beds are commonly stacked, forming
sequences up to 20 meters thick.

**Cross-bedded Sandstone:** Two types of cross-bedded sandstone occur.
One subfacies is of isolated, flat-bottomed, medium to thin beds which
are laterally discontinuous (Figure 13c). Cross-strata within this
type are tangential to the base and cut across the whole bed. The sand-
stone is iron-stained light orange brown to light gray, medium- to fine-
grained, and micaceous. Organic matter and mollusk fragments are common,
while small clay chips and burrows, including Ophiomorpha, are rare to
locally abundant.

The other type of cross-bedded sandstone is very pale orange to
yellowish gray, coarse- to medium-grained, trough cross-bedded with
some pebbly layers and rare wood fragments (Figure 13d). The sets of
Figure 12. Massive to laminated sandstone lithofacies exposed along the beach cliffs.

A) Horizontal planar to wavy laminae are especially obvious, when highlighted by plant fragment and biotite concentrations.
B) Planar laminae may become convoluted upward within a sedimentation sequence, developing contorted folds and flame structures (down-current direction to the right).
C) Uppermost portions of sedimentation sequences also display dish-and-pillar structures, the result of penecontemporaneous dewatering.
D) Mudstone beds overlying massive to laminated sandstone intervals are disrupted by sandstone dikes and sills (arrows).
Figure 13. Sandstone lithofacies exposed along the beach cliffs.

A) Outsize clast "floating" within the amalgamated pebbly sandstone.

B) Large displaced lithified block resting on the basal erosive contact of and surrounded by amalgamated pebbly sandstone.

C) Isolated, flat-based sandstone cross-bed with cross-strata tangential to the base. Note the overlying fine-grained deposits draped over the sandstone unit, and the bed's laterally-discontinuous nature.

D) Stacked sandstone cross-bed sets, predominantly displaying trough cross-strata.
cross-beds are stacked, each set ranging from 0.5 to 4 meters in thickness. These form sequences up to 25 meters total, although thick to medium beds predominate. More rare are planar cross-beds and undulatory bedding surfaces. Mollusk fragments, burrowing, and ripple cross-lamination are typically absent.

**Channelized Sandstone**: Channels of various dimensions, commonly 10's of meters wide and meters thick, are eroded into and isolated within finer-grained material or are amalgamated, cutting into underlying channel sandstone and intervening mudstone (Figures 14a and 14b). The sandstone is medium- to fine-grained, light brown to very light gray, and micaceous. Bases of channels are irregular and erosive, with load and flame structures, pry-ups, or rare flutes where cut into mudstone (Figure 14c). Rip-up mudstone clasts are common along channel bottoms, while crystalline clasts are much rarer.

A wide variety of sedimentary structures are found, although structureless sands, horizontal laminations, and multiple-graded fills broken by stratification bands are most common. Rarer features are stringers of mudstone clasts, convolute and flamed laminae, dewatering pipes, dish structures (Figure 14d), and isolated mudstone and crystalline clasts. Planar to trough cross-stratification and ripple cross-laminations are sometimes present at the tops of channel sandstone. Mud-lined *Ophiomorpha* burrows and mud-filled burrows are other rare but locally abundant features.

**Turbidite Sandstone**: This facies is uncommon. The sandstone is medium light gray, coarse- to fine-grained, and micaceous. Turbidite sands only overlie mudstone units or each other, and some display very
Figure 14. Channelized sandstone lithofacies exposed along the beach cliffs.

A) Amalgamated, cross-cutting channel-form sandstone units. Intervening fine-grained intervals are volumetrically subordinate, lining channel bases or draping upper flat surfaces of individual sandstone fills.

B) Small, isolated channel-form sandstone bed, with a concave-upward base and lateral thinning. Note hammer below the bed to the right for scale.

C) Base of channelized sandstone unit loaded into the underlying finer-grained strata.

D) Channelized sandstone fill displaying dish structures. Laminated fills lacking evidence of such dewatering are more common.
erosive bases (Figure 18a). However, perfectly flat, non-erosive bottoms predominate. Beds with planar contacts are laterally very continuous; stacked, cross-cutting turbidites are slightly channel-form and pinch out laterally. Flutes are present though rare.

The most common association of internal structures follows the Bouma (1962) T(a)bce sequence (Figure 15a). A thin, massive layer with mudstone rip-ups is uncommon, while horizontal laminations passing upward into climbing ripples (both type 1 and type 2, Jopling and Walker, 1968) and/or convolute laminations prevail. Collapse structures and dewatering pipes associated with syndepositional fluid escape are present (Figure 15b), as are small syndepositionally-faulted intervals. The climbing-ripple intervals are often quite thick (to 0.5 meters) and of two types; one lacks preservation of stoss-side laminations thus displaying only upcurrent-dipping remnants, while the more common is of complete out-of-phase drifting-ripple forms.

Burrowed Silty Sandstone: Light-gray to pinkish-gray, medium- to fine-grained, micaceous sandstone is associated with the trough cross-beded sandstone facies. Remnants of mudstone and siltstone laminae within the sandstone are cut by sand-filled burrows 0.5 to 5 cm in diameter (Figure 15c). Some burrows are pellet- or mud-lined *Ophiomorpha*, and the majority are vertical to near-vertical. Burrows are better cemented than the surrounding material and stained light brown by iron oxide. Molds of small, leached-out mollusk fragments are common, and plant fragments may be present.

The upper surface is typically erosive, cut by cross-beded sandstone or has a layer of mudstone intraclasts preceding the cross-beds.
Figure 15. Sandstone lithofacies exposed along the beach cliffs.

A) Turbidite sandstone containing a classic $T_{abc}$ sequence (massive to laminated to rippled internal structures). The thin basal massive (structureless) interval contains small rip-ups (arrows) of the underlying material.

B) Syndepositional dewatering pipes and collapse structures (arrows) in a turbidite sandstone.

C) Burrowed silty sandstone facies dominated by vertical to near-vertical Ophiomorpha traces. Burrows are iron stained, and the more resistant, well-cemented fills weather out in relief.

D) Burrowed concretionary sandstone bed interbedded with siltstone. Traces are variously meandering, vertical, and horizontal, and include Ophiomorpha, Thalassinoides, and Planolites.
The lower surface is often gradational, passing upward from faint cross-stratification into the burrowed intervals.

**Interbedded Sandstone and Siltstone:** Fine-grained light-gray to light-brown, micaceous sandstone is medium- to thinly-interbedded with gray siltstone. Burrows and complete bioturbation, especially in the sandstone beds, predominate. This includes *Ophiomorpha*, horizontal *Thalassinoides*, *Planolites*, and smaller branched and unbranched, horizontal to vertical to meandering, mud-filled and sand-filled, burrows (Figure 15d). Complete bioturbation and subsequent preferential iron staining of coarser-grained material has produced color mottling in some beds. Some sandstone beds also display faint low-angle cross-bedding or horizontal planar to wavy laminations, as well as isolated or stringers of mudstone and siltstone clasts. Beds with these features have an erosive, channel-form base, and gradationally pass upward into a burrowed top portion.

Many of the sandstone beds are carbonate cemented and concretionary to ledge-forming, and contain mollusk fragments. Upper surfaces of such beds display abundant horizontal branching and meandering burrows, plus dendritic grazing trails.

**Interlaminated Sandstone, Siltstone, and Mudstone:** Very thick to thin discontinuous laminae of fine- to very fine-grained, yellowish-brown to gray sandstone is interlaminated with varying amounts of light brown to yellowish gray siltstone and/or silty to clayey medium to light gray mudstone (Figure 16a). This ranges from wholly sandstone and siltstone to siltstone and mudstone with few sandstone laminae. Individual graded sequences, i.e. decreasing sandstone and increasing
Figure 16. Interlaminated and interbedded lithofacies exposed along the beach cliffs.

A) Interlaminated sandstone, siltstone, and mudstone facies with starved ripple and ripple cross-laminated intervals.

B) "Lam-scram" sequences typical of the interlaminated sandstone, siltstone, mudstone facies. The lower laminated interval becomes increasingly bioturbated upwards until completely scrambled. The lens cap marks the beginning of an overlying "lam-scram" sequence.

C) The thinly-interbedded/lenticular-bedded sandstone and mudstone facies displays starved ripples, mudstone drapes, grading of sandy intervals, and loading. These rhythmically-alternating layers may be described as thin-bedded ($T_{ce}$) turbidites.

D) Fossiliferous interbedded sandstone and mudstone facies. Whole and fragmented gastropods, bivalves, and worm tubes occur as layers and lenses within the sandstone units, especially as basal lags.
mudstone laminae upward, with thicknesses of 50 cm or less occur. These may be stacked as multiple-graded units to thicknesses of 2 meters.

Both sandstone and siltstone laminae are discontinuous and wavy, ripple cross-laminated, planar laminated, or occur as asymmetrical starved-ripple lenses. Ripple crests are somewhat sinuous and have a relatively large ripple index. Some sandstone layers have minute clay clasts along laminae. Rare thin sandstone beds are calcareous and contain shell debris. Mudstone laminae are also wavy and often have minute flame structures. Carbonaceous flecks are locally abundant, but fossil fragments are exceedingly rare.

Burrows and color mottling due to complete bioturbation are common. *Ophiomorpha* and *Scolicia* (heart urchin? burrows) are most prevalent in sandstone-siltstone sequences, while *Planolites*, large pyritized burrows, burrows with U-shaped backfills, and complete bioturbation are most common in muddier intervals. However, where mudstone alone predominates, biological structures decrease and *Chondrites* may be the sole burrow type present. A standard arrangement is for a completely laminated sequence to pass upward into small burrows, then larger burrows, and finally a bioturbated interval with complete mixing and disruption of laminations (Figure 16b). These laminated-to-scrambled intervals (termed "lam-scram" sequences by J. Warme) may be multiply stacked. Other intervals have tops truncated by erosion and only the large sand-filled burrows preserved.

*Thinly-Interbedded/Lenticular-Bedded Sandstone and Mudstone*: Thin to very thin beds of medium- to fine-grained, light-brown to yellowish-gray sandstone are ripple cross-laminated and commonly have an undula-
tory asymmetrically-rippled top and rhythmically alternate with light-
to dark-gray, clayey mudstone (Figure 16c). These may be described as
thin-bedded Tce turbidites (Bouma, 1962). Other fine- to very-fine
grained sandstone occurs as starved ripples which may be loaded and
therefore deformed into the surrounding muddy matrix. The coarse-
grained beds are flat based except where loaded into the mudstone or
where the mudstone has developed flame structures, while muddy layers
drape over underlying microtopography. Branching, mud-filled Chondrites
burrows extend down into the sandstone beds from their upper surface.

Where mudstone and sandstone are subequal in thickness, burrowing
is rare in the fine-grained units. However, when sandstone is solely
present as starved ripple lenses, burrowing and complete bioturbation
are prevalent. Ophiomorpha, Chondrites, Planolites, and vertical
pyritized or laminated sand-filled burrows are all present. Body
fossils are extremely rare.

This facies is gradational with the interlaminated sandstone, silt-
stone, and mudstone facies described above.

Fossiliferous Interbedded Sandstone and Mudstone: Thin to thick
beds of carbonate-cemented, fine- to very fine-grained, light-gray to
light-brown, micaceous, silty sandstone and sandy siltstone are inter-
bedded with thick to very thick beds of light- to medium-gray silty
mudstone to siltstone. While some sandstone and siltstone beds display
faint planar laminations and are ledge-forming, the distinguishing
characteristic of this facies is the abundant whole and fragmented
fossils in the mudstone, siltstone, and sandstone layers. Fossils are
particularly abundant in the coarser-grained intervals, locally com-
prising a large proportion of the rock (Figure 16d). These include both high- and low-spired gastropods, very large and small bivalves, and large colonial vermetid worm tubes. The shell material is often leached away, especially in the finer-grained beds. Layers and lenses of monospecific shells are present, especially in basal sandstone portions, although diversity overall is high. Little current orientation of shells is evident, although sub-parallel orientations of high-spired gastropods are present.

Mud-lined Ophiomorpha burrows and Chondrites are present throughout this facies. Bedding surfaces are irregular, with no evidence of extensive erosion, and appear to have been reworked by burrowing organisms. Mudstone beds are gradational upward to more fossiliferous siltstone layers, producing an inverse grading. Sand-filled burrows in the finer-grained intervals are often iron-stained and more resistant to weathering.

Laminated Siltstone and Silt-Shale: Thinly-bedded to thinly-interlaminated, gray to orange-brown siltstone and silt-shale contain rare thin, fine-grained sandstone lenses with siltstone rip-ups (Figure 17a). Complete intervals range up to 11 meters in thickness. Laminations are irregular to wavy, or ripple cross-laminated, and both biotite and plant fragments concentrate along bedding surfaces. Small concretions occur along some layers. Body fossils or their external molds are lacking. However, small burrows and bioturbated layers disrupt laminations. This facies displays some fissility but more often displays blocky to somewhat chippy weathering characteristics.

The rare sandstone lenses shallowly cut into the siltstone and
Figure 17. Fine-grained lithofacies exposed along the beach cliffs.

A) Laminated siltstone and silt-shale facies. Rare thin sandstone lenses and layers of body fossils are present, but monotonously inter-laminated siltstone with minute burrows predominates.

B) Cross-cutting channels filled with the mudstone facies. The cliff face above the basal landslide is approximately 70 m high (long thin arrow points to one-lane dirt road for scale). One channel base is draped with the mudstone facies and plugged with the laminated sandstone facies (short curved arrow).

C) Mudstone facies displaying typical blocky weathering and draping thin ripple cross-laminated sandstone beds and starved ripples.

D) Intervals of interlaminated siltstone and mudstone, and of burrowed mudstone. The mudstone facies is gradational with the interlaminated sandstone, siltstone, and mudstone facies.
in turn are draped by it. Upper surfaces of siltstone intervals are scoured by conglomerate or sandstone units.

**Mudstone (with Sandstone and Siltstone Beds):** Clayey to silty, light-gray to medium-dark-gray mudstone occurs as thick to very thick beds with thin to medium beds of fine- to very fine-grained grayish-orange to light-gray sandstone and siltstone. The coarser-grained beds are biotite-rich, often calcareous and ledge-forming, with slightly irregular bedding surfaces. Some laterally pinch out while others are continuous and concave upward, delineating very broad channels 10's to 100's of meters wide (Figure 17b). These coarse-grained beds are structureless or display a variety of sedimentary structures, including normal grading, planar to wavy horizontal laminations, cross-bedding, ripple cross-laminations, and mudstone partings (Figure 17c). Bidirectional small-scale cross-beds and ripple cross-laminations occur. Other beds are burrowed by mud-filled Planolites, Helminthoida, or are completely bioturbated. Some sandstones are calcareous, and contain abundant molluscan shell debris.

Mudstone intervals are 0.2 to 10 meters thick, and likewise contain a variety of internal structures. Chondrites, Helminthoida, composite burrows, and burrows with U-shaped back-fills are common, as are intervals of thin to indistinct, planar sandstone and siltstone laminae (Figure 17d). Starved and low amplitude siltstone ripples and sandy wisps are locally abundant. However, structureless, ungraded silty mudstone with uncommon external molds of small, leached, whole bivalves and gastropods or fragmented shell debris predominates. Carbonaceous flecks are present throughout. The mudstone weathers powdery to
splintery to blocky. In places this facies grades upward to fossiliferous interbedded sandstone and mudstone.

**Slurried Unit:** This facies is composed of syndepositionally-deformed mudstone or interlaminated sandstone, siltstone, and mudstone facies. Slump folds are simple, recumbent (like in a carpet which slid down an incline), or may be extremely contorted with fold axes at all angles (Figure 18a). These "olistostromes" may be quite thin, the slumped beds 10 or 20 cm in thickness and pinching out laterally, or very large, with the resultant crumpled mass up to 20 meters thick. In some cases, a zone of detachment can be seen up the inclined bedding plane along which the slump occurred. Original burrows may be stretched out and destroyed, and new ones may extend downward from the reconstructed upper bedding surface.

Some slurried units have a perfectly planar to somewhat irregular base, while others have an erosive sandy basal portion which cut across and incorporated underlying mudstone units as rip-up clasts. Partially pried-up flaps of underlying beds are present. Likewise, upper bedding surfaces may be irregular to planar, covered by passively-deposited units, or eroded by the next event. Bent, semi-lithified to well lithified sandstone blocks with irregular margins are incorporated into some slumps as olistoliths (Figure 18b). Some of these are stacked or imbricated. Large mudstone olistoliths are less common. In one rare case a slurried unit incorporated crystalline cobbles into the matrix.

**Pebbly Mudstone:** This last facies is transitional to syndepositional slurried units. Clayey to silty, light-brown to gray mudstone
Figure 18. Deformational lithofacies exposed along the beach cliffs.

A) Slurried unit overlain by sandstone turbidite (note irregular, erosive contact). Syndepositional slumping produced the extreme contortion and variety of fold-axis orientations within the slurried interval.

B) Slurried unit ("olistostrome" or "penecontemporaneous slump") which contains relatively undeformed, better-lithified sandstone blocks or "olistoliths" (arrows).

C) Pebby mudstone facies with remnant folded laminations, demonstrating a transitional, genetic relationship with the slurried unit facies.

D) "Pebby" mudstone facies containing a variety of clasts. Not only are pebbles represented, but crystal-line cobbles, whole and fragmented mollusk shells, and clastic intraclasts (mudstone and sandstone; arrows) are also present.
contains rounded mudstone and crystalline clasts as well as mollusk shell fragments. Remnant slump folds of laminated mudstone may be present, but the matrix is largely structureless with "floating" isolated clasts (Figures 18c and 18d). The upper bedding surface is typically eroded by overlying sandstone. The facies often has a channel-form base cutting across underlying mudstone or sandstone.
Lithofacies Exposed Inland

All of the facies, except for pebbly mudstone, occurring along the beach-cliff transects are present inland within the study area. Some minor variations (subfacies) were observed and are detailed below. Six additional facies are also described, and as discussed later are the signature of a transition from shallow marine to deep marine depositional environments.

Clast-Supported Conglomerate: Disorganized, crystalline-clast conglomerate equivalent to that of the beach cliffs occurs at the Genesee Avenue North location, as well as in the Los Peñasquitos Canyon area (where no stratigraphic section was measured). A distinct subfacies not present along the beach cliffs occurs inland and contains even less coarse- to medium-grained sandstone matrix, with a field estimate of 5% or less volumetrically (compared to 5 to 20% matrix along the beach cliffs). Lenses of sandstone with scoured, flat to concave-upward bases and abrupt changes in conglomerate clast-size attest to the cross-cutting, amalgamated nature. Distinguishing characteristics of this subfacies are: 1) predominance of normal and inverse-to-normal grading, with subordinate inversely-graded and disorganized intervals, 2) abundance of inclined to horizontal stratification, 3) lack of shell material, 4) rare remnant silty mudstone beds, and 5) very large crystalline clasts up to 0.75 meters in the a-axis dimension (Figure 19a). Mudstone rip-ups are scarce, though locally abundant. The fabric of these conglomerates is a-axis parallel to flow and imbricated, with other fabrics being rare. Examples of this
Figure 19. Coarse-grained lithofacies exposed inland.

A) Matrix-poor, clast-supported conglomerate. Note the multiple, thick, graded intervals, very large crystalline-clasts, and the sandstone lens with a concave-upward base and scoured upper contact. The lens is approximately one meter thick.

B) Channel-form conglomerate facies with imbricated clasts (a-axis parallel to flow) and filled with thin individual sedimentation units.

C) Channelized sandstone facies I with steep, stepped channel margin (arrows) and basal cobble and intraclast stringers. Note laminated appearance of fill.

D) Pried-up layer containing flame structures, typical of the basal contact of the channelized sandstone facies II.
facies are found at the Morena Boulevard and Rose Canyon localities.

Some channel-form conglomerate facies also display variations from that occurring along the beach-cliff transects. These channels are thicker, up to 5 meters, although individual conglomeratic sedimentation units may only be one or two clasts thick (Figure 19b). More typically, the channels have a clast-supported conglomerate basal layer with some mudstone rip-ups which passes upward to matrix-supported conglomerate, then to sandstone displaying lateral infill, trough cross-bedding, and horizontal laminations as detailed below (Figure 19c). Graded-stratified units with both horizontal and inclined layers predominate. A-axes of the cobbles are typically imbricated, although a component of a-axes transverse to flow and b-axes imbricated is locally well-developed. Channel margins are steep, often "stair-stepped" and undercut (Figure 21c). Erosion was dominantly directed downward with little meandering, as the width-to-depth ratios of this channel conglomerate subfacies are much lower than those along the beach cliffs (typical range of 1 to 5 versus 10 or much greater along the seaciffs). This facies is found in the Miramar Road, lower Price Club and upper Morena Boulevard sections.

Channelized Sandstone: A wide variety of subfacies occur, being completely transitional from one to another. For ease of classification, three distinct types are defined, all slightly different from the channel sandstones of the beach cliffs but also displaying many of the same features. Perhaps the major distinguishing characteristic, as with the channel conglomerates, is the average lower width-to-depth ratios of these inland sandy channels, although this is not always
diagnostic. These subfacies also lack flutes, dewatering pipes and dish structures, convoluted and flamed laminae, and multiply-graded fills.

The channelized conglomerate described above is transitional to one of these channelized sandstone subfacies. Channel margins are very steep, often undercut or stepped. Basal cobble conglomerate with mudstone rip-ups passes upward to inclined clast-supported conglomerate stringers interlayered with coarse-grained structureless sandstone (Figure 19c). Cross-cutting pebble stringers and rip-ups attest to cut-and-fill of the lower sandstone intervals within the channels. In some rare cases, flat-based sheets of graded conglomerate pass laterally into matrix-supported inclined conglomerate. The overlying sandstone is yellowish gray to light gray, coarse- to medium-grained, sometimes displaying both cut-and-fill and faint trough cross-beds. This interval, which is minor or lacking in some channels, then passes up to horizontal planar laminae which may have minute mudstone clasts arranged as interlaminated stringers. The very upper portion of this type of channel is medium- to fine-grained sandstone with small "current-ripple" trough cross-laminae and/or a rare thin interval of small burrows or complete bioturbation. Mudstone interlaminations are also a feature of this uppermost sequence. Other channelized sandstones and conglomerates often cut down into the underlying channel, thus removing varying amounts of the upper sequence. Sandstone channels of this subfacies are present in the Miramar Road, upper Morena Boulevard, and lower Price Club sections, and display total thicknesses ranging up to 7 meters.

This channelized sandstone facies grades to a second subfacies
found at the Price Club, Pacific Theater, and Balboa Avenue localities. Unlike the conglomerate-based channels described above (which are cut into semi-consolidated material by numerous flows, resulting in steep margins and cut-and-fill structures), these sandstones were deposited on much less consolidated mudstone. Basal contacts display spectacular flame structures, mudstone pry-ups, and convolutions within the underlying material (Figure 19d). Channels of this subfacies are thinner, less than 3 to 5 meters thick, and very often less than 1 meter. The sandstone fill is typically graded, grayish yellow to light gray, medium- to fine-grained, with mudstone rip-ups in its massive base passing up to a laminated interval. The upper portion is composed of interlaminated sandstone and siltstone with very low-amplitude starved ripples or climbing ripples. *Ophiomorpha* burrows are sometimes present throughout this fill sequence. No trough cross-bedding or multiple cut-and-fill intervals were found, and unlike the conglomerate-floored sandstone channels, these forms were never observed to erode down into one another. Small burrows and complete bioturbation are also lacking.

A final channel type is extremely subtle, usually less than 2 meters deep and less than 10 meters wide (Figure 20a). These channels lack crystalline clasts, burrowing, flamed or deeply-eroded bases, grading, or a consistent vertically-changing sequence of sedimentary structures. Instead they may display a slightly irregular erosive base with mudstone rip-ups (Figure 20b) and pass upward into light-brown to yellowish-gray sandstone that is either: 1) coarse- to medium-grained, sometimes pebbly, structureless, and rarely amalgamated with mudstone-
Figure 20. Lithofacies exposed inland.

A) Channelized sandstone facies III. Note the very subtle channel margin (arrows) discordant to the horizontal bedding of the underlying mudstone.

B) Slightly irregular basal erosive contact of the channel-form sandstone in A (arrows). The sandstone fill is structureless, and contains abundant mudstone rip-up clasts.

C) Interlaminated sandstone, siltstone, and mudstone facies dominated by rippled layers and lenses. Flame structures due to loading and drag also occur. Opposing current senses are common in adjacent layers.

D) Alternating planar-laminated sandy siltstone and burrowed muddy siltstone layers. Dark plant fragments are common, highlighting both laminae and burrows.
clast lenses, or 2) medium- to fine-grained, laminated, with layers of whole to fragmented, monospecific shells such as Acila decisa (Conrad) or Turritella uvasana Conrad. Fill of some associated channels is of the same lithology as that encasing these subtle channelized sands, i.e. bioturbated, laminated sandy siltstone and mudstone. Mudstone slurried units filling similar subtle channel-forms are also associated with this channelized sandstone subfacies. These channels occur along Gilman and Clairemont Drives (no stratigraphic sections measured), and at the Soledad Canyon, Balboa Avenue, and Genesee Avenue South sites.

Interlaminated Sandstone, Siltstone, and Mudstone: This is the last facies of the beach-cliff transect which displays a significant subfacies variation inland. This subfacies is present only at the Miramar Road locality, and consists of silty to clayey, laminated medium-gray mudstone with rare thin to very thin, very-pale-orange, iron-stained, fine- to very fine-grained silty sandstone (Figure 20c). Many of the sandstone laminae have an undulatory (rippled) surface draped by the mudstone. Isolated siltstone ripples are common within the mudstone leading to lenticular bedding with thin, flat lenses of cross-laminated siltstone surrounded by mudstone. Dipping laminae with opposing (current) senses are common in adjacent ripple layers. Bioturbation is rare, though some simple, medium-sized (5 cm thick), vertical sand-filled burrows are locally present in low numbers. Shell fragments are absent.

Laminated to Burrowed Sandy Siltstone: This facies is different from the beach-cliff siltstone and silt-shale in that it displays alternating laminated and bioturbated sequences (Figure 20d). Planar-
laminated, pale-yellowish-brown to light-brown, very fine-grained sandstone is interlaminated with muddy very-light-gray to grayish-orange siltstone. Some small (0.5 to 2 cm) to medium (5 cm) sand-filled meandering to vertical burrows (Ophiomorpha, Planolites, and others) disrupt the laminae. Ripples are extremely rare. Minute plant fragments are abundant, and shell fragments are uncommon. These zones (5 to 40 cm thick) alternate with and grade upward into completely bioturbated sandy siltstone zones (usually thinner than 20 cm). Thin to very thin, fossiliferous to rippled sandstone beds are rare though present. These sequences are found at the Rose Canyon, Soledad Canyon, San Clemente Canyon Road, and Monangahela Drive sites.

**Pinch-and-Swell Sandstone:** Found only at the Miramar Road locality and along the railroad cuts east of the Soledad Canyon site, beds of pale-yellow-brown to light-gray, medium-grained sandstone rapidly swell from 0 cm up to 250 cm (Figure 21a). Laterally the beds are extremely discontinuous, usually pinching out within 10 meters maximum to about 2 meters minimum. Upper surfaces of these beds are undulatory and dip in all directions, then are draped by organic-rich claystone and interlaminated to thinly-interbedded sandstone and mudstone. Lower surfaces are flat to undulatory; when irregularly based, the sandstone bed usually is filling in a low area predisposed by the underlying pinch-and-swell sandstone bed's topography. Muddy intervals completely encase these coarser-grained units, filling in swales between beds and draping their swelling upper surfaces.

Internally, these sandstones are gently trough cross-bedded or planar laminated. Where cross-bedded, the sedimentary sequence passes
Figure 21. Sandstone lithofacies exposed inland.

A) Bar-like pinch-and-swell sandstone beds (arrows). Note the discontinuous nature of these units. Upper surfaces are draped with mudstone, and subsequent, overlying sandstone beds sequentially fill in topographic lows.

B) Turbidite sandstone beds present inland have thin planar-laminated (b) intervals and thick ripple cross-laminated (c) sequences. These Tbcė turbidites contain abundant organics highlighting the internal sedimentary structures.

C) The planar-laminated sandstone facies with "perched" cobbles is shown in the lower right, cut by a steep-sided conglomerate-filled channel.

D) The fine- to medium-grained planar-laminated sandstone of C contains isolated cobbles and cobble zones which display a thin surrounding scoured zone (arrows).
upward from planar-laminated medium- to fine-grained silty sandstone. The upper surface is burrowed by mud- and organic-lined Ophiomorpha and Gyrolithes, and in one case, contains a large irregularly-margined ellipsoidal structure resembling a brood pouch. Other upper surfaces are completely bioturbated. Mudstone rip-ups are aligned along sandstone laminae or at the bases of trough cross-bed sets.

**Turbidite Sandstone:** Present only at the Miramar Road and Soledad Canyon sites, 70 to 20 cm-thick beds of pale-yellowish-brown, fine- to very fine-grained sandstone are draped by 10 to 5 cm-thick silty to clayey, light-gray to brownish-gray mudstone. The beds are of Tbc sequences (Bouma, 1962). Within the sandstone, a subordinate basal horizontally planar-laminated unit passes upward to a thick climbing ripple cross-laminated (type 2 of Jopling and Walker, 1968) sequence, and then to an upper undulatory surface with asymmetrical, low-amplitude ripples (Figure 21b). This bed may in turn be draped by micrograded mudstone intervals on the order of 1 to 3 cm thick, or first pass through a sequence of lenticular bedding with starved silty sandstone ripples. The rippled-interval is highlighted by concentrations of macerated plant fragments on lee sides of ripples. Some small- to medium-sized, simple vertical burrows are present. Sandstone (clastic) dikes cutting the complete sequence are rare. The sandstone beds are somewhat lens-shaped, pinching out over a distance of 10's of meters.

**Conglomerate-Based Sandstone:** Medium to very thick, laterally continuous and non-channelized sandstone beds are interbedded with burrowed sandy siltstone and mudstone intervals. These take on two
different forms. At the Rose Canyon locality, slightly irregular- to flat-based beds of white to grayish-yellow, coarse- to medium-grained sandstone may contain or lack 20 to 50 cm-thick basal layers of disorganized crystalline clasts, mudstone rip-ups, and shell fragments. This passes upward into massive sandstone and/or trough cross-beds with isolated clasts. The uppermost portion may display laminated, medium- to fine-grained sandstone with stringers of mudstone clasts or rare thin mudstone laminae with flame structures. This interval then may gradationally become burrowed to completely bioturbated, with mud-lined vertical Ophiomorpha common. This complete sequence is interrupted in some beds due to the amalgamation of another conglomerate-based sandstone sequence, without an intervening siltstone/mudstone sequence.

Another form is present at the San Clemente site. While a detailed stratigraphic section was not measured, a repetitive sequence of interbedded conglomerate-based sandstone and burrowed silty mudstone was described. The sandstone beds have a disorganized basal layer of crystalline clasts and abundant thick-shelled, fragmented mollusk shells. This passes up into a thick sequence of very low angle (average of 5 to 10°) planar laminations arranged in slightly cross-cutting sets dipping in all directions. This type of bedding has been termed "truncated wave-ripple laminae" (Campbell, 1966). Although these sedimentary structures superficially resemble "hummocky cross-stratification", they do not display obvious positive-relief bedding associated with this stratification form (Harms et al., 1975; Hamblin and Walker, 1979). This sequence passes upward through a structureless interval, and finally a thin ripple trough cross-laminated uppermost unit. These beds
may be amalgamated, and lack portions of the upper sequence. Intervening beds are composed of interlaminated silty sandstone and siltstone with ripple cross-laminae, and burrowed interlaminated sandy siltstones and mudstone. Shell fragments in these finer-grained beds are rare, though present.

**Planar-Laminated Sandstone with "Perched" Cobbles:** Another facies found only at the Miramar Road locality, light-gray to yellowish-gray, fine- to very fine-grained sandstone, is planar laminated. These laminae dip about 5° and have isolated pockets of and solitary crystalline clasts and rare shell fragments (Figure 21c). The cobbles are up to 25 cm in length, but typically less than 10 cm along the a-axis. Pockets of cobbles contain fewer than 10 clasts and more often only two or three. These clasts are disorganized to oriented with a-axis parallel with one another but dipping in opposing directions. All cobbles display a small, thin scoured zone surrounding them, cutting into the sandstone laminae to a depth of approximately 1 cm or less (Figure 21d).

Some intervals are siltier, lacking abundant cobbles, and appear to have been rippled. However, a mottled appearance presumably due to complete bioturbation has destroyed sedimentary structures. Another inaccessible interval has trough cross-bedded sandstone at its very top. All of these beds are cut by the steep-margin conglomerate passing up to channelized sandstone.

**Cross-bedded Sandstone:** Steeply-dipping light-gray to yellowish-gray, pebbly granule to medium-grained sandstone contains multiple
steeply-inclined planar cross-bed sets which in some places become somewhat tangential at their base. Sets of cross beds are on the order of 1 to 5 meters thick. Gravel layers and mudstone rip-ups are inclined along the bedding surfaces. These dip 20 to 30°. This facies is found only in the Friars Formation at San Clemente Canyon.
Vertical Facies Transitions

Lithofacies Exposed Along The Beach Cliffs

Fining- and thinning-upward sequences characterize the vertical facies changes exposed along the Torrey Pines-Scripps sea-cliff transect as well as at Tourmaline Surfing Park. However, some coarsening- and thickening-upward trends are also present. As will be detailed in the discussion section, the fining-upward sequences typically begin with erosive coarse-grained sandstone or conglomerate channels that occur on a variety of scales. The sandstone beds become thinner and finer-grained upward, and intercalated mudstone or siltstone units become dominant. This is common in any kind of channel system in which there has been gradual abandonment and more passive deposition through time in response to the depositional site shifting elsewhere, the channel becoming plugged, or sedimentation rates decreasing. In the depositional system under study, these "positive" cycles (terminology of Ricci-Lucchi, 1975) often represent a channelized sequence associated with a submarine canyon and fan, and will be discussed later.

In contrast, "negative" or coarsening- and thickening-upward sequences are indicative of progradational (regressive) trends. Both positive and negative cycles occur as organized groups of beds within a much larger suite or "first order cycle", i.e. a basin-margin suite. To distinguish between the two types of cycles, Mutti and Ricci-Lucchi (1972) termed the second order (internal cycles) "megasequences". This terminology will be followed below. Some sections lacking definable megasequences also occur.
Amalgamated pebbly sandstone, massive to laminated sandstone, interbedded sandstone and siltstone, and mudstone facies occur as fining-upward megasequences in Torrey Pines Reserve Canyon, Bathtub Rock Trail, and Canyons #1, #2, and #3. The amalgamated pebbly sandstone cuts into underlying Delmar Formation and/or Torrey Sandstone. Although the amalgamated pebbly sandstone and laminated sandstone facies have previously been mapped as Torrey Sandstone (Kennedy, 1975), an irregular, angular unconformity cutting across Delmar and Torrey bedding surfaces distinctly separates them from the overlying amalgamated pebbly sandstone. Sedimentary and biogenic structures show a dramatic change across this boundary (see Boyer and Warme, 1975, for a discussion of "typical" Torrey Sandstone depositional styles), and the overlying mudstone facies contains microfossils indicating concomitant abrupt variations in age and depth of deposition. The details and significance of this surface are outlined later. Some interbedding of these facies occurs, forming short, repetitive positive megasequences, and the sandstone/siltstone and mudstone facies are not both present in all these sections.

In Canyon #1, the mudstone facies in the 22.5 to 36 meter interval contains no discernible megasequence, then passes upward into various coarsening-upward and fining-upward intervals caused by the variable development of the interlaminated sandstone, siltstone, and mudstone and the fossiliferous sandstone and mudstone facies. This pattern is also present in Canyon #3. Cross-bedded sandstone and/or burrowed silty sandstone facies in the upper portions of these sections continue the uppermost coarsening-upward trends.
The remaining beach-cliff sections contain only fining-upward megasequences. Canyon #4 contains a massive channelized system approximately 1 km wide which begins with approximately 27.5 meters of amalgamated clast-supported conglomerate and some rare matrix-rich conglomerate passing upward into massive to laminated sandstone and finally into laminated siltstone with medium beds of massive to laminated sandstone. A number of shorter positive megasequences occur, some also beginning with the conglomeratic facies. All these smaller megasequences contain channelized sandstones which are amalgamated or isolated, and pass upward to thinly-interbedded/lenticular-bedded sandstone and mudstone, interlaminated sandstone, siltstone, and mudstone, laminated siltstone, and slurried unit facies; channelized and massive sandstone interbeds decrease in thickness up through each megasequence. Canyons #5 and #6 contain similar cycles, although mudstone and pebbly mudstone facies also occur.

The cliff at Tourmaline Surfing Park is significant in having only one major positive megasequence that begins with coarse conglomerate, and displaying a preponderance of interlaminated sandstone, siltstone, and mudstone facies with "lam-scram" sequences. Mudstone laminae are distinctly subordinate; rippled siltstone and sandstone laminae are ubiquitous. Relatively flat-based, but laterally-discontinuous, cross-bedded sandstone occurs within the sandstone/siltstone/mudstone facies. An association of channelized sandstone and slurried units, along with interlaminated sandstone, siltstone, and mudstone and lenticular-bedded sandstone and mudstone, characterize a minor, second positive mega-
sequence. Thus, these units are much like those of upper Canyon #4 and Canyons #5 and #6, with the decrease in channelized sandstone and increase of interlaminated sandstone and siltstone being the major difference.

Lithofacies Exposed Inland

The Torrey Pines Road, Monangahela Drive, and Balboa Avenue sections all have facies and negative megasequences similar to the upper portions of Canyons #1, #2, and #3. However, the Balboa Avenue locality contains abundant intervals where no megasequences are definable, and also has some superimposed short positive megasequences due to the presence of the subtle channelized sandstone subfacies. Similar channelized sandstones create fining-upward trends in overall coarsening-upward stratigraphic sections at Genesee Avenue South and Soledad Canyon.

Channelized sandstone facies also occur at the Price Club and Pacific Theater sites. These channels differ from those at the sites listed above not only in their internal sedimentary and biogenic structures (as described in the previous 'Facies Descriptions' section), but they also occur as part of single positive megasequences over the whole of each exposure. Single positive megasequences evident over the complete outcrop are also present at the Morena Boulevard and Genesee Avenue North localities. Within the field area, single large fining-upward trends such as these are subordinate to multiple positive megasequences related to discrete channel-fill events or to coarsening-upward sequences associated with progradation. As will be discussed later, these thick, continuous positive megasequences are related to transgressive depositional cycles.
The Genesee Avenue North positive cycle is extremely similar to the lowermost large channel-fill sequence in Canyon #4, with amalgamated, disorganized clast-supported conglomerate passing upward to massive to laminated sandstone and siltstone. Morena Boulevard is also composed of conglomerate changing upward to sandstone and finally interlaminated sandstone, siltstone, and mudstone. The major difference is that the conglomerates are of different age (Scripps Formation at Genesee Avenue North and Mount Soledad Formation at Morena Boulevard), and the conglomerate at Morena is amalgamated, normally and inverse-to-normally graded and has less matrix material.

This conglomerate subfacies present at Morena Boulevard is also found in Rose Canyon, where it cuts down into a short coarsening-upward megasequence of bioturbated silty sandstone and cross-bedded sandstone. This in turn overlies a much thicker fining-upward megasequence composed of conglomerate-based sandstone and laminated to burrowed sandy siltstone.

Steep-walled conglomerate and sandstone channel facies cut into laminated sandstone with "perched" cobbles at Miramar Road, then form a positive megasequence as they are overlain by pinch-and-swell sandstone. The laminated sandstone with "perched" cobbles is the culmination of one negative megasequence, and occurs above interlaminated unfossiliferous rippled sandstone, siltstone, and mudstone. A second negative cycle is present in the upper half of this section. Interlaminated sandstone, siltstone, and mudstone facies grades upward into turbidite sandstone, then is cut by steep-margined sandstone and finally
conglomerate channels.

At the remaining locality (San Clemente Canyon Road), an overall coarsening- and thickening-upward trend is present. Laminated to burrowed sandy siltstone becomes interbedded upward with thin beds of conglomerate-based sandstone. The sandstone units thicken upward, some becoming amalgamated. A thick sandstone with steeply-inclined sets of planar cross-beds is present in the upper quarter of the section. This cross-bedded sandstone is part of the Friars Formation.

Lateral Facies Changes

Torrey Pines-Scripps Beach-Cliff Transect

Lithofacies exposed along the Torrey Pines-Scripps and Tourmaline sea-cliff transects are graphically portrayed on the three foldouts of Appendix III. Mudstone, siltstone, sandstone, pebbly sandstone, crystal-line-clast conglomerate, intraclast conglomerate, and slurried unit distribution is shown. Also portrayed is an interpretation of the depositional environments represented by these units (which is discussed in a later section).

As already mentioned, relatively flat-lying Delmar Formation and Torrey Sandstone are cut by an undulating erosional unconformity which separates them from the overlying amalgamated pebbly sandstone. This irregular surface rises from sealevel just south of Bathtub Rock to an elevation of 54.5 meters over a distance of approximately 0.65 km in the Torrey Reserve Canyon. The average slope of this unconformity is thus a little over 4.75°. The basal portion of the amalgamated pebbly sandstone is typically clast-rich, composed of abundant mudstone rip-
ups with very irregular to rounded margins. This facies is present south to Canyon #3.

Both the amalgamated pebbly and laminated to massive sandstones are irregular and discontinuous, locally cut out by very broad mudstone-filled channels of the overlying Ardath Shale. The underlying Torrey and Delmar units also display much relief where cut into by the pebbly sandstone and Ardath Shale. The laminated to massive sandstone is interbedded in places with Ardath Shale, and laterally pinches out within it.

While most channels within the Ardath are filled with mudstone and contain varying amounts of thin graded, rippled, cross-bedded, fossiliferous, and/or bioturbated sandstone and siltstone layers, some have a draped mudstone base, then are "plugged" with laminated sandstone. One example is above the Torrey Pines Landslide. Bases of many broad channels (to 100's of meters wide and 10's of meters thick) are lined with fossiliferous, concretionary sandstone, which outlines their channel-form.

The base of the Ardath Shale mudstone-filled channels between Bathtub Rock and Canyon #2 rises high in the section at Canyon #3; these channels dominate the complete cliff face below the Hang Glider Port. Thick siltstone and sandstone units cap the channel sequences between Canyons #2 and #3 and along the Hang Glider Port. Sandstone also becomes somewhat more prevalent within the mudstone-filled channels along the Glider Port cliff. These sandy fills are typically massive to laminated sandstone facies with rare intraclast-rich intervals,
whereas the capping units along the upper cliffs are interbedded bioturbated sandstone and siltstone and fossiliferous interbedded sandstone and mudstone facies. The channels within the Ardath not only become somewhat sandier, but also are overall shallower and smaller scale along the southern Hang Glider Port cliff face. The Scripps Formation is faulted up into contact with the Ardath Shale south of the Glider Port, and marks the first occurrence of clast-supported conglomerate and exceptionally thick, massive- to laminated-sandstone sequences. Overlying this thick positive megasequence north of Canyon #4, the sandstones lose their channelized form, and interbedded sandstone and siltstone similar to that further north occurs.

Between Canyons #4 and #5, slurried units, turbidite sandstone, and somewhat thinner, multiple sequences of conglomerate and sandstone channel-fill are present. The margin of the very large channel at Canyon #4 rises above this sequence, implying that its development occurred later than the lower units further south.

To the south, deep channels of coarse-grained fill and well-developed megasequences give way to broader and flatter units, some of which are laterally continuous over the whole outcrop distance. However, within any one coarse-grained interval, intense cut-and-fill is evident. Because of the intense channeling and hence lateral discontinuity of any one bed (versus a laterally-continuous coarse-grained interval), a vertically-measured section is representative of only that one specific location and is correlatable in only a general sense laterally (Figure 22). Intervening siltstone and silty mudstone
Figure 22. Correlation of measured sections, Torrey Pines-Scripps beach-cliff transect. The stratigraphic sections are vertically restored to their relative pre-faulting positions.
units separate these coarse-grained units except just north of Scripps Institution of Oceanography, where the complete amalgamation of conglomerate and sandstone channels has removed all fine-grained layers. The upward progression of coarse-grained to silty to muddy layers produces numerous short positive megacycles. Slurried units are common in this southern portion, especially in the uppermost Ardath Shale just below Scripps Formation deposits. The largest olistostrome occurs at Canyon #5.

In summary, a major erosional unconformity with much topographic relief separates flat-lying Delmar and Torrey units from a channelized system present along most of the transect. Interbedded pebbly and laminated sandstone is cut by and interbedded with extremely large channels containing fine-grained fill of the Ardath Shale. This is capped by fossiliferous sandstone and siltstone units present from Canyon #1 to the southern margin of the Hang Glider Port. At that point, a 1 km-wide Scripps Formation channel with a thick positive megasequence cuts into the Ardath Shale. This channel also overlies a somewhat older Scripps Formation channelized system composed of a wide array of cut-and-fill sequences on a variety of vertical and lateral scales. All these channels show fining- and thinning-upward cycles; amalgamation creates multiple positive megasequences. This channelized system in turn overlies a turbidite and mudstone Ardath Shale sequence capped by a horizon of large, relatively continuous slurried units.

Tourmaline Surfing Park Beach-Cliff Transect

Because of the dipping beds, the lateral facies transitions present along the Tourmaline Surfing Park cliffs were essentially outlined
within the 'Vertical Facies Transitions' section. Abundant interlayered slurried units and cross-bedded or channelized sandstone are separated by intervening interlaminated siltstone/sandstone. Both the slumps and sandstone beds pinch out toward the main clast-supported conglomeratic channel axis and thicken away from it. These units dip slightly away from the conglomerate even after accounting for tectonic tilting.

Laterally, the sequence becomes finer-grained. Syndepositional slumps and thick lenticular-bedded sandstones give way to thinly-bedded siltstone and sandstone, and then to bioturbated mudstone with starved sandy ripples. The irregular microtopography on top of slump units is mimicked along the bases of overlying sandstones even though separated by draped, interlaminated sandstones and siltstones. This transect resembles some of the thin channelized cycles south of Canyon #5, although the intense multiple cut-and-fill along the southern Torrey Pines-Scripps transect is absent at Tourmaline.

Regional Correlations

Although regional correlations of the depositional environments are detailed in the discussion section, it is relevant here to mention certain similar facies patterns (see Figures 49 and 50). The Genesee Avenue North section contains a positive, channel-fill megasequence closely matching that of Canyon #4. The same stratigraphic relationships hold inland as along the coast, and this cycle overlies the lower portion of Genesee Avenue South. This latter locality has a negative megasequence truncated by fining-upward channelized sandstone to sandy siltstone facies, closely akin to the pattern present at Soledad Canyon.

The fossiliferous interbedded sandstone and mudstone facies and
interbedded sandstone and siltstone facies capping the Ardath Shale from Canyon #1 to Canyon #4 occupy similar stratigraphic positions at Monangahela Drive, Torrey Pines Road, and Balboa Avenue. Older Mount Soledad Formation clast-supported conglomerate at Morena Boulevard forms a single positive megasequence that appears to be relatively continuous into and directly underlying positive megasequences occupying the complete stratigraphic outcrop at the Pacific Theater and Price Club localities.

Further inland, negative megasequences are more common. Crossbedded sandstone caps a sandy siltstone to conglomerate-based sandstone interbedded with sandy siltstone transition at San Clemente Canyon Road. Clast-supported conglomerate cuts into crossbedded sandstone and bioturbated silty sandstone at Rose Canyon; this overlies a positive cycle of conglomerate-based sandstone passing upward to sandy siltstone. Finally, conglomerate caps another negative megasequence at Miramar Road. As at Rose Canyon, this overlies a fining-upward cycle which in turn overlies a coarsening-upward trend.
TEXTURAL ANALYSIS

Friable samples were studied texturally by use of a rapid, automated settling-tube system as described in the methods section. Wet sieving of the total sample separates sand-sized (>63μ) material from the mud, then the clays and fine silts are decanted from this fine fraction leaving behind coarse- to medium-grained silt (63 to 16μ). Dry weighing prior to and following this process yields sand:silt:clay ratios. The textural parameters - moment measures of mean grain size and standard deviation and graphic measures of skewness and kurtosis - as well as cumulative and frequency curves are calculated by computer.

The mean (average) grain size (M_φ) of a sample, using the method of moments, is affected by the total weight distribution. For each phi (ϕ) interval, that group's phi size midpoint (D) is multiplied by the total weight of that interval (W). This is summed for the entire sample and divided by the total sample weight to give the moment measure of mean:

\[ M_\phi = \frac{\sum DW}{\sum W}. \]

The standard deviation (σ_ϕ) is a measure of the spread about the mean, i.e. sorting (Figure 23A) and is determined by:

\[ \sigma_\phi = \frac{\sum W(M_\phi - D)^2}{\sum W}. \]

This number is comparable to the following verbal descriptions:

\[ \sigma_\phi = 0.35 \phi \text{ or less} \quad \text{very well sorted} \]
\[ 0.35 - 0.50 \phi \quad \text{well sorted} \]
\[ 0.50 - 0.71 \phi \quad \text{moderately well sorted} \]
Figure 23. Relationships between statistical output and graphic displays in textural analysis (from Folk, 1974).

A) Frequency Curve - in essence, a smoothed histogram showing weight percent for each data point. In the "normal" distribution shown, most grain-sizes cluster about the mean and "tail off" symmetrically to the coarse- (lower \( \phi \)) and fine-grained (higher \( \phi \)) extremes.

B) Cumulative Frequency Curve - showing same data as above. The weight percent of each data point is added to the preceding points to yield a running total.

C) Frequency curves a and b are equally sorted and have an equal standard deviation. However, they have different means and a is positively skewed while b is negatively skewed.

D) Frequency curve a is extremely leptokurtic, excessively peaked in the middle and deficient in the "shoulders". Curve b is extremely platykurtic as well as bimodal.
0.71 - 1.00 $\phi$ moderately sorted
1.00 - 2.00 $\phi$ poorly sorted
2.00 - 4.00 $\phi$ very poorly sorted
4.00 $\phi$ or greater extremely poorly sorted

The inclusive graphic skewness (SkI) is a measure of a sample's deviation from a symmetrical grain-size distribution (a "normal" or bell-shaped frequency curve), and in which direction it "tails-off". It is determined by measuring the displacement of the median from the average $\phi$ size taken at the 16 and 84 cumulative weight percent points, and also the average between the 5 and 90 cumulative weight percent points (Figure 23B and 23C):

$$
SkI = \frac{\phi_{16} + \phi_{84} - 2\phi_{50}}{2(\phi_{84} - \phi_{16})} + \frac{\phi_{5} + \phi_{95} - 2\phi_{50}}{2(\phi_{95} - \phi_{5})}
$$

The verbal descriptors are:

SkI = +1.00 to +0.30 strongly fine-skewed (tail to the right)
+0.30 to +0.10 fine-skewed
+0.10 to -0.10 near-symmetrical
-0.10 to -0.30 coarse-skewed
-0.30 to -1.00 strongly coarse-skewed (tail to the left)

Graphic kurtosis (Kg) is also a measure of a sample's deviation from a Gaussian (normal) weight distribution, and compares the sorting in the tails to the sorting in the central portion (Figure 23D). A very peaked curve is better sorted in the central portion, and a flat-peaked curve has better sorting in the tails. Mathematically, the spread between the grain sizes at the 25 and 75 cumulative weight percent points is compared to those at the 5 and 95 cumulative weight
percent points:

\[ K_G = \frac{\phi_{95-\phi5}}{2.44(\phi_{75-\phi25})} \].

For a verbal comparison, the following are used:

- \( K_G = 0.67 \) or less \quad \text{very platykurtic}
- \( 0.67 - 0.90 \) \quad \text{platykurtic (flat-peaked)}
- \( 0.90 - 1.11 \) \quad \text{mesokurtic (Gaussian distribution)}
- \( 1.11 - 1.50 \) \quad \text{leptokurtic (high-peaked)}
- \( 1.50 - 3.00 \) \quad \text{very leptokurtic}
- \( 3.00 \) or more \quad \text{extremely leptokurtic}

A variety of the sandstone facies were analyzed (Table 2, Figures 24 to 36). The results for different types of facies were grouped to distinguish broad patterns and natural sedimentological breaks, which are then used to interpret depositional mechanisms and environments in the discussion section. It should be emphasized that a specific set of textural parameters is not unique to any one specific depositional environment. Rather, these parameters reflect given hydrodynamic conditions acting upon a uniquely-defined source material. Thus the processes operative in transporting and depositing sediments, and their relationships in time and space, are expressed by the textural data.

In an effort to construct a qualitative sand budget delineating sediment storage, transport, and transformation through the study area, it is necessary to define the type of source material. To characterize this detrital input, sandstones deposited within the Eocene fluvial systems were processed (samples from the Ballena Channel and fluvial Torrey Sandstone channels north of the field area in Encinitas) (Figure
<table>
<thead>
<tr>
<th>Sample</th>
<th>Mean Grain Size</th>
<th>Standard Deviation</th>
<th>Skewness</th>
<th>Kurtosis</th>
<th>Sand:Silt:Clay</th>
</tr>
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<tbody>
<tr>
<td>36</td>
<td>2.83</td>
<td>0.54</td>
<td>-0.07</td>
<td>1.43</td>
<td>84:7:9</td>
</tr>
<tr>
<td>38</td>
<td>3.19</td>
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<td>81:9:10</td>
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<td>0.76</td>
<td>0.37</td>
<td>1.37</td>
<td>95:2:3</td>
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<td>0.90</td>
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</tr>
<tr>
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<td>1.23</td>
<td>41:14:45</td>
</tr>
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<td>72</td>
<td>2.59</td>
<td>0.59</td>
<td>-0.10</td>
<td>1.09</td>
<td>58:10:32</td>
</tr>
<tr>
<td>82</td>
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<td>0.17</td>
<td>0.96</td>
<td>94:3:3</td>
</tr>
<tr>
<td>86</td>
<td>2.04</td>
<td>0.63</td>
<td>0.17</td>
<td>0.96</td>
<td>95:3:2</td>
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<td>94</td>
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<td>0.00</td>
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<td>218</td>
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<td>1.43</td>
<td>80:10:10</td>
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<tr>
<td>222</td>
<td>1.53</td>
<td>0.84</td>
<td>0.05</td>
<td>0.90</td>
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</tr>
<tr>
<td>226</td>
<td>3.29</td>
<td>0.50</td>
<td>0.00</td>
<td>1.23</td>
<td>75:11:14</td>
</tr>
<tr>
<td>304</td>
<td>2.35</td>
<td>0.74</td>
<td>0.13</td>
<td>0.89</td>
<td>86:9:5</td>
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TABLE 2. Textural Analysis For Sand-sized Fractions (cont.)

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mean Grain Size</th>
<th>Standard Deviation</th>
<th>Skewness</th>
<th>Kurtosis</th>
<th>Sand:Silt:Clay</th>
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</thead>
<tbody>
<tr>
<td>307</td>
<td>2.44</td>
<td>0.94</td>
<td>-0.06</td>
<td>0.90</td>
<td>88:5:7</td>
</tr>
<tr>
<td>309</td>
<td>2.67</td>
<td>0.86</td>
<td>-0.04</td>
<td>1.14</td>
<td>88:5:7</td>
</tr>
<tr>
<td>311</td>
<td>2.67</td>
<td>0.82</td>
<td>-0.13</td>
<td>1.04</td>
<td>82:7:11</td>
</tr>
<tr>
<td>313</td>
<td>2.36</td>
<td>0.91</td>
<td>0.12</td>
<td>0.95</td>
<td>79:6:15</td>
</tr>
<tr>
<td>314</td>
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<td>0.90</td>
<td>-0.04</td>
<td>1.58</td>
<td>90:5:5</td>
</tr>
<tr>
<td>321</td>
<td>1.98</td>
<td>0.97</td>
<td>0.26</td>
<td>1.07</td>
<td>86:5:9</td>
</tr>
<tr>
<td>326</td>
<td>2.45</td>
<td>1.11</td>
<td>-3.84</td>
<td>0.28</td>
<td>82:6:12</td>
</tr>
<tr>
<td>329</td>
<td>3.21</td>
<td>0.38</td>
<td>0.23</td>
<td>0.89</td>
<td>76:10:14</td>
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<tr>
<td>340</td>
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<td>0.95</td>
<td>-0.10</td>
<td>1.01</td>
<td>92:3:5</td>
</tr>
<tr>
<td>341</td>
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<td>1.00</td>
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<td>0.94</td>
<td>93:3:4</td>
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<tr>
<td>344 f.l.</td>
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<td>0.17</td>
<td>0.97</td>
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<tr>
<td>345 B</td>
<td>3.16</td>
<td>0.36</td>
<td>0.32</td>
<td>0.99</td>
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<tr>
<td>349</td>
<td>3.14</td>
<td>0.41</td>
<td>0.17</td>
<td>0.93</td>
<td>82:9:9</td>
</tr>
<tr>
<td>Sample</td>
<td>Mean Grain Size</td>
<td>Standard Deviation</td>
<td>Skewness</td>
<td>Kurtosis</td>
<td>Sand:Silt:Clay</td>
</tr>
<tr>
<td>--------</td>
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<td>-------------------</td>
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</tr>
<tr>
<td>350</td>
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<td>1.02</td>
<td>-4.20</td>
<td>0.41</td>
<td>86:5:9</td>
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<tr>
<td>351</td>
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<td>0.15</td>
<td>1.21</td>
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</tr>
<tr>
<td>0-8</td>
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<td>0.45</td>
<td>-0.07</td>
<td>1.43</td>
<td>82:6:12</td>
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<tr>
<td>Encb</td>
<td>2.34</td>
<td>0.55</td>
<td>-0.07</td>
<td>1.43</td>
<td>79:5:16</td>
</tr>
<tr>
<td>RC-D-1</td>
<td>1.83</td>
<td>0.70</td>
<td>0.27</td>
<td>1.23</td>
<td>88:5:7</td>
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<tr>
<td>RR-I-2</td>
<td>1.74</td>
<td>0.63</td>
<td>0.16</td>
<td>1.84</td>
<td>92:2:5</td>
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<td>RR-H-2</td>
<td>2.36</td>
<td>0.60</td>
<td>0.23</td>
<td>1.09</td>
<td>85:6:9</td>
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</tbody>
</table>
24). This material is very immature, containing the highest silt and clay components of any sandstone. The frequency curve for the sand fraction is negatively skewed, and would be even more strongly skewed if the associated pebbles and gravels were included. The coarse-grained tail between 1.25 and 1.75 $\phi$ represents the channel-lag component. The two strong peaks coarser than 3 $\phi$ indicate separate modes of tractional deposition, while the finely-skewed tail is the result of deposition from intermittent and/or total suspension (Visher, 1969; Reed et al., 1975). The lack of material between 1.25 $\phi$ and the associated gravels is somewhat puzzling, but may have resulted in a number of ways, including: 1) mechanical breakdown of coarse-grained lithic components, 2) chemical breakdown of unstable lithic components on outcrop, and/or 3) final filling of the fluvial system only during later erosional stages when the transport gradient was much reduced (J. Anderson, personal communication, 1981). Therefore the source material brought into the depositional system under investigation was very poorly sorted, with all detrital size-ranges probably represented.

Sandstone lenses within Mount Soledad and Stadium conglomerates are portrayed in Figure 25. Little residual lag is present. Instead, the frequency curves are strongly positively skewed (average of 1.43) with a sharp truncation at about 1 $\phi$ due to a predominance of tractive depositional processes. The sharp peaks in the fine tail of number 304 and 321 probably indicate other traction modes formed by waning deposition with sporadic pulses (intermittent suspension). The smooth fine tail of samples RC-D-1 and RR-I-2 denotes more rapid deposition from a some-
Figure 24. Frequency and cumulative curves for samples from Eocene fluvial systems. Sample 70 is a sandstone matrix from the basal conglomerate of the Ballena Channel west of the study area, while sample 72 is sandstone from the channel's main fill. Sample ENC-B is from small cross-cutting fluvial channels north of the study area in Encinitas. Lag (coarse-grained tail), traction (central modes), and suspension (fine-grained tail) components are all present. The fine-grained constituents (silt and clay) are the highest in these 3 samples.
MEAN GRAIN SIZE
2.52 $\phi$

SAND:SILT:CLAY
59:10:31

SAMPLE NUMBERS
72
ENC-B
70
Figure 25. Frequency and cumulative curves for sandstone lenses in the Stadium (RR-I-2) and Mount Soledad (304, 321, and RC-D-1) conglomerates. A lag component is lacking, with modes representing tractional and intermittent suspension deposition dominating. This produces a strong positively-skewed frequency curve.
MEAN GRAIN SIZE
1.98 φ

SAND:SILT:CLAY
88:5:7

SAMPLE NUMBERS
RR-I-2 ———
304 ·····
321 ———
RC-D-1 o o o
what instantaneously-decreasing flow. All of these had a consistent high energy mode which formed the break at 1 \( \phi \); this indicates physical removal from any lag-type depositional subenvironment.

Cross-bedded sandstone (46 and 222) is much "cleaner" (less fine-grained component) than the previously described samples, probably due to active winnowing of fines. These sands also contain coarser-grained material than the prior two types, and have a small amount of lag material (Figure 26). Although the frequency curves are somewhat positively skewed, they are truncated at 3 \( \phi \), further evidence for active winnowing. The strong bimodality of the tractional component emphasizes the coarse- and medium-grained laminated nature of these deposits.

Sandstones of pinch-and-swell and conglomerate-based facies (inland) also display some bimodality, but frequency curves are shifted to between 1 and 4 \( \phi \) (Figure 27). These are much less positively skewed to unskewed; both residual and intermittent suspension plus suspension components are evident. These sandstone are also less "clean" than the cross-bedded group, with an average of only 80.5% sand-sized material, and overall the most poorly-sorted sands of any group (average standard deviation of 1.01 \( \phi \)). If the finer-grained material had been analyzed, these sands probably would not display any truncation in the fine tail and hence be even more positively skewed.

Subtle, laminated to structureless channelized sandstone (inland facies) falls within the same grain-size interval as the pinch-and-swell and conglomerate-based sandstone facies (Figure 28). However, these contain less fine-grained material, the sand component is better
Figure 26. Frequency and cumulative curves for cross-bedded sandstone samples from the Torrey Sandstone (46) and Ardath Shale (222). Lack of material finer-grained than 3 φ is indicative of winnowing in a high-energy environment. The small weight percent of silt and clay also are evidence of hydrodynamic "cleaning" of the detrital material deposited within this setting.
MEAN GRAIN SIZE
1.34 φ

SAND:SILT:CLAY
94.5:2.5:3

SAMPLE NUMBERS
46
222
Figure 27. Conglomerate-based (313) and pinch-and-swell (326) sandstone samples produce frequency and cumulative curves with a bimodality like that from the cross-beded sandstones. However, these curves are "shifted" to the finer-grained sandstone modes, are more positively skewed, and display the poorest sorting of any samples analyzed. These features indicate lower-energy deposition and possibly a lack of availability of coarser-grained sandstone during the depositional phase.
MEAN GRAIN SIZE
2.41 \( \phi \)

SAND:SILT:CLAY
80.5:6:13.5

SAMPLE NUMBERS
313 ---
326 -----
Figure 28. Subtle channelized sandstone samples from the Scripps Formation (307, 314) and Ardath Shale (309, 311) fall in the same sand-sized range as the pinch-and-swell and conglomerate-based units. However, their frequency and cumulative curves indicate much better sorting and more dominant tractive deposition.
MEAN GRAIN SIZE
2.47 \( \phi \)

SAND:SILT:CLAY
87:5.5:7.5

SAMPLE NUMBERS
309
307
314
311
sorted (average standard deviation of 0.88 $\phi$), and discrete traction modes are much more in evidence. This tractive deposition is especially brought out by the frequency curve of sample 314. The remaining channelized sandstones of this group are somewhat more negatively skewed (average of -0.08), indicating a minor coarse lag component present within this system.

The second most poorly sorted material is the amalgamated pebbly sandstone facies (average standard deviation of 0.95 $\phi$). The frequency curves for these samples are negatively skewed with an average of -0.14, indicating abundant residual material (Figure 29). All of these sands display a degree of truncation at 3 $\phi$, which indicates that a winnowing process might have been associated with deposition. Some incipient grading might also be present, as numbers 48 to 340 to 341 were sampled in that order up through one pebbly sandstone unit. There appears to be a concomitant shift in their respective cumulative curves to somewhat finer grades. However, any grading present is very poorly developed.

The laminated sandstone facies which overlies and is interbedded with the amalgamated pebbly facies in contrast is: 1) moderately well sorted, 2) contains much more fine-grained material, 3) is positively skewed, and 4) has an extremely distinct traction mode which begins at approximately the point where the amalgamated pebbly sands show effects of winnowing (Figure 30). The presence of some mudstone laminae within this facies indicates that these samples are even more positively skewed than shown. As previously discussed, the fine-grained tail indicates deposition from suspension. Although closely related stratigraphically,
Figure 29. Frequency and cumulative curves for the amalgamated pebbly sandstone samples attest to the poorly-sorted nature of this facies. The lag (residual) component is very strong. A minor amount of deposition from suspension is indicated, possibly due to a winnowing process having removed the fine-grained component.
MEAN GRAIN SIZE
1.86 Φ

SAND:SILT:CLAY
93:3:4

SAMPLE NUMBERS
340
341
48
Figure 30. Frequency and cumulative curves for the laminated sandstone facies contrast sharply with those from the underlying amalgamated pebbly sandstone (Figure 29). Sorting is much better, one very distinct tractional mode predominates, and no lag component is present. Of special interest is that the finer-grained material deficient within and winnowed(?) from the amalgamated pebbly sandstone composes the laminated sandstone's grain sizes.
MEAN GRAIN SIZE
3.17 \( \phi \)

SAND:SILT:CLAY
79:8:13

SAMPLE NUMBERS
345B
344f1
these two facies provide evidence for vastly different depositional mechanisms.

Deposition from traction is also evident in another positively-skewed (average of 0.17), but not as well-sorted, massive to laminated sandstone present at Canyon #4 (Figure 31). However, these sandstones are much "cleaner", with 94.5% sand-sized material, and like the amalgamated pebbly sandstone, the fine tails of the frequency curves are truncated at about 3 φ. There appears to be a small coarse-grained lag component, again typical of channelized deposition.

Sandstone facies from smaller channels filled with laminated to apparently graded-sequences occur south of Canyon #4. One of these apparently graded-intervals in an amalgamated sequence within Canyon #5 was sampled from the base to the top every 10 cm. These samples in order are 349, 350, 351, and 38. This interval becomes better sorted, with a small residual component occurring only near the base, producing a somewhat coarse-tail graded sequence upward (Figure 32). Traction deposition predominated. Probably the coarse- and fine-grained materials were deposited together early in the depositional cycle; then in the late stages of flow only the fine material was present to be deposited from traction and suspension. This deposit has the highest percentage of fine-grained material of any channelized system, and the finest mean grain size (3.01 φ average).

Isolated sandstone beds within the mudstone facies and interbedded/ lenticular-bedded sandstone and mudstone facies are also fine-grained and texturally similar to 350 and 351 (Figure 33). Samples 36, 94, 218, and 0-8 all contain a small lag component, have almost equivalent
Figure 31. Massive to laminated sandstone from Canyon #4 produces frequency and cumulative curves dominated by tractional modes. Any lag component is very deficient, and possible winnowing effects may account for the sharp decrease in detritus finer-grained than 3 $\phi$. 
MEAN GRAIN SIZE
2.05 $\phi$

SAND:SILT:CLAY
94.5:3:2.5

SAMPLE NUMBERS
82 ——
86 ·····
Figure 32. Frequency and cumulative curves from laminated sandstone of small channels display good sorting and dominant tractional deposition. The basal portions of apparently graded sequences contain a lag component that is lacking higher up in each sedimentation interval.
MEAN GRAIN SIZE
3.02 \( \phi \)

SAND:SILT:CLAY
84:7:9

SAMPLE NUMBERS
38
349
350
351
Figure 33. Isolated sandstone beds within the mudstone facies (0-8, 218) and interbedded/lenticular-bedded sandstone and mudstone facies (36, 94) display frequency and cumulative curves much like those of the basal portions of small sandstone channel sedimentation units (Figure 32). Tractive deposition dominated, but lag and suspension components are also present.
MEAN GRAIN SIZE
2.94 \( \phi \)

SAND:SILT:CLAY
82.5:7.5:10

SAMPLE NUMBERS
36
94
0-8
218
mean grain sizes, and contain approximately the same amounts of sand: silt:clay, though slightly more fine-grained component. However, the internal sedimentary structures of these facies, the bedding thickness, and the associated lithologies are very different. But even considering these morphological differences, the flow conditions under which deposition took place were probably quite similar.

Samples 226 and 329 of bioturbated silty sandstone (Figure 34) are texturally quite similar to laminated sandstone samples 344 f.1. and 345 B. They are only slightly finer-grained (3.25 φ versus 3.17 φ averages) and have relatively equivalent sorting values and sand:silt:clay ratios. However, a minor lag component is evident in the bioturbated silty sandstone frequency curves, while a very sharp traction mode defines the depositional style of the coarsest laminated sandstone component. Both types of deposits are positively skewed, probably more so if finer-grained material were included. Again, although these two facies differ morphologically, the processes responsible for their deposition were similar.

The final facies analyzed is laminated sandstone with "perched" cobbles (Figure 35). This unit is moderately well sorted, positively skewed, with deposition from traction and suspension dominant. A minor residual portion may be present. It appears that two main modes may compose the frequency curve, one at 2 φ and one at 2.75 φ, with the finer-grained one masked due to overlap. This conjecture is borne out in the field as this deposit displays planar laminations and an abundant fine-grained component.
Figure 34. Frequency and cumulative curves from samples of bioturbated silty sandstone closely resemble those from the laminated sandstones (Figure 30). A minor lag component and more obvious fine-grained suspension tail are the major differences.
MEAN GRAIN SIZE
3.25 φ

SAND:SILT:CLAY
75.5:10.5:14

SAMPLE NUMBERS
226
329
Figure 35. Sandstone with "perched" cobbles produces frequency and cumulative curves with two obvious tractional modes typical of laminated deposits.
MEAN GRAIN SIZE
2.36 $\phi$

SAND:SILT:CLAY
84.5:5.5:10

SAMPLE NUMBER
RR-H-2
When grain-size distribution is plotted on a sand:silt:clay diagram, four main groups become evident (Figure 36). The fluvial sandstones (square symbols) have the most clay and silt, occurring over a very wide range but all within the clayey sandstone field. All but sample 0-8 of the isolated sandstone and interbedded sandstone/mudstone facies grouping (diamond symbols) also occur within the clayey sandstone area. All remaining samples fall within the sandstone corner of the diagram. Those that cluster the closest to the pure sand-sized corner (triangular symbols) are the amalgamated pebbly sandstones, cross-bedded sandstones, and the massive to laminated sandstones occurring within the very thick megasequence present within the 1 km-wide channel at Canyon #4. All remaining samples (closed circles) have a decreased sand-sized component and scatter between the extremely sand-rich cluster and the boundary with the clayey sandstone field.
Figure 36. Sand:silt:clay diagram for the sandstone samples analyzed texturally. Fluvial samples (square symbols) all fall within the clayey sandstone field, and contain the most clay and silt. Isolated sandstone and interbedded/lenticular-bedded sandstone and mudstone facies (diamond symbols) also fall within or very near the clayey sandstone subdivision. All remaining samples fall within the sandstone corner, with the amalgamated pebbly sandstone, cross-bedded sandstone, and massive to laminated sandstone facies (all triangular symbols) falling closest to the pure sand-sized corner.
PETROGRAPHIC ANALYSIS

Thin-section Study

Although most sandstones in the study area are extremely friable, some of the rare carbonate-cemented layers were sampled for thin-section study. Because of this, the cement types described are somewhat biased. Nineteen sandstone thin sections were examined to determine their mineralogies, textures, cements, and diagenesis.

Mineralogy

Table 3 and Figure 37 present mineralogic data for the examined thin sections. Groupings of the various detrital grains follow the classification scheme outlined by McBride (1963). Microgranular grains of quartz and feldspar are included with the unstable lithic fragments as opposed to being grouped with detrital feldspars, as proposed by Folk (1968), or by counting the individual constituent with the crystalline aggregate that falls under the counted point, as proposed by Dickinson (1970). This method better delineates type and proximity of the source terrane of the samples. Thus, most samples fall into the lithic arkose sandstone classifications of both McBride (1963) and Folk (1968). However, if the granitic (quartz and feldspar) fragments are included with feldspar grains, the samples would shift slightly towards the feldspar corner on the triangular diagram, some falling into the arkose classification.

The dominant stable grains are monocrystalline quartz of common plutonic type (Folk, 1968), with lines of vacuoles and straight to slightly-undulose extinction. Microlites (mineral inclusions) are not
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<th>Sample Number</th>
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<th>Detrital</th>
<th>Microcline</th>
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*Sample RC-A-1 also contained shell fragments at 22 count points
Figure 37. Triangular compositional diagrams displaying normalized thin-section mineralogy.
A) Q-F-L (quartz-feldspar-lithic fragments) plot. Two separate clusters of samples are portrayed. The amalgamated pebbly and cross-bedded sandstones are distinctly more quartz-rich (diamond symbols). These same samples also have some of the highest sand/(silt+clay) ratios (Figure 37). High-energy hydrodynamic processes probably sorted and "cleaned-up" (mechanically) the unstable and fine-grained components during deposition.
B) Qm-P-K (monocrystalline quartz-plagioclase-potassium feldspar) plot for the same samples as above. Once again, the cross-bedded and amalgamated pebbly sandstones (diamond symbols) appear to cluster together toward the quartz-rich corner. The possible potassium feldspar-rich cluster (triangular symbols) includes thin- and non-channelized, laminated sandstone units isolated within mudstone.
C) Qp-Lv-Ls (polycrystalline quartz-volcanic lithic fragments-sedimentary lithic fragments) plot with no evident clustering of samples.
common but are present, and include epidote, rutile and muscovite.
Strongly-undulose extinction occurs in both quartz and feldspar grains.
Polycrystalline quartz ranges from numerous quartz crystals with aligned
c-axes to subequant crystals with irregular, crenulate or more rarely
relatively straight boundaries. Altogether these polycrystalline grains
indicate a high-grade metamorphic origin (Folk, 1968).

Feldspar percentage is generally, though not always, less than
quartz. Orthoclase and microcline are the dominant potassium feldspars.
Plagioclase grains display weak zoning. Many feldspar grains are
undergoing diagenetic alteration, particularly by calcite, sericite,
or clay mineral replacement. Still others are being leached.

Rock fragments include three main types. Microgranular grains of
plutonic origin are composed of quartz, potassium feldspar, and/or
plagioclase, with subordinate mica. This mixed feldspar-plus-quartz
suite indicates a granitic composition for the source rock. In con-
trast, metamorphic grains include schistose forms, microgranular to
macrogranular quartz identical in appearance to the matrix of metarhyo-
lite cobbles found in Eocene strata, and mixtures of fine-grained equant
quartz and feldspar also similar to metamorphic cobbles present in
adjacent rock units. Therefore, numerous high-rank metamorphic rock
fragments, with lesser low-rank forms, are present. The third type of
lithic grain is sedimentary in origin. Shale and rarer siltstone clasts
are often partially compacted between more competent grains. In extreme
cases, pseudomatrix is formed by the complete mashing of the argilla-
ceous grains and fills primary pores.
Opaque minerals, predominantly magnetite and hematite with some pyrite, are quite abundant in many samples. In RC-A-1 these are authigenic in origin, localized within gastropod chambers. In many other samples their detrital origin is evident as they occur along microlaminations. However, the origin of these opaques is debatable in RC-P-2; although concentrated along laminations, they appear to have replaced some other unidentified, well-rounded grains.

Minor constituents include felsitic volcanic fragments, chert, micas, and other heavy minerals.

When the mineralogic data are plotted on triangular compositional diagrams modeled after Dickinson (1970), a variety of patterns emerge. On the Q-F-L plot (Figure 37a), two relatively distinct groupings emerge. While both clusters have approximately equivalent proportions of feldspar-to-lithic grains, quartz provides a greater percentage of the total rock for samples RC-P-2, TR-G-1, TR-A-1, 48, 0-9, and 0-11. Samples 48, 0-9, and 0-11 all are members of the amalgamated pebbly sandstone facies, and TR-G-1 and TR-A-1 belong to the cross-beded sandstone facies. These samples probably were hydrodynamically "cleaned-up" during deposition, with mechanical breakdown and removal of unstable grains. RC-P-2 is from the Mission Valley Formation, included for general comparison to the older units under study.

The Qm-P-K plot does not display such well-defined clusters, although three possible assemblages are delineated (Figure 37b). RC-P-2, TR-G-1, TR-A-1, 48, 0-9, and 0-11 all may once again bunch together, although such a grouping might also include Mt. Soledad
Formation samples RR-D-1 and 36 and cross-bedded Friars Formation sample RC-J-2. The most potassium feldspar-rich collection includes samples 0-4, 0-7, and 0-11, plus also possibly 34 and RC-A-1. Of special note is that these samples are all from laminated, thin-bedded, thinly- to non-channelized sandstones isolated within the mudstone facies. The remaining samples taken from this sea-cliff transect are intermediate between the more potassium feldspar-rich versus more quartzose assemblies.

The $Q_p-L_v-L_s$ plot (Figure 37c) shows a wide scatter of samples with no evident clustering. A $Q_m-F-L_t$ plot is not included because it essentially mimics the $Q-F-L$ plot. The significance of the mineralogy data is analyzed in the discussion section.

**Fabric Analysis**

The majority of samples display normal packing (Table 4); i.e. grains are predominantly in point (tangential) contact with one another. Compaction effects are minimal, with few overly close-packed samples, and little evidence of embayed or sutured grain boundaries or point contact fusion. Long and concavo-convex contacts are subordinate to point contacts. Some samples (Table 4) also have floating grains, i.e. an obvious separation of grain boundaries by pore space. However, incipient compaction did occur in these rocks. Fractures formed in some quartz and feldspar grains, and some sedimentary rock fragments occur as pseudomatrix deformed by more competent grains. All have micas, predominantly biotite, which are either bent around grains or into which other grains project, forming concavo-convex boundaries.
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<th>Cement Type</th>
<th>Packing</th>
<th>Grain Contacts&lt;sup&gt;1&lt;/sup&gt;</th>
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<td>iron oxide, authigenic clay and clay matrix</td>
<td>floating to overly close</td>
<td>floating, point, long, and concavo-convex</td>
<td>very poor</td>
<td>subangular to curvilinear</td>
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<td>iron oxide, clay matrix, and some blocky calcite</td>
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<td>point, long, and concavo-convex</td>
<td>moderate</td>
<td>angular to subround</td>
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<td>moderate to well</td>
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<td>blocky calcite over authigenic clay coatings</td>
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<td>poikilotopic calcite over authigenic clay coatings</td>
<td>normal</td>
<td>floating and point, some long</td>
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<td>Grain Contacts ¹</td>
<td>Sorting ²</td>
<td>Roundness ³</td>
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<td>point, long, and concavo-convex</td>
<td>moderate</td>
<td>subangular to subround</td>
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</table>

¹From Pettijohn, Potter, and Siever, 1972
²From Pettijohn, Potter, and Siever, 1972, and Folk, 1968
³From Krumbein, 1941
This deformation is never intense enough to form pore-filling pseudo-matrix.

Most samples of amalgamated and channeled facies (numbers 34, 48, 0-9, 0-11, 0-13, 0-16, 0-17) are poorly to very poorly sorted and have admixtures of coarse-grained and medium- to fine-grained material. The remaining samples are well to moderately sorted. The majority of the examined grains are subangular to subround. However, samples 36, 0-3, 0-9, 0-13, 0-17, Encb, RC-A-1, RC-J-2, RR-D-1, and TR-A-2 have abundant angular material. Many of these also fall into the group having poor to very poor sorting, which suggests a genetic relationship discussed in a later section. But it can be noted here that the channel sandstones along the beach cliff transects tend to have more angular and poorly-sorted material, and are therefore considered texturally immature (Folk, 1968).

**Cementation**

Four types of cementation are defined from the specimens: 1) phyllosilicate cement, 2) pore-filling blocky and/or poikilotopic calcite, 3) pseudomatrix, and 4) iron oxide cement. Many samples have varying degrees of a thin, irregular crust to isopachous druse coating the grains (Figure 38a). This is formed by authigenic clay, perhaps of the smectite-illite group based on its brown color, morphology, and birefringence. This type of cement is irregular to completely lacking in portions of some thin sections, whereas it forms incipient films completely around grains in others. Only in sample RC-P-2 does well-crystallized, apparently authigenic phyllosilicate (equivalent in appear-
Figure 38. Photomicrographs of the various styles of cementation.

A) Authigenic clay coating clastic grains. This dark brown phyllosilicate cement varies from a thick, regular, isopachous druse to thin, irregular to incipient films along grain boundaries. Plane-polarized light.

B) Blocky calcite (C) is the predominant pore-filling cement. Plane-polarized light.

C) Poikilotopic or "luster mottled" calcite cementation. The brightly-colored pore-filling cement is a single calcite crystal in optical continuity. Crossed nichols.

D) Pseudomatrix formed by compaction and deformation of detrital argillaceous grains. Distribution of this type of pore-filling material is patchy and localized. Plane-polarized light.
ance to that coating the grains) fill in primary porosity. It is possible that this is actually orthomatrix formed by recrystallization of detrital lutum or pseudomatrix. However, its homogeneity and lack of a relict clastic texture argues against this.

The major type of cement that is found to fill primary pores and overlie authigenic clay coatings is blocky and/or poikilotopic calcite (Figures 38b and 38c). Twin lamellae are especially common in the poikilotopic form. The presence of one type of calcite or the other in clastics has been related to pore sizes (thus affected by detrital grain sizes and sorting), the occurrence of original shell debris in the sample, and/or the distribution of nuclei for calcite crystallization (Scholle, 1979). However, no clear-cut relationships are observed in these specimens. RC-A-1 does contain abundant shell fragments and displays poikilotopic calcite, while many other samples have blocky calcite that is somewhat cloudy, perhaps arguing for interstitial silts acting as closely-spaced nuclei. But similar occurrences are quite variable. Pseudomatrix formed as a result of deformation of detrital argillaceous grains (Figure 38d). Evidence for this is relict grain boundaries and large matrix-filled areas within the sample. There is a complete spectrum from partially-squashed sedimentary rock fragments to severe pseudoplastic flow. This type of feature is very localized and is not significant as a cementation agent.

Another minor cement is iron oxide. This material coats fractures which cut calcite-cemented samples and also forms opaque, irregular patches which line grain boundaries and obliterate porosity in some
samples. It is never found as a cement when calcite is present as the primary porosity fill.

Another interstitial constituent is "protomatrix", apparently unre-crystallized detrital clay and silt which occurs as irregular, locally abundant patches that partially to completely fill primary porosity. This lacks the high birefringence of the authigenic phyllosilicates, and often forms cloudy areas which are surrounded by and interfinger with calcite cement. A second line of evidence for its detrital nature is that its presence does not create large "gaps" (larger-than-normal pores) between competent grains, which occur during the formation of pseudomatrix from sedimentary rock fragments, nor are relict argillaceous grain boundaries obvious.

**Diagenesis**

A later section covers possible paragenetic sequences for these samples. However, the diagenetic features are briefly described at this point. Mention has already been made of grain replacement, especially feldspars, by sericite and clay. Calcite has replaced both quartz and feldspar, as evidenced by extremely ragged grain boundaries (Figure 39a) and the "ghosts" of original grains indicated by authigenic clay coats now "floating" within the calcite cement. Calcite cement also filled fractures within quartz and feldspar, and cleaved micas apart by displacive precipitation, forming open, ragged grains (Figure 39b).

Shell fragments in sample RC-A-1 have been altered in a number of ways. Microboring organisms penetrated the shells, and micrite rinds
Figure 39. Photomicrographs of diagenetic features.

A) Calcite replacement. The black feldspar grain in the center has ragged boundaries, fractures, and internal portions being altered to light-colored calcite in optical continuity with the blocky calcite cement. Crossed nichols.

B) Displacive precipitation. The central biotite grain was cleaved apart by crystal growth of the blocky calcite cement (C). Plane-polarized light.

C) Secondary porosity development. The cleaved mica (upper left quadrant) and fractures within the quartz grains (q) contain no cement filling these now-open networks. This attests to dissolution of the original pore-filling calcite (?) cement. Plane-polarized light.

were then produced. In some cases the original shell material was leached away and calcite in optical continuity with that surrounding the micrite envelopes has been reprecipitated within them. Another calcite alteration of shell fragments is that of neomorphism, in which large blocks of calcite are in the process of altering the original shell mineralogy. A final alteration is silicification, with the original carbonate shell being replaced by silica.

Of major importance in the diagenesis of these samples has been secondary porosity development (Schmidt, McDonald, and Pratt, 1977). Dissolution of original carbonate cement is the most striking type. Oversize pores and "floating" grains, partial dissolution, inhomogeneity of packing, elongate pores, and grains with extremely irregular margins (presumably having undergone some calcite replacement prior to decementation) all attest to this process. Further lines of evidence are open fractures in quartz and feldspar which should have been closed upon burial and compaction, and micas which are split apart with their cleaved layers floating in open pores (Figure 39c). Some cement, probably calcite, filled these fractures and caused displacive precipitation but has since been removed.

Dissolution of feldspar is also evident (Figure 40). This may have occurred as clay or calcite replacement, then by dissolution of these materials or by true leaching of the feldspars themselves. Evidence for this type of porosity development includes partially-dissolved grains of feldspar and relicts of authigenic clay coatings which are present in oversized pores as external molds of former grains. Some
Figure 40. Photomicrographs of successive stages in feldspar dissolution.

A) Incipient dissolution along fractures (arrows). Note the clayey rind. Plane-polarized light.

B) Open lattice-like network of a remnant feldspar. Large pores (arrows) are also developing. Plane-polarized light.

C) Small remnants (arrows) of an original feldspar grain. The authigenic clay rind is preserved, in this case "floating" within an open pore formed by dissolution of the original calcite(?) cement. Plane-polarized light.

D) Complete dissolution of a feldspar. Alteration to phyllosilicates, especially along fracture boundaries, commonly forms a lattice-like or banded network within the original or authigenic clay rind. Plane-polarized light.
feldspars developed clay-filled cleavage fractures along cleavages, probably due to preferential replacement along these zones; these clays remain in partially- to completely-dissolved grains as a box-like network mimicking original feldspar cleavages.

One minor type of secondary porosity development is microfracture porosity (Figure 39d). This is relatively rare and volumetrically insignificant even in those samples where present.

Heavy-mineral Study

In order to investigate provenance variations with time and aid in stratigraphic correlation, twenty-one different heavy minerals were identified and counted for twenty-six samples (Table 5). The six most abundant grain types (epidote, topaz, kyanite, actinolite, black opaques, and chlorite) were then normalized to 100% for each sample (Table 5). A variety of interesting patterns are evident. When these normalized minerals are plotted at their respective positions along the Torrey Pines-Scripps beach-cliff transect, kyanite is seen to increase while actinolite tends to decrease upward in most sections (Figure 41). Epidote and topaz remain about the same in all samples, except for the low amounts of epidote in numbers 82 and 86. Reduced epidote percentages as well as very low actinolite values are also found for samples 33, 70, 72, RC-I, RC-J-2, and RC-P-2. Actinolite is especially high in RR-I-2 and RC-D-1. The significance of these trends, and the significance of the heavy mineral types themselves, are explored in a later discussion section.

Weight percent of heavy minerals versus total sample are also
### TABLE 5: Heavy Mineral Analysis for Selected Frangible Samples

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<th>Colorimeter</th>
<th>Topaz</th>
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**Weight % Therapy/Total Sample**
Figure 41. Relationship of heavy-mineral abundances to stratigraphic interval along the Torrey Pines-Scripps transect. Total weight percent of the heavy-mineral fraction is shown at the appropriate stratigraphic position for each sample, followed by the normalized relative percents of epidote:topaz: kyanite:actinolite:black opaques:chlorite determined by point counting. Sampling intervals are indicated by: ←. The stratigraphic sections are reconstructed in their relative pre-fault positions. In most sections, kyanite increases upwards while actinolite decreases. Thick channel sandstones and sandstone matrices of cobble conglomerates contain the highest total weight percent of heavy minerals.
given (Table 5). These are likewise plotted along the Torrey Pines-Scripps Transect. Higher values (4.8 to 2.5%) are present in the major, thick channel sands and conglomerate matrices (Figure 41). Minor channels and thin sandstone beds have much lower values (1.0 to 0.4%). The significance of this and other weight percentage trends is also examined later.
SEDIMENTARY FABRIC ANALYSIS

Conglomeratic Fabrics

Measured attitudes of clast-supported conglomerates show fabrics that are predominantly imbricated along their a-axis. The b-axes are therefore perpendicular to the presumed paleoflow directions, and the a-axes dip toward the upcurrent direction. The prevailing currents responsible for the deposition of these rocks ranged from northwesterly-to southwesterly-flowing, with only four localities yielding results indicating flow components towards the east (Appendix II and see Figure 61). This pattern was expected, as previous work (e.g. Kennedy, 1975; Howell and Link, 1979) indicated an Eocene fluvial system to the east of the study area that transported detrital materials to the western coastal zone.

Mt. Soledad Formation conglomerates (b,c,d,e,f, of Appendix II) show a predominant northwestward paleoflow pattern. In evaluating the paleogeographic significance of this trend, a palinspastic reconstruction of lateral movement along the Rose Canyon fault zone was attempted. As discussed later, two different loci of input and deposition are indicated. One main area of deposition would have been located south of the present-day study area, while the other was centered within the area under investigation.

Along the Torrey Pines-Scripps beach-cliff transect, a divergent pattern of cobble imbrications is evident. A southeastward flow is indicated in the amalgamated pebbly sandstone at Canyon #1 (g, Appendix II). In contrast, the Scripps Formation conglomerates yield abundant
northwestward and southwestward current directions from Canyon #4 to Scripps Pier. The northern group (h, i, j, k, l, m, and n of Appendix II) is dominantly directed towards the northwest, and the southern group (o, p, q, r, and w) is southwesterly oriented. This yields a conspicuous radiating pattern.

Of special note are two measured intervals in Canyon #5 (m and n). These are within an upper disorganized-to-graded amalgamated unit. The basal disorganized portion (l, Appendix II) shows the typical pattern of a-axes dipping upcurrent and b-axes aligned transverse to flow. A thin overlying portion of the unit is graded and strongly bimodal (m), with the a-axes transverse to flow and b-axes imbricated. The uppermost matrix-supported layer of this unit then has a-axes dipping south-westward, yielding an apparent northeast current direction. Many measured intervals within the field area contain minor components of bimodality or a-axis imbrication down current, but these are usually subordinate to a-axis imbrication upstream. These two intervals in Canyon #5 are the only instances where typical a-axis imbrication patterns give way to other varieties. The significance of these two varieties in relationship to hydrodynamics is discussed later.

Inland, the Scripps Formation facies at Los Peñasquitos Canyon and Genesee Avenue North closely resemble those present at Canyon #4. Large channels with a single positive megasequence are cut into Ardath Shale and are filled with amalgamated coarse conglomerate grading to thick massive to laminated sandstone, then a variety of sandstone and siltstone intervals. At Los Peñasquitos, the measured current orienta-
tions (s and t) are directed toward the southeast (the only occurrence of this trend other than the pebbly sandstone scour of Canyon #1). At Genesee Avenue North (u and v of Appendix II), the paleoflow was more variable, with both southwestward and northwestward orientations.

Amalgamated and channelized Stadium Conglomerate exposed along the Atchison, Topeka, and Santa Fe railroad cuts (Soledad Canyon, Miramar Road, and Rose Canyon sites) is dominantly southwesterly directed (z, aa, bb, cc, dd, and ee of Appendix II), with only one measured westerly interval (ff). The average mean vector magnitude for this group is higher (0.60) than for any other areally closely-spaced group (for the Mt. Soledad Formation, this parameter is 0.52; for the Scripps Formation along the beach cliffs it is 0.53). Thus the Eocene currents were relatively more effective in producing both a strong and a consistent fabric in this area for these Stadium Conglomerate units.

Sandstone Fabrics

Grain orientations were determined from thin sections cut parallel to bedding. A sense of grain imbrications and hence unique paleocurrent directions was not determined. However, because elongate grains dominantly align themselves parallel to a flowing current (Parkash and Middleton, 1970; Potter and Pettijohn, 1977, p. 47-55; Hiscott and Middleton, 1980), a bidirectional line of movement may be attained.

Fifteen thin sections all display well-developed grain orientations. For each sample, one directional trend is usually predominant over a second trend formed at approximately right angles (a-o, Appendix II). Surprisingly little scatter of the elongate grains within each
sample is evident. Twelve of the samples contain a dominant trend which follows a southwest-northeast strike. The remaining three (h,j, and o) have a northwest-southeast orientation. If these grains are imbricated with a sense of (westerly) flow analogous to that of the conglomerate clasts, then the paleocurrents indicated by these elongate grains were predominantly towards the southwest (see Figure 61).

While strong grain orientations might be expected in channelized sandstones (such as f,e,i,k,l,n, and o, Appendix II), just as strong orientations are found in flat-based laminated to graded units (d,f,g, h, and j). The marked parallel arrangement of the majority of trends is striking; units deposited by overbanking perpendicular to the main channel trends and by random settling out from suspension did not dominate sample collection. Especially surprising is that the sampled bioturbated sandstone and silty sandstone (a,c, and m) were not disrupted to such an extent that good grain orientations were destroyed.

Other Paleocurrent Indicators

Other paleocurrent indicators are consistent with the presence of an Eocene westerly-directed drainage system. Parting lineations at Miramar Road average just about due east-west, consistent with ripple crests indicating average flow toward 280° (N80W). Channel axes at Soledad Canyon, one of which is filled by a slump, trend both southwest-northeast and northwest-southeast (strikes of 205° and 324°). Similarly, channels to the west of the Rose Canyon locality, west of the Soledad Canyon site, along Gilman Drive, and at Torrey Pines Road also trend in a variety of directions (respectively 345°, 225°, 297°, and
335°). Slumping down a channel wall at Price Club indicates an axial trend of 221°. Ripples with a sense of flow toward 343° at this locality probably are related to an overbank type of deposition.

Channels along the Torrey Pines-Scripps beach transect also display a variety of orientations. The large sandstone-plug channel exposed above Torrey Pines Landslide strikes 255°. Above the interbedded sandstone and siltstone facies at Canyon #3, successively higher channels trend 345°, 305°, 200°, and 235°. Thus these are meandering and cross-cutting. Within Canyon #4, channel axes which also display cut-and-fill just above the massive to laminated sandstone strike at 285°, 255°, and 210°. The small sandstone channel isolated within Ardath Shale near the base of Canyon #5 has flutes which show an average flow toward 246°. Ripples near the top of this channel display a similar current direction, indicating transport toward 230°. Near the top of Canyon #5, channel trends swing back to a northwest-southeast orientation; the channel 65 meters above the section base strikes 295° and that at 96 meters strikes 305°. Finally, at 45 meters in the Canyon #6 section, an isolated sandstone channel has a similar direction, trending 315°.

At Tourmaline Surfing Park, slumps with average axial fold trends of 260°, 265°, 281°, and 284° slid toward the south away from the conglomerate-filled channel. Ripples indicate flow relatively parallel to that of the conglomerate-filled channel, toward 337°.

Trends somewhat different than the regional paleocurrent distribution (i.e. channels and non-channelized deposits trending both northwest and southwest with minor components related to slumping and over-
bank currents directed away from, or in the case of other slumps, toward channel axes) are rippled turbidite sequences. Thick Bouma (1962) $T_C$ intervals north of Canyon #5 at beach level display average flow towards $338^\circ$ ($r = 0.96$, $s = 15^\circ$, $n = 23$). Almost directly opposite trends are present just north of Scripps at beach level, with currents directed toward $155^\circ$ ($r = 0.97$, $s = 14^\circ$, $n = 31$). Some possible reasons for this variance are discussed later.
PALEONTOLOGY

Previous Work

As already stated, Kennedy (1973a) carried out a detailed biostratigraphic correlation of the complete Eocene sequence of San Diego County, and summarized the results in Kennedy (1975) and Kennedy and Peterson (1975). Golz (1971) and Golz and Kennedy (1971) studied mammalian fossils from a variety of localities. Taxonomy of the rich molluscan faunal assemblages of San Diego was presented in classical work by Hanna (1926), plus additional work by Givens and Kennedy (1976) and Givens (1978). Calcareous nanoplankton collected by Bukry and Kennedy (1969) from the Ardath Shale provide the bulk of the previous floral work, although palynomorphs have also been examined (Elsik and Boyer, 1977).

In comparison, foraminifers have been extensively studied by various researchers. In early work, Cushman and Hanna (1927) collected specimens just south of Canyon #3. Gibson (1971) reported solely on benthonic forms from the Ardath Shale and Stadium Conglomerate and discussed their biostratigraphy and paleoecology. Henry (1972) examined both benthonic and planktonic foraminifers from an Ardath Shale locality as a senior thesis. Steineck et al. (1972) examined the planktonic faunas of the Ardath Shale and Stadium Conglomerate, which helped lead to the realization that the benthonic foraminiferal stages of Mallory (1959) are time-transgressive due to their ecologically-controlled distribution. Finally, Lohmar (1978) made a collection of foraminifers from the Torrey Pines-Scripps beach-cliff transect,
summarizing the results in Lohmar and Warne (1978). These works indicate that the Middle and Late Eocene boundary (given as 40 million years B.P. by Vail and Hardenbohl, 1979) lies near the transition from the La Jolla to the Poway Groups. This boundary falls within the Uinta B North American land mammal age of Wood et al. (1941), based on work by Golz and Kennedy (1971). Golz (1971) found that the complete Eocene sequence of San Diego ranged at least from the Bridgerian to Uinta C mammal ages (Figure 42).

Work on molluscan biostratigraphy (Givens & Kennedy, 1976) correlates the younger Uinta C mammals with the Tejon Californian molluscan stage, the Uinta B age with a "Transition" stage, and places the majority of the La Jolla Group within the Middle Eocene Domengine molluscan stage (Figure 42). Assemblages of Turritella uvansana applinae Hanna, Ficopsis cooperiana Stewart, and Tejonia lajollaensis (Stewart) from the middle to upper portion of the Mount Soledad Formation are restricted to the Domengine (Kennedy, 1973a). A similar assemblage was found in both the Ardath Shale and Scripps Formation. A Friars Formation collection including Nekewis io (Gabb) and Ectinochilus canalifer supraplicatus (Gabb) indicates at least partial incorporation of this unit into the Middle to Late Eocene "Transition" stage (Kennedy, 1973a).

Calcareous nannoplankton collected from the Ardath Shale at a variety of localities contain Discoaster sublodoensis Bramlette and Sullivan, Discolithina exislis (Bramlette and Sullivan), D. frimbiata (Bramlette and Sullivan), Lophodolithus mochlophorous Deflandre, and Rhabdosphaera inflata Bramlette and Sullivan (Bukry and Kennedy, 1969).
Figure 42. Chart showing possible correlations of various biostratigraphic zonations used in determining stratigraphic and age relations for the Eocene of San Diego. The absolute time scale correlated to the Eocene epoch, European stages, and planktonic foraminiferal zones is from Vail and Hardenbol, 1979. Californian molluscan stages, West Coast marine stages, North American mammal ages, and stratigraphic sequence is modified from Kennedy, 1975, and the Geologic Time Table of Elsevier Scientific Publishing Company, compiled by F. W. B. Van Eysinga, 1975. Benthonic foraminiferal zones are modified from Lohmar and Warme, 1978, and the nannoplankton zonation is from Okada and Bukry, 1980.
These species correlate with the upper part of the early Middle Eocene Discoaster Sublodoenis Concurrent-range Zone of Hay and others (1967). Bukry and Kennedy (1969) also showed that this group is correlative with Lutetian (European stage) strata of the Paris Basin and the Hantkenina aragonensis Range Zone of foraminiferans from Trinidad (Figure 42).

Mallory (1959) assigned the Upper Eocene sequence to his (Late Eocene) West Coast marine Narizian stage, and the lower portion to his Ulatisian (Middle Eocene) stage. Ardath Shale fauna collected at the type locality on the west side of Rose Canyon by Gibson (1971) support this interpretation, and place the Ardath at this locality into the Amphimorphina californica Zone (late Middle Eocene or late Ulatisian stage) for benthonic foraminiferans (Figure 42). A diverse arenaceous fauna with large bathyal Cyclammina, shelfal Eponides, and abyssal Anomalina, Gyroidina, Buliminella, and Cibicides led Gibson (1971) to consider this group as mid-bathyal (600-1500 meters). The presence of basal Middle Eocene to medial Middle Eocene planktonic Pseudohastigerina micra (Cole), Subbotina patagonica (Todd and Knicker), and Truncorotaloides spinuloinflatus (Bandy) supported Gibson's (1971) benthonic age interpretation. Henry (1972) made analogous findings. Steineck et al.'s (1972) work substantiated the planktonic data of Gibson (1971) and included the Mount Soledad Formation outcrop at Tourmaline Surfing Park. The Ardath and Mount Soledad localities were both bracketed within the early Middle to late Middle Eocene Hantkenina aragonensis and Globigerapsis kugleri Zones of Bolli, 1957 (later revised in Bolli,
Benthonic foraminiferans from the Ardath Shale along the Torrey Pines-Scripps beach-cliff transect indicate the presence of three of Mallory's (1959) zonations: 1) **Alabamina wilcoxensis** Zone (late Early Eocene or late Penutian Stage), 2) **Vaginulinopsis mexicana** Zone (early Middle Eocene or early Ulatisian Stage), and 3) **Amphimorphina californica** Zone described above (Lohmar, 1978). The **A. wilcoxensis** Zone was interpreted from mudstones of Canyon #1, the **V. mexicana** Zone was present beneath the Hang Glider Port and immediately inland of the sea cliffs, and the **A. californica** Zone was recognized near the base of Canyon #5 and along the east wall of Rose Canyon. However, this apparently "younging-to-the-south" trend along the beach transect was interpreted by Lohmar (1978) as being predominantly facies, not age, related. The **A. wilcoxensis** Zone contains undifferentiated bathyal forms (200-2000 meters); transitional bathyal to inner abyssal forms (>2000 meters) are found in the **V. mexicana** group, and the **A. californica** Zone assemblages represents an inner abyssal fauna (>2000 meters).

**Timing of Deposition**

The works cited above indicate that the Mount Soledad Formation to Friars Formation sequence ranges in age from late Early Eocene (late Penutian) to early Late Eocene (early Narizian). In order to get a better idea of timing of the depositional sequence along the Torrey Pines-Scripps Transect, nannofossil samples were processed by C. H. Ellis and W. H. Lohman of Marathon Oil Company, Denver Research Center.
A total of 51 samples were examined; 21 were barren and 14 contained an indeterminate or undifferentiated middle Eocene assemblage. The list of the identified species, their ranges and locations, and interpreted subzones are presented in Table 6 (Ellis, written communication, 1978).

Samples 3244-6, 3244-7, and 3244-9 (Canyon #1) contain *Rhabdosphaera inflata* which places them in the middle Eocene *R. inflata* Subzone of the *Discoaster sublodoensis* Zone. Samples 3402-1 (between Canyons #3 and #4) and 3729-6 (Canyon #4) contain *Nannotetrida quadrata* which occurs just above the *R. inflata* Subzone. However, the presence of *Discoaster sublodoensis* and *D. Iodoensis* in these samples is a puzzle. These two forms supposedly do not range above the *R. inflata* Subzone, and thus these samples are best assigned an age at the boundary of the older *R. inflata* Subzone and the younger *D. strictus* Subzone (*Nannotetrida quadrata* Zone). *Lophodolithus mochlophorus*, indicative of the *R. inflata* Subzone, is found in Sample 3719-5 (between Canyons #4 and #5). Sample 3720-2 (Canyon #6) contains *Chiasmolithus gigas*, placing it in the *C. gigas* Subzone above the *D. strictus* Subzone (*N. quadrata* Zone), the youngest interval identified.

Thus all of these samples support a Middle Eocene depositional interpretation. A weak pattern of "younging-to-the-south" is present, but does not include nearly as much of a time span indicated by the facies-controlled benthonic foraminifers from these same cliffs. The possible significance for this trend is examined later.

**Environments of Deposition**

When investigating previous work, it becomes apparent that taxon-
Table 6: Nannofossil Occurrences* and Age Interpretations along the Beach-cliff Transect

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* *r = rare  
 f = few

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<td>N. quadrata</td>
<td>C. staurion</td>
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<td>C. gigas</td>
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<td>D. sublodensis</td>
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<td>D. lodoensis</td>
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188
omic and biostratigraphic correlations have been the predominant paleontologic work, with only cursory attempts at paleoecologic interpretations by using benthonic foraminiferans. The object of this study was not to carry out such an investigation because that in itself is worthy of a whole thesis. However, certain easily-identifiable mollusks were noted wherever found, and help provide some clues in paleogeographic reconstruction.

The large landslides along the Torrey Pines-Scripps transect, especially between the Torrey Pines Landslide south to Canyon #4, have carried fossiliferous debris down from the upper sandstone and siltstone beds. These same fossils are found in place (see, for example, the measured stratigraphic section of Canyon #3). Similar fossiliferous layers are present in the Monongahela Drive, Balboa Avenue, and Pacific Theater localities. Some of the macrofossils found include: 1) bivalves - *Atrina sp.*, *Solen ?parallelus*, *?Acrilla tejonensis*, *?Pitar joaquinensis*, *Acila decisa*, *Benercardia hornii*, *Macrocallista ?hornii*, *Optrea sp.*, and *Nuculana sp.*; 2) gastropods - *Turritella sp.*, *Amaurellina moragi lajollensis*, *?Trichotropis lajollaensis*, and *Crithium sp.*; 3) *Dentalium ?aspicostatum* scaphopods; 4) *Flabellum sp.* coral; and 5) abundant colonial vermetid worm tubes. For a synonomy see Hanna, 1927, Vokes, 1939, and Givens, 1974. Many of these forms display borings by endolithic microorganisms, including algae, fungi, and sponges. Natural casts of microborers within leached shells are common. While some shells are fragmented, a great number of whole, articulated specimens such as *Turritella* and *Acila* form lenses and
layers. In contrast, very small, thin-shelled and often fragmented mollusks and gastropods are found in the mudstone facies of the Ardath Shale north of Canyon #4. *Nucluna* is one common form. Within the Ardath Shale at Canyon #6, Givens (oral communication to J. Warme, 1979) identified *Loxotrema turritum* Gabb within a sandy interval. Just below this, he found specimens of *Portlandia* and *Macoma*. Within the channelized Scripps Formation south of the Hang Glider Port, all the mollusks described above can be found. However, they are obviously resedimented, having been intensely fragmented and transported with the associated coarse conglomerates. Evidence of microboring activity within these shells is also abundant.

A different assemblage of mixed broken and whole shells is found inland. This type is present in channelized sandstones within the lower Price Club and Pacific Theater sections and associated with basal clast-rich layers of the conglomerate-based sandstone facies at the Rose Canyon and San Clemente Canyon Road sites. The macrofossils are dominantly disarticulated, thick-shelled bivalves and gastropods. Some of these forms are *Glycymeris sp.*, *Venericardia horni*, *Ostrea indriaensis* and *Ostrea sp.*, *Crassatellites sp.*, and *?Ampulina hannibali*.

These macrofossils are used in a later section to help interpret depths and environments of deposition. Common depth ranges and habitats of extant relatives for some of these species are shown in Table 7.

Samples from the Torrey Pines-Scripps transect sent out by Marathon Oil Company, Denver Research Center for foraminiferal analysis
**TABLE 7: Common Distribution of Extant Relatives of Eocene Macrofossils**

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<th>Genus</th>
<th>Distribution (and Habitat)</th>
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<tr>
<td>Acila</td>
<td>9 - 1500 m, usually less than 915 m (fine-grained or sand substrate)</td>
</tr>
<tr>
<td>Atrina</td>
<td>6 - 27 m (muddy or clean sand bottom)</td>
</tr>
<tr>
<td>Chione</td>
<td>6 - 232 m (clean to slightly muddy sand)</td>
</tr>
<tr>
<td>Crassatella</td>
<td>0 - 712 m, usually less than 73 m</td>
</tr>
<tr>
<td>Glycymeris</td>
<td>0 - 640 m, usually less than 110 m</td>
</tr>
<tr>
<td>Macrocallystra</td>
<td>0 - 27 m (clean or slightly muddy sand)</td>
</tr>
<tr>
<td>Mytilus</td>
<td>0 - 91 m, usually less than 3 m (attached to rocks)</td>
</tr>
<tr>
<td>Nuculana</td>
<td>5 - 3658 m (muddy substrate)</td>
</tr>
<tr>
<td>Ostrea</td>
<td>0 - 91 m, usually less than 18 m (brackish to open marine waters)</td>
</tr>
<tr>
<td>Pitar</td>
<td>0 - 540 m</td>
</tr>
<tr>
<td>Solena</td>
<td>0 - 46 m (sandy mud to fine sand)</td>
</tr>
<tr>
<td>Turitella</td>
<td>20 - 200 m</td>
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<tr>
<td>Venericardia</td>
<td>9 - 130 m, usually less than 45 m (coarse-grained sand)</td>
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</tbody>
</table>

*from Stanley, 1970; Hertlein and Grant, 1972*
were interpreted as predominantly mid-bathyal (greater than 500 meters, D. G. McCubbin, oral communication to J. Warme, 1978). Many of the assemblages were identified as mixed mid-bathyal and reworked shallow forms with evidence of size and taxonomic sorting. North of Canyon #3, most samples were barren, although one isomorph of the Recent mid-bathyal Histominella scitula was found below the cross-bedded sandstone facies. Other samples contain outer neritic forms. There is a tremendous variation in foraminiferal abundance presumably related to sedimentation rates, not to dissolution on outcrop. Some rare Early Eocene specimens were found, but the dominance of forms indicative of Middle Eocene deposition suggest the older forms were reworked.
INTERPRETATION AND DISCUSSION

FACIES ANALYSIS

Beach-cliff Transects

Summary

Lohmar (1978) and in other publications (Lohmar and Warme, 1978 and 1979) defined seven separate environments of deposition from facies exposed along the Torrey Pines-Scripps sea cliffs and those at Tourmaline Surfing Park. The purpose here is not to repeat this work; instead a further breakdown of Lohmar's interpretation is accomplished for some environments, and some of his interpretations are questioned. Otherwise, this work is in agreement with his general model and that of Howell and Link (1979). Terminology follows that of Walker for submarine canyon-fan environments (Figure 43).

Pebbly amalgamated sandstone, planar- to convolute-laminated sandstone, and channelized mudstone with sandstone and siltstone sequentially fill a submarine canyon tributary cut into lagoonal Delmar and nearshore Torrey units. A progradational sequence of transitional outer- to inner-shelf deposits of interlaminated sandstone, siltstone, and mudstone facies and fossiliferous interbedded sandstone and mudstone facies is present overlying the canyon tributary system. South of the Hang Glider Port (Plate 2, Appendix III), an inner-fan channel-fill sequence is faulted into place adjacent to shelf deposits. This unit overlies base-of-slope deposits and gives way to inner mid-fan distributary channels further south.
Figure 43. Submarine fan environmental model (Walker, 1979).
D-B CGLS. represents disorganized-bed conglomerates. The inner or upper fan contains a major channel and levee system; the coarsest sediment is restricted to the channelized portion, while overbank deposition mixed with down-slope slumping dominates the remaining inner fan area. Smaller braided distributaries shift position and debouch sediment onto the mid fan, producing a coalesced channel system deficient in fine-grained levee units. The outer or lower fan then represents formation by turbidity currents which have flowed beyond the confines of this channelized portion.
An original interpretation by Lohmar (1978) of inner-fan margin deposits just below the Hang Glider Port are reassigned to a shelf sequence. Similarly, his fan fringe deposits at Tourmaline Surfing Park more closely resemble a mid-fan channel and levee sequence. Finally, a wave-worked fan-delta front deposit of the Mount Soledad Formation at Canyon #3 is recognized. This unit is topped by a kaolinitic (?) weathering profile and overlain by Delmar Mudstone and Torrey Sandstone. Shallow-marine Deposits

Boyer and Warme (1975) defined five facies in the shallow marine Delmar Formation and Torrey Sandstone. These units represent an interfingering transgressive sequence predominantly exposed north of Canyon #1. Lagoonal and tidal flat Delmar sedimentation gives way upward to the Torrey tidal channel and subaqueous bar sequence (Plate 1, Appendix III).

In the uppermost portion of Canyon #1, burrowed silty sandstone and trough cross-bedded sandstone also represent shallow marine environments (see Figure 51). These are part of Lohmar's (1978) prograding shelf environment. The trough cross-bedded sandstone is arranged in multiple sets of very coarse- and coarse-grained sandstone. Grain-size frequency curves for cross-bedded Torrey exposed in the base of the Bathtub Rock Trail and that exposed in the uppermost part of Canyon #1 are very similar; these deposits have the largest average grain size of any unit sampled (Figure 26). Abundant wood and rare thick-shelled mollusk fragments are found. Relatively high heavy-mineral abundances are present. All these observations indicate a
high-energy, nearshore zone, where coarse-grained and hydrodynamically-heavy material is deposited by traction. Through cross-bedded sands of this type are typical of the shoreface-foreshore transition zone or upper shoreface (terminology variation represents barred versus non-barred settings) (Davidson-Arnott and Greenwood, 1976). In this environment, the coarsest materials are typically concentrated along the breaker zone (Harms et al., 1975). In the Torrey Sandstone, wave action has winnowed the material finer than 3\(^\circ\).

The associated silty bioturbated sandstone is medium- to fine-grained, structureless, and contains abundant simple vertical Ophiomorpha burrows. This is typical of the lower shoreface zone. Box cores in the Gulf of Gaeta, Italy (Reineck and Singh, 1971) contain cross-stratified sand in both large and small sets (upper shoreface) overlying low-angle cross-stratified burrowed sands which display an increase in bioturbation downward. In the Eocene San Diego example, the intervening low-angle cross-stratification (middle shoreface) is lacking, perhaps attesting to a steep-gradient shoreline and rapid depositional-environment transitions. Truncations of burrows are evidence for periodic, high-energy, erosional events. The abundant silt-sized material indicates deposition below fair-weather wave base with complete overturning and reworking of the sediments by burrowing organisms.

A minor occurrence of Mount Soledad Formation conglomerate occurs at the base of Canyon #3 (Plate 2, Appendix III). Kennedy and Moore (1971a) suggest these rocks are non-marine and grade into the Mount Soledad unit at Tourmaline Surfing Park. This unit is composed of
matrix-rich conglomerate facies, with crystalline-clast a-axes transverse to flow (Figure 10d). Conglomeratic units interfinger with sandstone beds. This exposure is deeply weathered, obviously having been subaerially exposed, allowing for a kaolinitic (?) soil production prior to transgression by shallow marine Delmar and Torrey beds. This sequence is also located relatively further seaward of the typical proximal fan-delta Mount Soledad Formation at Morena Boulevard and along Interstate 5 (see below). Based on these data, the exposure below Canyon #3 is preliminarily assigned to a fan-delta front environment where wave-worked deposition took place. Clifton (1973) characterizes this zone as having discrete, laterally-extensive beds of gravel alternating with sandstone. Wescott and Ethridge (1980) describe a similar sequence from Jamaica at the transitional zone between the subaerial fan delta and submarine fan delta.

**Shelf Deposits**

The rapid transition from shelf to overlying nearshore facies is indicative of rapid progradation and/or high depositional-gradient during the Eocene. The inner-shelf environment (see Figure 50) of deposition is characterized by fossiliferous interbedded sandstone and mudstone facies. An arbitrary break between inner and outer shelf units is placed at sand:mud ratios of about 1:5 (this ratio becomes as high as 1:1 in the inner shelf environment), but this transition in actuality is entirely gradational. The abundant fossil debris along sandstone bases is more diagnostic, indicative of periodic storm transport of this material offshore. These forms include typical
neritic organisms (Table 7), among them *Turitella*, *Acila*, *Atrina*, *Solena*, and colonial vermetid worms. Many shells are abundantly bored by microorganisms, including sponges, indicating post-mortem exposure then later resedimentation. Another line of evidence for at least shelf depth is the abundance of microboring fungi present in these shells, typical of exposure at the sediment-water interface below the photic zone (see, for example, Golubic et al., 1975; Zeff and Perkins, 1979).

Reineck and Singh (1975, pp. 320-321, 323-329) discuss shelfal muds with storm sand-layers from the Nordergrunde Busum Regions, North Sea, and off Galveston Island. Below about 15 meters water depth, clayey silt alternates with fine-grained sand and sandy silt. Autochthonous shell layers at somewhat erosional bases of sand units indicate periodic high energy conditions moving sediment seaward from the near-shore zone. Sand layers are typically laminated to somewhat cross-bedded. Intervening fine-grained material is typically laminated with increasing bioturbation upwards; starved ripples may be present. While these are much thinner than many of the beds observed throughout the San Diego study area, thicker analogous sandstone units have also been described.

These sublittoral sheet sands with storm lag layers are typically 5 to 30 cm thick and occur in both shelf and offshore settings. Planar laminations have been attributed to both upper flow regime and lower flow regime deposition on the Oregon shelf (compare Komar et al., 1972, and Kulm et al., 1975). Thin graded units reflect fallout from suspension, perhaps from small turbidity currents. The Eocene beds under
study herein show a strong tractional mode plus a fine-grained tail
typical of deposition from suspension (Figure 33). A wide range of
bed thicknesses, types of internal structures, and degrees of frag-
mentation of bioclasts indicates a variety of interacting tidal, wave,
and storm currents (see Johnson, 1978, pp. 245-255 for a complete dis-
cussion). Some analogous ancient examples are described by Brenner and
Davies, 1973, 1974; Bridges, 1975; Kelling and Mullin, 1975; and de Raf
et al., 1977. One example of a prograding Late Precambrian shelf (Banks,
1973) contains many features directly comparable to the San Diego
sequence.

These units grade into fossiliferous interbedded sandstone and
mudstone with decreasing sandstone to mudstone ratios (less than 1:5)
and an increased amount of interlaminated sandstone, siltstone, and
mudstone. Banks (1973), Reineck and Singh (1972), de Raf et al. (1977),
and Bourgeois (1980) all discuss similar units within ancient and
modern outer-shelf environments. Typically these layers are arranged
as "rhythmites" (laminated mudstone and sandstone) with intervening
very thin sand layers. "Lam-scram" sequences are present. Reineck
and Singh (1972) attribute these features to periodic low-density
suspension deposition related to storms, with organic reworking of
increasing density upward in intervening fine layers during quiet
water periods. Stow and Shanmugam (1980) have called upon fine-grained
turbidites as the generating mechanism, and provide convincing vertical
sequences analogous to the Bouma (1962) sequence in coarse-grained
turbidites.
Submarine Canyon Deposits

These units were discussed by Lohmar (1978) and will only be highlighted here. An erosional unconformity representing the floor of a submarine canyon-head tributary cuts across the older, horizontally-bedded, shallow-marine Delmar and Torrey units with an axial gradient of approximately 5° (Plate 1, Appendix III). The erosive surface is stepped and plucked, with features of injection, pry-ups, scour, erosional remnants projecting upward into the overlying fill, and undercut ledges (Figures 11 and 44a). The overlying canyon fill proper is tripartite, representing progressive detachment of the canyon tributary from a coarse-grained, nearshore source (Figure 44b).

The basal fill unit is an amalgamated pebbly sandstone, an abrupt facies change from typical cross-bedded Torrey Sandstone. Cross-cutting, very poorly-sorted, pebbly, granule to medium-grained sandstone is structureless to faintly laminated. Basal portions of many cut-and-fill units contain irregularly-margined mudstone ripup clasts that are matrix supported. These clasts are of Delmar Formation plus others containing shallow water foraminiferans not reported in exposed Eocene or Cretaceous strata. Each separate unit becomes laminated upward above the clast-rich portion. Outsize mudstone clasts to 6 m in length "float" within the sandstone at all levels. Large blocks (up to the size of train boxcars) of well-lithified mudstone and burrowed siltstone (with Scolicia) rest upon the basal erosional surface or fill in basal lows (as opposed to the outsize clasts floating in the amalgamated pebbly unit). Most of these blocks simply represent rock-
Figure 44. Characteristics of an Eocene submarine canyon tributary erosional surface and fill.

A) Nearshore Delmar Formation and Torrey Sandstone deposits were truncated, forming an irregular submarine canyon floor that is plucked, stepped, and scoured. Erosional remnants project into the overlying fill, and injection features, prynups, and canyon-wall rockfalls filling topographic lows are common.

B) The submarine canyon fill is tripartite. A basal amalgamated pebbly sandstone is composed of cross-cutting units containing intraclast-rich bases grading into laminated fills with outsize clasts. The overlying laminated sandstone becomes convoluted upward, with a concomitant increase in dish structures, fluid-escape pillars, flamed mudstone laminae, and clastic dikes (sedimentary structures are diagrammatic and not to scale). Cut-and-filled hemipelagic mudstone channels cap the sequence, and contain siltstone and sandstone layers that are variably graded, rippled, cross-bedded, bioturbated, and/or fossiliferous.
CHARACTERISTICS OF
SUBMARINE CANYON FLOOR

TRANSPORT

A

Plucked
Canyon
Floor

Foundered Shoreline Deposits

Debris-Filled
Channel

Erosional
Remnant

CHARACTERISTICS OF
SUBMARINE CANYON FILL

LAMINATED TO
CONVOLUTED S.S.
Flamed Mudstone
Laminae

Bioturbated
Sandstone

HEMIPELAGIC
MUDSTONE
S.S. Plerks
Dish
Structures
& Fluid Escape
Pillars

AMALGAMATED
PEBBLY SANDSTONE

Cgl. Pocket
Injection

Digludged
Clast

20 m
falls and slides off the canyon wall.

A second facies deposited within this environment is the planar-to convolute-laminated sandstone which overlies the amalgamated pebbly sandstone. The striking textural differences between these two facies (Figures 29 and 30) has already been discussed. A fundamental hydrodynamic change is represented, and described later. Obviously this facies was overcharged with pore water whose expulsion produced box-like convolutions, dewatering pipes, dish-and-pillar structures, and clastic dikes.

The third facies included within this environment is the uppermost channelized mudstone (with siltstone and sandstone beds). Channels to 0.5 km wide and 75 meters deep are filled with bathyal, hemipelagic mudstone (Figure 17b). Thin beds of bioturbated, cross-bedded, rippled, and graded siltstone and sandstone attest to a variety of depositional processes. These were dominantly directed down canyon, but evidence of up-channel flow is also displayed. These processes are discussed later. The shelf facies described above prograde out over this depositional sequence. Intervening passive upper-slope deposits are not recognized; this canyon tributary probably headed directly into the shelf environment. An analogous situation is the modern Scripps Canyon whose head is within 60 meters of the surf zone (Shepard and Dill, 1966).

Slope Deposits

The hemipelagic material filling the channelized, upper submarine canyon tributary system is recognized inland in unchannelized sequences where it gradationally overlies and underlies outer shelf deposits.
It is considered to be "passive" slope deposits where slumping and channeling did not dissect the continental margin. Lohmar (1978) described this environment from the east wall of Rose Canyon (my Monongahela Drive section), where a mixed inner abyssal and neritic foraminiferal assemblage was recovered.

A base-of-slope transitional environment is displayed south of Canyon #4 to Scripps. It is characterized by classical turbidite sandstones, large and laterally-extensive olistostromal accumulations with flat bases, and wedges of pebbly mudstone. These features are considered diagnostic of the base-of-slope environment (Stanley, 1969; Stanley and Unrug, 1972; Mutti and Ricci Lucchi, 1972; Nelson and Kulm, 1973). Lohmar (1978) defined these units as lower slope based on an inner abyssal foraminiferal assemblage and the sedimentary features listed above. However, based on standard terminology used in published works, this environment is better designated as base-of-slope.

Inner-fan Deposits

A 1 km-wide inner-fan channel is exposed at Canyon #4, cutting into base-of-slope units (Plate 2, Appendix III). Lohmar (1978) considered deposition of the thick basal sequence of amalgamated, clast-supported conglomerate and massive to laminated sandstone facies the result of a meandering channel thalweg current. This channel fill extends inland; it was measured at Genesee Avenue North and its transition to a submarine canyon was examined in Los Peñasquitos Canyon (where the basal conglomerate layer becomes much thinner, the massive to laminated sandstone is approximately 50 meters thick, and
Jurassic volcanic remnants border and protrude into this fill. This system grades imperceptibly from an inner-fan setting into a major submarine canyon tributary. The amalgamated pebbly sandstone/laminated sandstone/channelized mudstone triad exposed from Bathtub Rock Trail south to Canyon #3 thus represents only one of numerous large and small canyon tributaries which fed into the Eocene submarine canyon-channel system.

Johnson and Walker (1979) and Hein and Walker (in review) have further subdivided coarse-grained inner-fan environments into a main thalweg channel with braid bars, marginal terraces with point bars and secondary channels, and high terraces of massive sandstones and classical turbidites (Figure 45). The inner-fan channel exposed at Canyon #4 is much narrower and thinner than that described from the Cap Enragé Formation (10 km wide and 300 meters deep). However, the complete fill sequence does contain two positive megasequences (best exposed on the south side of Canyon #4). The lower appears to be the result of progressive burial of a main channel (amalgamated, scoured, disorganized conglomerate) by marginal terrace (flat based, inclined and stratified conglomerate plus isolated conglomeratic channels) and then higher terrace (massive to laminated sandstone and laminated siltstone) deposits. The upper megasequence represents reestablishment of an active braided channel.

Inner-fan channel-levee deposits are exposed in the upper beach cliffs just south of the Hang Glider Port, lateral to the channel-fill sequence. Interbedded/lenticular-bedded sandstone and mudstone
Figure 45. Conceptual reconstruction of the inner-fan channel from the Cap Enragé (Hein and Walker, in review). Abbreviations are explained in the interpretative cross section below; L. A. indicates lateral accretion. A fining- and thinning-upward sequence (due to lateral channel migration) is sketched.
and interlaminated sandstone, siltstone, and mudstone facies with starved ripples and "lam-scram" sequences appear to be the result of overbank deposition. Planar-laminated sandstone beds with laminae dipping northward, rippled sandstone and starved ripples, graded sandstone layers, mudstone drapes, slurried units, and "lam-scram" sequences are evidence of periodic turbid suspension and tractional flow directed away from the main channel. Similar sequences have been attributed to overbank levee deposition (see, for example, Nelson and Kulm, 1973).

Lohmar (1978) instead assigned inner-fan levee deposition to fossiliferous sandstone, siltstone, and mudstone beds just beneath the southern portion of the Hang Glider Port. This interpretation is not supported by the field or laboratory evidence. These units consist of fossiliferous mudstone and sandstone and bioturbated sandstone and siltstone. Instead these deposits resemble outer- and inner-shelf sequences exposed northward and inland from these cliffs. Fossils collected in this area by Lohmar (1978) include vermetid worm tubes, oxidized diatoms, arenaceous foraminifers, and the calcareous benthonic foraminiferan *Eponides mexicana*. This assemblage indicates a middle to outer neritic environment (Gonzaga, 1977, written communication to Lohmar). Overlying this, samples with *Haplophragmoides sp.* were interpreted by Gonzaga (as above) to represent a shallow-water environment. These shelf units are also laterally separated from the inner-fan channel system by a major fault which uplifted the inner-fan channel sequence from a previously much lower stratigraphic posi-
tion (inner-abyssal foraminifers are present in the mudstone under-
lying the inner-fan channel).

Middle-fan Deposits

South of Canyon #4 to Scripps, isolated and amalgamated channel-
ized sandstone and conglomerate are separated by overbank thinly-
interbedded/lenticular-bedded sandstone and mudstone facies and by the
laminated siltstone facies (Plate 2, Appendix III). Lohmar (1978)
assigned these units to an inner-fan setting, but the presence of
cross-cutting braided sequences, radiating paleocurrent patterns,
extensive overbank units, slurried units, and pebbly mudstones indicate
that assigning these facies to an inner mid-fan distributary system is
more consistent with terminology of Normark (1970) and Walker (1978).
The vertical stratigraphic sequences are also more closely aligned
with models developed for deposition on channelized mid-fan areas
(Figure 46). Unchannelled suprafan lobes are not recognized.

Textural characteristics of mid-fan channels may also be used to
distinguish them from inner-fan channels (compare Figures 31 and 32).
The mid-fan channel deposit studied is finer-grained, better sorted,
and lacks the coarser components present in the massive to laminated
fill of the inner fan channel. Hydrodynamic sorting and winnowing
has progressively enriched the sands in finer-grained components down-
system. The complete 1 to 3 $\phi$ size range is represented in the submarine
canyon tributary fill, plus two coarser modes are present. These
coarsest modes were not transported down-canyon, and removal of much
of the 2 to 3 $\phi$ interval further out onto the fan enriched the massive
Figure 46. Hypothetical submarine fan stratigraphic sequence produced by fan progradation (Walker, 1979). C. T. = classical turbidite, M.S. = massive sandstone, P.S. = pebbly sandstone, CGL. = conglomerate, D.F. = debris flow, and SL. = slump. Arrows show thickening- and coarsening-upward (C-U) sequences and thinning-and fining-upward (F-U) sequences.
t to laminated sandstone of the inner-fan channel in the 1 to 2 φ component. Most material finer-grained than 2 φ was winnowed, with only a minor amount deposited from suspension in the inner-fan channel; the remainder was moved out into the mid-fan system and deposited by traction plus suspension.

Lohmar (1978) assigned the Mount Soledad Formation at Tourmaline Surfing Park to a fan-fringe environment. The evidence instead indicates that this is simply a mid-fan channel deposited in an environment slightly more distal than those present north of Scripps. Channelized conglomerate and sandstone define major mid-fan distributaries; however, the predominant deposition at Tourmaline consists of syndepositional slumps, cross-bedded sandstone, "lam-scram" sequences, and lenticular-bedded ("starved-ripple") sandstone and mudstone. These units pinch out onto the flanks of the one distributary channel present, and rapidly thicken away from it. All of this evidence indicates the dominance of overbank processes associated with deposition on a mid-fan levee. Lower reaches of the Astoria Fan inner-fan channel display levees that are similarly dominated by rippled and rhythmically-laminated silts, plus thin sand beds and lenses (Nelson and Kulm, 1973). Mutti et al. (1975) diagrammatically displayed mid-fan levee deposits with abundant slumping down levee margins strikingly similar to that at Tourmaline Surfing Park (Figure 47 and Plate 3, Appendix III).

Another factor arguing against Lohmar's (1978) interpretation for this sequence is that fan-fringe deposits are defined as being formed by non-channelized turbidites and laterally continuous lutitic and
Figure 47. Facies relations and terminology of channel and interchannel deposits, based on observations of mid-fan deposits from the Eocene Hecho Group, south-central Pyrenees (from Mutti et al., 1975). Channel-fill and channel-margin facies are arranged in distinctly fining- and thinning-upward cycles, and are variably composed of conglomerates, pebbly sandstone, massive sandstones, thick- or thin-bedded turbidites, mudstones, and/or slurried units. Interchannel units mainly consist of thin-bedded turbidites and mudstone. The intervening levee materials gradationally incorporate both depositional environments and styles, but also provide a unique topographic and hydrodynamic setting.
sandy beds (Ricci Lucchi, 1975). Marginal fan deposits should also form the lower part of negative megasequences, whereas positive megasequences are evident at Tourmaline Surfing Park.

It is possible that all of these units are contained within a much larger channel-form whose axis is buried to the south, and that the slumps simply slid down toward the axis. However, to be consistent with field relationships observed at Tourmaline, progressively finer-grained fill would then have been deposited in this major channel's axis, and coarser material would have been deposited away from its axis - the exact opposite of what is typical.
Inland Area

Summary

The depositional environments represented by the facies exposed inland include a submarine canyon-channel system, passive slope, inner to outer shelf, shelf channels, transitional offshore to nearshore sequence, and fan delta (Figure 50). Extending inland from Canyon #4 to Genesee Avenue North and then to Los Peñasquitos Canyon is a major inner fan channel/submarine canyon. Adjacent to this system, canyon tributaries/outer shelf channels cutting slope and outer shelf deposits are present at Genesee Avenue South. A similar setting is represented by the Soledad Canyon section.

Further inland, a regressive sequence is formed by the progradation of the Stadium Conglomerate fan-delta system. This cuts down into a variety of paralic deposits at Miramar Road and Rose Canyon. A similar sequence is evident at San Clemente Canyon Road, where shoreface Friars Formation is developed on top of a regressive offshore sequence.

A different type of regressive sequence is present at the Balboa Avenue, Monongahela Drive, and Torrey Pines sites. This is a progradational, outer to inner shelf, negative megasequence developed on top of passive slope deposits. This sequence is also partially developed along the upper sea cliffs from Bathtub Rock to south of the Hang Glider Port.

Older units demonstrate a transgressive (retrogradational) progression. Mount Soledad Formation fan-delta deposits at Morena Boule-
ward interfinger with and are covered by marine Ardath Shale units. The continuation of this positive megasequence occurs at the nearby Price Club and Pacific Theater locales. Channelized sandstones give way to nearshore, then offshore, and finally shelf units. A synchronous (?) transgressive sequence is evident at Canyon #3.

**Fan-delta Deposits**

The Mount Soledad Formation at Morena Boulevard and Stadium Conglomerate at Rose Canyon and Miramar Road are interpreted as portions of fan (tectonic) deltas. Fan deltas are simply alluvial fans which prograde into a standing body of water from an adjacent highland (Holmes, 1965). They are especially common along convergent continental margins in areas of high seasonal rainfall (Wescott & Ethridge, 1980), both conditions that existed in the San Diego area Eocene. Kennedy and Moore (1971a) and Howell and Link (1979) came to a similar facies interpretation for these units.

The conglomerates of these formations are typically amalgamated, flat-based, stratified, and normally and inverse-to-normally graded, with well-imbricated clasts. Similar fabrics are described from both recent (Wescott & Ethridge, 1980) and ancient (Nemec et al., 1980) fan deltas. The lack of laterally-continuous interbedded sandstone indicates that deposition occurred within a proximal-fan setting (McGowan and Groat, 1971). However, the lack of narrow, steep-margined channels, except at Miramar Road, indicate that most deposition took place in the outer (distal) portions of the proximal fan. This depositional site is characterized by sedimentation during floods forming graded and crudely-
bedded longitudinal bars and gravel sheets on a low relief plain (Harms et al., 1975; Howell and Link, 1979). The inverse and inverse-to-normal grading and very low matrix content is also typical of more distal proximal-fan deposition (Nemec et al., 1975).

Associated sandstones are typically flat-based, with scoured upper surfaces, and internally laminated. These lenses probably represent bar deposits, an interpretation supported by the textural data. The lack of a residual component indicates removal from lag material. Tectonic deposition and then rapid fallout from suspension with some tectonic surges, as indicated by the grain-size frequency curves, are typical of deposition on bars. This style of sedimentation produced the laminated internal structures. Sandstone bodies with concave-up bases are subordinate, and probably result from deposition within the braid channels. The mineralogic immaturity (e.g. high plagioclase to potassium feldspar ratio), textural immaturity, and high heavy mineral content in these sandstones all indicate rapid, high energy deposition near the source.

Nearshore Deposits

The terminology for nearshore and offshore environments is somewhat contradictory and confusing. Figure 48 is an attempt to provide a consistent set of terms to be used in this paper. The laminated sandstone with "perched" cobbles is interpreted as a beach deposit, specifically foreshore and berm environments. Texturally the unit is not well-sorted. However, in comparing frequency curves of this facies and its presumed source (the fan-delta sandstone), some winnowing has
Figure 48. Diagrammatic nearshore and offshore profile and the corresponding facies zones. Wave activity and transformations affecting the bedforms across these environments are also shown. Taken from numerous sources (see text).
evidently taken place. Wave action in the foreshore was probably the agent responsible for removal of the fine-grained component. An incipient bimodality present in the frequency curve is also consistent with foreshore deposition (Visher, 1969). The facies is horizontally laminated to slightly inclined, with pockets of and isolated cobbles completely surrounded by small scoured zones. Heavy-mineral concentrations are intermediate between that of high-energy, channelized deposits and thin-bedded sandstone interbedded with finer-grained sediments present elsewhere in the field area.

Similar internal structures are found in beaches along the Yallahs Fan Delta, Jamaica (Wescott and Ethridge, 1980), as well as along the present San Diego area beach. At the Jamaica locality, cobbles in paralic zones are most common in berm and lower foreshore deposits and the beach is essentially a lag, composed of fluvially-emplaced material minus fine-grained detritus lost by winnowing. These textural characteristics are analogous to those of the Miramar Road units.

Interbedded with this facies are muddy sandstone units which display some disrupted rippling and complete bioturbation. Similar burrowed intervals associated with nearshore deposition are reported from ancient and modern environments; both lower shoreface-upper offshore and tidal flats are areas of intense bioturbation (see, for example, Reineck and Singh, 1975, p. 316, p. 359-363; 1978b, p. 148-153, p. 174). The association with foreshore deposits and the rippled appearance indicate deposition of these beds in a sand flat or mixed mud flat.

Another nearshore environment is represented by steeply-dipping,
planar cross-bed sets of coarse-grained sandstone in the Friars Formation at San Clemente Canyon Road. This probably is the product of a ridge-runnel system associated with the breaker zone at the shoreface-foreshore transition. Cobble layers inclined along the dipping sets attest to the high energy of this environment. A nearshore setting is indicated by the presence of fragmented, thick mollusk shells typical of shallow marine habitats (see below). Davies et al. (1972) and Davidson-Arnott and Greenwood (1976) describe similar high-angle cross-beds which dip landward and result from the migration of foreshore ridges. Planar, landward-dipping cross-beds with inclined pebble layers have also been found along the nonbarred high-energy shoreline off Oregon (Clifton et al., 1970).

**Offshore Deposits**

The pinch-and-swell sandstone facies is stratigraphically associated with many of the nearshore facies but is distinctly different. Lensoid sandstone bodies with conspicuous gentle trough cross-bedding, planar laminations, and an upper burrowed surface are draped by fine-grained material. Upper surfaces dip in all directions. Texturally these units have a coarse-grained cut-off at about 1.25 φ, similar to that of the foreshore deposits. However, this offshore facies is more poorly sorted, with well-developed lag and suspension components. This facies is interpreted to be lower shoreface or shoreface-offshore transition bars formed below fair-weather wave base, probably in response to storm conditions.

The abundant *Gyrolithes* and simple, large, vertical mud-lined
Ophiomorpha burrows in this facies are typical of shallow water deposits (see, for example, Boyer and Warme, 1975), and trough cross-beds with some horizontally-laminated intervals are typical of nearshore bars produced by lunate-megaripple migration (see, for example, Davidson-Arnott and Greenwood, 1976). However, periods of decreased wave energy are indicated by fine-grained draped layers. Each succeeding storm period produced sets of bars which filled in previous microtopographic lows. The poor sorting and low sand content in these beds also argue for periodic high-energy lag deposition interspersed with quiet water fallout from suspension. However, tractional peaks are dominant in the frequency curve.

Large-scale sand ribbons, sand waves, and tidal current ridges have all been described from modern and ancient shelves (see, for example, Reineck and Singh, 1975; Walker, 1979), but examples of small-scale bar-like features encased in a fine-grained matrix are noticeably lacking. The closest approximation is hummocky cross-stratified sandstone with intervening thinly-interbedded shales and sandstones (Harms et al., 1975; Hamblin and Walker, 1979; Bourgeois, 1980; Wright and Walker, 1981). These have been interpreted as gently swaled sandstone deposits formed by storm-wave migration then draped during hydraulically quieter times. While morphologically similar, hummocky cross-stratification is internally very different, composed of: 1) sets of low-angle erosional lower surfaces, 2) laminae parallel to and immediately overlying these surfaces, 3) upward thickening laminae with a concomitant decrease in surface dips, and 4) scattered dip directions. This deposi-
tional style is interpreted as the result of scour and deposition by currents with a variety of orientations. Perhaps with well-oriented storm waves and high sediment volume, directed lunate-megaripple migration below fair-weather wave base will yield trough cross-beds and a swell-and-swale sea floor. As wave energy diminishes, lower plane beds and rapid fallout from suspension are dominant. This could be the scenario for production of pinch-and-swell sandstone. Indeed, similar low sandy swells with intervening mud and clay-filled swales have been noticed offshore on the Texas Gulf Coast, but have not yet been internally sampled (J. Anderson, oral communication, 1981).

Another type of offshore storm deposit has been documented in the literature, and is characterized by the conglomerate-based sandstone facies. Texturally these sandstones are almost equivalent to pinch-and-swell sandstone, arguing for analogous hydrodynamic conditions during formation. The basal cobble and fragmented shell layer probably represents storm lag, and draped interbeds of fine-grained material were most likely due to fair-weather deposition below wave base.

That these are indeed storm-generated beds rather than simply a different type of lag deposit can be demonstrated. Brenner and Davies (1973, 1974) interpreted similar laterally-extensive sheet-like beds with planar erosive bases and non-erosive tops as storm deposits. In contrast, shallow-marine channel lags should line concave-up sandstone bases, and contain a low diversity community of whole to completely comminuted shell debris. Shelf storm-swell lags are often of monospecific basal shell layers and much thinner bedded. Transgressive lags (Clifton,
1973 and 1981) are yet another possibility. However, the examined units are more rapidly repetitive and do not occur at the tops of shoaling units. And finally, intervening fine-grained beds argue against deposition as a beach layer (see, for example, Bourgeois, 1980).

The shallow-marine nature of these deposits is indicated by abundant fragments of thick-shelled bivalves indicative of shallow sublittoral environments (Table 7). These include Ostrea, Crassatella, Glycymeris, and Venericardia. A similar assemblage, but far less abundant, is found in the steeply-inclined planar cross-bedded Friars sandstone.

A common sedimentary sequence described from storm lag deposits is due to decreasing flow regime: basal gravel giving way upward to evenly laminated sand, then to bioturbated or wave-ripple cross-laminated sand (Sanders and Kumar, 1975a; Kumar and Sanders, 1976). Cross-laminations, climbing ripple cross-laminations, and grading have also been described (Johnson, 1978, p. 251). As in the pinch-and-swell sandstone, the lower cross-bedded portion in Rose Canyon units probably represents storm wave-generated lunate megaripple migration over the basal lag. Overlying horizontal laminations could then represent lower flat bed and intermittent suspension deposition in the waning stages of flow. Upper surfaces with small trough cross-beds represent the last stage of deposition by storm wave-ripples.

At San Clemente Canyon Road, the trough cross-bedded interval is replaced by low angle, cross-cutting, wedge-shaped sets of planar laminations and by horizontal planar laminations. The latter is easier to explain, possibly formed by upper flow-regime, flat-bed
deposition. The former superficially resembles hummocky cross-stratification (Harms et al., 1975) or some figured examples of truncated wave-ripple laminae (Campbell, 1966), but lacks the low-angle curved and upward-doming laminae. These structures also are not simply low-angle trough cross-beds because there is no asymptotic pinch-out of cross-laminae. While the mechanism of formation is not known, these superficially appear to be produced in a like manner to hummocky cross-stratification, although lacking the domed accretionary deposition. Perhaps this indicates higher energy and the dominance of erosional over depositional processes.

The deepest offshore coarse-grained deposits are represented by turbidite sandstone sequences at Miramar Road and Rose Canyon. Major storms produce storm-surge tides that erode and entrain sands from the nearshore environment. Once the storm abates, a seaward-flowing density current may be generated (Walker, 1979). Below storm wave base, a turbidite with typical Bouma (1962) divisions would then form. A scenario such as this may have produced the observed beds, or input of sediment-laden flood waters from the contemporaneous fan-delta/fluvial system may have provided the necessary density flow.

Lower-energy, offshore deposits are represented by: 1) interlaminated sandstone, siltstone, and mudstone facies dominated by lenticular bedding and bidirectional ripple cross-laminations, and 2) laminated to burrowed sandy siltstone facies with small-scale "lam-scram" sequences. These probably represent normal low-energy regime deposition between storm wave base and fair-weather wave base. These facies
are transitional with one another; the former is most often interbedded with sandy intervals and probably represents slightly higher energy and coarser sediment input.

Davies et al. (1971) assigned bioturbated mudstones and siltstones containing isolated ripples to a lower shoreface environment in the Lower Cretaceous Muddy Sandstone. However, offshore bioturbated sandstones and siltstones are described from Upper Cretaceous rocks of Utah as well as from recent material off Fire Island and Sapelo Island (Elliott, 1978b, p. 165). The concomitant decrease in grain size and ripple structures indicates an offshore decrease in bottom current activity (Reineck and Singh, 1977). Bidirectional ripple senses indicate the interplay of sediment movement seaward during onshore winds and short-period waves, and landward during longer-period swells (Brenninkmeyer, 1978). "Lam-scram" sequences in both recent and ancient silty offshore sands have been documented by Howard (1972), Howard and Reineck (1972), and Bourgeois (1980). These indicate periodic low-energy sediment influx which punctuates periods of non-depositional biogenic reworking.

A final, somewhat enigmatic, offshore facies noted is channelized sandstones with shell hash, rip-ups, and more rarely cobble lag bases. These overlie and cut into the fine-grained offshore materials described above; spectacular soft sediment deformation and convoluted and flamed structures are present in the fine-grained material directly underlying these sandstone channels. Fill sequences are typically graded to laminated to rippled, and lack cross-beds. A variety of paralic
channels are possible models, including tidal, rip-current, delta front, and longshore channels. Longshore channels are an ephemeral feature, destroyed by the migration of nearshore bars. These are also the sites of fine-grained sandstone and siltstone accumulations, with coarser material restricted to the associated bars. For these reasons, longshore channels are not considered as possible models.

Tidal channels are often sinuous and form by lateral migration of the cut bank and corresponding lateral accretion on the depositional bank. The point bar is typically composed of thin, interlaminated clay-silt and sand gently dipping into the channel (Elliott, 1978b, p. 175). Other tidal channels in sandy environments extend seaward from estuarine systems. These may contain sand-wave, megaripple, or ripple bedforms and a fill of medium- and large-scale cross-bedding (Elliott, 1978b, p. 175). Still other tidal channels cut across barrier islands and shoals, and these too migrate laterally (partially in response to longshore drift). Previously deposited barrier sands are eroded, and lateral accretion takes place on the depositional bank. Interfingering dunes and sandwaves are also characteristic of this system (Elliott, 1978b, p. 152-156). None of these sequences is seen in the channel facies exposed at the Price Club and Pacific Theater localities.

Therefore, a rip-current or delta front channel origin is favored. Little work has been carried out on modern rip-current processes (McKenzie, 1958; Ingle, 1966; Cook, 1970), and even less has been published on possible ancient examples. Rip currents head in the surf
zone where they are narrow and scoured to 1 to 3 meters. These channels extend through the longshore bars out into the offshore zone. The currents carry sediment out into the offshore zone (Brenninkmeyer, 1978). In Southern California, very large rip-current channels may extend more than a kilometer offshore, transporting abundant material onto the continental shelf (Cook, 1970). These channels are floored by lag deposits of coarse sediment, shells, and mudstone rip-ups, and transport material via traction and suspension through the surf zone. Past this area, material up to 1.5 $\phi$ in diameter has been found being carried by suspension within the rip-current channels (Cook, 1970).

In the geologic record, corresponding rip-current channel fill is expected to consist of elongate bands of coarse sand and shell fragments, and with average widths of 5 to 15 meters, thicknesses of 1 to 3 meters, and lengths of 25 to 100 meters (Cook, 1970). In the near-shore surf area, this fill would be cross-bedded, deposited by sand waves or large ripples. Offshore, graded sequences related to high-energy periods would be interspersed with massive to laminated normal-energy deposition.

This work in the modern is consistent with the interpretation of these offshore channels being of rip-current origin. However, most rip-current channels are ephemeral features unless hydrodynamically related to geomorphic controls, such as submarine canyon heads or seaward-projecting headlands. Therefore, submarine channels in front of the fan delta must also be considered as possible sites of origin for these features. Where river outflow enters marine basins, hypo-
Pycnal flow is typical, carrying fine-grained material further seaward in a sediment plume. Sand-sized detritus rapidly drops out of suspension in distributary mouth bars. Waves generally rework the coarse-grained sediment onto laterally-continuous sand sheets (see Elliott, 1978a). However, lenticular sand units representing subaqueous extensions of distributary channels may occur. The carboniferous Kinderscout Delta of the Central Pennine basin displays such a delta-front sequence of steep-sided, sand-filled channels encased in mudstone and siltstone. Further offshore, these channels were filled with turbidites, but closer to distributary mouths, sequences of erosive bases overlain by massive-bedded sandstone, then horizontally-bedded sandstone, and finally cross-bedded sandstone were developed (Elliott, 1978a, p. 136).

Thus, these ancient-channel examples resemble those exposed at the Price Club and Pacific Theater localities. However, deltaic cycles are generally coarsening- and thickening-upward, representing progradation; the channels of possible delta-front origin in my field area are instead associated with thinning- and fining-upward sequences. Little work has yet to be carried out on ancient deltaic deposits which are not fluvially-dominated, and no adequate facies model has been developed for the poorly-studied fan-delta system. Therefore, determining an unequivocal origin for these channels is probably impossible. Indeed, they may have even been transitional feeder channels connecting the deltaic distributary complex directly to the submarine canyon head.

**Continental Shelf Deposits**

The typical suite of shelf facies has been covered above in the
'Beach Cliff Transects' section. However, a different environment of deposition is present inland associated with shelf sequences. These are very subtle channelized sandstone units with either massive or laminated sandstone fill. Monospecific layers of articulated _Acila_ and whole, unabraded _Turritella_ indicate a neritic environment of deposition (Table 7). 

Texturally these units resemble offshore storm deposits, but have more distinct traction modes (Figure 28). The presence of traction-deposited, coarse-grained layers, with irregular erosive bases and rip-ups, argue for an upper flat-bed flow regime and/or mass flow processes. A wide variety of channels, many acting as feeders for submarine canyons in both ancient and modern shelf settings, have been reported. Possible mechanisms of formation and transport are discussed later. The interested reader may find examples and further information in Whitaker, 1962; Sedimentation Seminar, 1969; Galloway and Brown, 1973; Jeletzky, 1975; Lewis, 1976; Haner, 1979; and Dott and Bird, 1979. 

The remaining shelf facies exposed at the inland sections have been detailed above, and will not be repeated here.
Shallow- to Deep-marine Facies Transitions

A series of three stratigraphic cross-sections along the depositional dip have been constructed and correlated into a fence diagram to provide insight into regional paleogeographic transitions with time. Individual stratigraphic sections are projected into the lines of dip. The southernmost Dip Section I (Figure 49) presents progradation of the Mount Soledad fan delta in Penutian (Early Eocene) time, then a later retrogradational phase. Mid-fan channel levee deposits at Tourmaline Surfing Park have been correlated with Ardath Shale slope deposits, indicating deposition slightly later than the Mount Soledad fan delta. The Price Club and upper Morena Boulevard stratigraphic sections display this retrogradation of the Ardath shelf and nearshore units over the Mount Soledad fan delta. Further north and environmentally farther inland (see Canyon #3, Figure 50) Delmar and Torrey units retrograded over a correlative and previously subaerially-exposed Mount Soledad fan-delta front. A second progradational stage is represented further inland on Dip Section I at Balboa Avenue, where Scripps Formation offshore, then inner shelf, units were deposited over the Ardath Shale outer shelf.

The younger progradational event evident at Balboa Avenue is directly correlative with that at Monongahela Drive, Dip Section II (Figure 49). Scripps Formation inner-shelf units were deposited over the Ardath Shale outer shelf and slope. Further inland, some minor progradational-retrogradational couplets give way to the progradation of the late Ulatisian to Narizian (late Middle Eocene to Late Eocene)
Figure 49. Dip sections across the Eocene basin margin showing facies relationships and stratigraphic development. The scale given on Dip Section II pertains to all three sections. Datum is sea level. Relative locations of the numbered stratigraphic sections are shown on Figure 50, and include:

Dip Section I
10 - Tourmaline Surfing Park  15 - Morena Boulevard
14 - Price Club  17 - Balboa Avenue

Dip Section II
9 - Torrey Pines Road  26 - Rose Canyon
13 - Monangahela Drive  21 - San Clemente Canyon
19 - Miramar Road

Dip Section III
6 - Canyon #4  12 - Genesee Ave. South
11 - Genesee Ave. North  18 - Soledad Canyon

Two regressive sequences punctuated by a transgressive phase characterize the Middle Eocene sedimentation. Regressive trends produce large-scale negative megasequences as coarser-grained, landward units prograde over finer-grained, basinward deposits. In contrast, a general positive megasequence is preserved due to transgression and depositional retrogradation.
Figure 50. Fence diagram showing areal stratigraphic development for the Eocene basin margin exposed north of San Diego. The upper surface represents present-day erosional truncation; dashed intervals indicate probable facies extensions prior to erosion. Locations for the stratigraphic columns used in this reconstruction and for the dip sections of Figure 49 are shown in the insert, and include:

1 - Torrey Pines Reserve  14 - Price Club
3 - Canyon #1  15 - Morena Boulevard
5 - Canyon #3  16 - Pacific Theater
6 - Canyon #4  17 - Balboa Avenue
9 - Torrey Pines Road  18 - Soledad Canyon
11 - Genesee Avenue North  19 - Miramar Road
12 - Genesee Avenue South  20 - Rose Canyon
13 - Monangahela Drive  21 - San Clemente Canyon
Stadium Conglomerate fan delta.

Along the northernmost Dip Section III (Figure 49), a basal progradational sequence is overlain by headward retrogradation (erosion) of the Scripps Formation submarine channel-canyon fill. A second progradational phase is related to the second positive megasequence present at Canyon #4 that reestablished braided thalweg inner-fan channel deposition. The intervening retrogradational stage is younger than that associated with Delmar, Torrey, and Ardath units onlapping the Mt. Soledad Formation fan delta.

Construction of a fence diagram (Figure 50) provides for the areal correlation of depositional sequences obvious on the dip sections. The Genesee Avenue North submarine channel-fill units correlate to those of Canyon #4, producing a positive megasequence. This overlies and truncates a negative megasequence composed of slope to shelf units exposed at Genesee Avenue South and Soledad Canyon (and at Torrey Pines Road). This canyon-channel fill also cuts nearshore units exposed from Torrey Pines Reserve south to Canyon #3 along the Torrey Pines-Scripps beach-cliff transect (see below).

Fossiliferous coarsening-upwards shelf-units capping the Ardath Shale from Canyon #1 south to Canyon #4 produce a younger negative megasequence, and occupy similar stratigraphic positions at Monanganahela Drive, Torrey Pines Road, and Balboa Avenue. This negative megasequence is represented further inland by offshore deposits overlying coarsening-upwards shelf units at San Clemente Canyon Road and Balboa Avenue, and by fan delta progradation over nearshore and offshore deposits at Miramar Road and Rose Canyon. Older fan delta sequences exposed at Price Club and Morena Boulevard and at Canyon #3 are capped
by retrogradational units also present at Pacific Theater, thus producing a positive megasequence.

Correlation of these depositional sequences suggests that the broad Middle Eocene transgressive-regressive cycle recognized by Kennedy (1975), and represented by retrogradation of marine units over the older Mt. Soledad fan delta and then later progradation of the younger Stadium fan delta and associated paralic to shelf deposits, was punctuated by a minor cycle. Such a minor regressive-transgressive cycle has been recognized worldwide during the Middle Eocene (Vail and Hardenbol, 1979), although it may be slightly older in the southern California region (Howell, 1980). A more complete discussion of the significance of this stratigraphic development is given below.

Depositional Timing Along the Torrey Pines-Scripps Transect

The same minor transgressive (retrogradational) punctuation of the Middle Eocene sequence is evident along the Torrey Pines-Scripps beach-cliff transect. Timing for this event is provided by the nannoplankton data (Figure 51). Much of this interpretation is similar to Lohmar's (1978). An Early Eocene (pre-Rhabdosphaera inflata nannoplankton subzone) weathering profile present at the base of Canyon #3 represents non-marine erosion, during which a fluvial channel may have truncated subaerially-exposed shelf units further basinward. During deposition of the ensuing retrogradational Delmar-Torrey-Ardath sequence, a submarine canyon previously initiated as this fluvial system probably eroded headward much like the Scripps
Figure 51. Torrey Pines - Scripps beach-cliff transect restored to the interpreted pre-fault configuration. The nannoplankton subzones and intervening isochrons are diagrammatically shown. Isochrons dip basinward, representing progressive filling of the submarine canyon system and basinal deposition.
Submarine Canyon has done during the Holocene sealevel rise (see Shepard and Dill, 1966, for a detailed discussion of this and other examples of submarine canyon development). The tripartite fill of the canyon tributary exposed along the sea cliffs reflects a fining-upward trend deposited during this retrogradational phase. This sequence also correlates to the retrogradational event evident in Dip Section I (Figure 49), with deeper-water Ardath Shale facies onlapping the Mount Soledad Formation fan delta.

The mudstones filling the channelized submarine canyon tributary were for the most part deposited from the *Rhabdosphaera inflata* to *Chiasmolithus gigas* Subzones. However, the isochrons dip southward as increasingly younger nannofossil subzones are found in stratigraphically lower positions (Figure 51). The 1 km-wide inner-fan channel exposed at Canyon #4 cuts into base-of-slope mudstone apparently containing the *Chiasmolithus gigas* Subzone. It also cuts across part of an older large submarine channel. Thus, the period of cutting was post *C. gigas* Subzone or about medial Middle Eocene, and represents a punctuation in the broad Middle Eocene transgressive-regressive depositional cycle. Mid-fan facies then retrograded into the inner-fan channel, as part of this intervening cycle. This is the retrogradation seen in Dip Section III (Figure 49) as the inner-fan channel-canyon system became progressively filled landward. This ensuing stage was probably post-medial Middle Eocene.

The broad late Middle Eocene regressive phase is represented by progradation of Torrey Sandstone shelf deposits over the submarine canyon tributary channelized mudstone. Rejuvenation of braided inner-fan channel deposition is probably related to this
phase, and is evident on the cliff transect and Dip Section III. This regression is represented inland on Dip Section II (Figure 49) by the progradation of nearshore Scripps and Friars deposition over offshore Scripps units and development of the Stadium Conglomerate fan delta. At Balboa Avenue, Monongahela Drive, and Torrey Pines Road (Dip Sections I and II), inner-shelf facies apparently prograded out over outer-shelf sequences.

Provenance Variations in Time

Stratigraphic patterns of heavy-mineral abundances along the beach transect lend further support to the structural and stratigraphic beach-cliff reconstruction based on nannofossils and three-dimensional relations seen on outcrop. Kyanite relative abundances increase with time (up through the sections), and actinolite decreases with time (Figure 42). This occurs regardless of depositional environment. However, without stratigraphic reconstruction, sample 63 in Canyon #6 appears to refute these patterns for actinolite and kyanite. On outcrop, it seems to have been deposited stratigraphically lower than inner-fan channel samples 82 and 86. Yet upon removal of the apparent faulting, sample 63 becomes stratigraphically higher than the lower inner-fan channel megasequence of Canyon #4 (Figure 41), and is consistent with the observed heavy mineral patterns. This supports the reconstruction showing a minor transgressive sequence of mid-fan units which retrograded into the inner-fan channel during post-medial Middle Eocene.
These heavy-mineral trends also sustain the interpretation of a major uplift of the inner-fan channel relative to adjacent younger (hence originally stratigraphically higher) shelf units. These shelf units were interpreted as laterally-equivalent channel margin units by Lohmar (1978). However, there is an obvious discrepancy when comparing the relative abundances of actinolite and kyanite for the two. Kyanite composes an average of 10.5% of the normalized inner-fan channel heavy mineral suite versus 25% for the shelf layers, and actinolite makes up an average of 33.5% of the channel suite versus 5% for that of the shelf units. With an interpretation supported by the observed heavy mineral patterns, these clearly are not time-equivalent sequences, and the shelf layers are instead correctly placed above (younger than) the inner-fan fill.

To discover why these patterns are observed, the provenances of the major heavy mineral components must be determined. Three major provenances are suggested. Chlorite, actinolite, and epidote are the diagnostic assemblage of the greenschist facies, formed during regional metamorphism at medium grades (200 to 400° C; see Deer et al., 1975). These may originate from a variety of precursors, including basalts, tuffs, marls, pelites, and graywackes. The presence of schistose lithic fragments also indicates a low- to medium-grade metamorphic source. The obvious choice for the source area of these minerals is the nearby regionally-metamorphosed Santiago Peak volcanics, volcanioclastics, and sediments. The fresh, unabraded appearance of actinolite is further evidence for a nearby source. The greenschist mineral assemblage is more abundant in the
early Middle Eocene deposits, reflecting a dominant input from the local source rocks at that time.

In contrast, kyanite, a typical product of high-grade regional metamorphism of pelitic rocks (Deer et al., 1975), becomes more abundant later. Other evidence for a separate high-grade metamorphic source is abundant polycrystalline metaquartzite grains. Although a fairly stable detrital mineral, the grains of kyanite are noticeably abraded, also indicating a source somewhat further afield. Indeed no nearby high-grade metamorphic units are known. One probable source area is located near the Sonoran metarhyolites (which in turn are the probable source of the "Poway" clasts). Northern Sonora, southern Arizona, and areas east of the San Andreas in California all contain numerous outcrops of Precambrian, Lower Paleozoic, and other age quartzites (Minch, 1979). Black to blue-gray quartzites and argillites are prominent cobbles in northern Baja and represent a second possible high-grade metamorphic source. These also outcrop in the Sonoran region (Minch, 1979). Both metamorphic suites had pelitic precursors and could be the source for kyanite found in the Eocene samples.

Topaz is the remaining very abundant heavy mineral, and displays a consistent relative abundance. It is typical of acid igneous rocks, especially granites and rhyolites, and is often found in detrital sediments near acidic intrusions (Deer et al., 1975). The occurrence of this mineral may also indicate derivation from the metarhyolites of the Sonoran area (Minch, 1979), and/or an origin from nearby plutonic sources. The former is favored because the
Peninsular Range batholiths are typically quartzmonzonites and granodiorites, i.e. not acidic enough to expect abundant topaz formation.

Pettijohn (1975, p. 499) developed three models to explain heavy mineral zonations (i.e. variations in relative abundances) in stratigraphic sections. The first two, progressive denudation and unroofing of a source terrane and progressive tectonism of a source, cannot explain the observed sequence. As already stated, one single source cannot explain the observed mixed assemblage of high-grade and low-grade metamorphic minerals. Increasing intrasratetal solution is the third model; it is untenable because the low stability actinolites appear very fresh. Another line of evidence that progressive solution with depth of burial has not been a factor is that sample 63, while at the same elevation as numerous others, does not share the same relative abundances as those samples at its elevation. Instead it shares trends with sandstones much higher because it has been faulted downward; i.e., it maintains its true stratigraphically-controlled relative abundances regardless of structural position.

Thus the observed patterns reflect a systematic interplay of source terrane influx with time, probably related to a complex interaction of tectonic, climatic, and/or eustatic controls. Although more samples would need to be examined stratigraphically and areally in order to delineate these factors and their relative importance, some simple models may be put forth.

The abrupt appearance of the metarhyolitic clasts in lower Eocene strata may represent headward (eastward) fluvial downcutting and
stream piracy within the region east of the Peninsular Ranges (Howell, 1980). High-grade metamorphic units were introduced as an Eocene source. In contrast, the kyanite thought to be from the same Sonoran region appears in great abundance only much later. This may reflect a variety of possible scenarios:

1) The headward-eroding streams did not tap the specific kyanite source until medial to late Middle Eocene. Then kyanite deposition outpaced that of the greenschist mineral assemblage from the local source.

2) The kyanite source was continually being tapped, but increased mechanical abrasion due to fluvial transport along with the cobbles broke down the kyanite grains, and they were carried out further into the basin. At approximately medial Middle Eocene, cobble transport diminished, allowing for transport and deposition of kyanite in greater abundance.

3) As the local source terrain became worn down, its influence as a source decreased, and/or later increased uplift of the metarhyolite source caused a concomittant increase in the detritus shed into the drainage system and transported to the basin margin.

4) The changing climate from hot and humid to temperate and arid may have produced a change in: a) the volume of drainage or development/downcutting of the drainage system within the local vs. distant source areas due to changing weather patterns, thereby tapping the separate
source areas to varying amounts; or b) relative
dissolution of the various heavy minerals (i.e.
actinolite vs. kyanite) prior to deposition.

The fact that topaz remains relatively constant in proportion
to the other heavy minerals, and may be from the same source area
as the kyanite, argues against some of these scenarios. Comparison
to the heavy-mineral assemblages of the Cretaceous strata could
further restrict the possibilities. And as stated above, other
more complex models could more easily be developed based on further
data from a detailed areal investigation. Regardless of these
restrictions, the observed patterns are real, and are useful in
aiding in the structural, stratigraphic, and geochronologic
reconstruction of the facies development along the Torrey Pines-
Scripps beach-cliff facies.

Regional Paleogeography and Geologic History

Regional paleocurrent and facies patterns are consistent
with sediment input into the San Diego Embayment via fluvial and
fan-delta processes, and then redistribution in shallow marine, shelf,
and submarine canyon-fan settings (see following section). Similar
systems (Figure 52) are common in both continental and island arc
settings associated with convergent margins (Wescott and Ethridge,
1980).

The stratigraphic relations of the associated deep-marine,
shelf, shallow-marine, and fan-delta facies occur in consistent
patterns recognized throughout the study area. These were
Figure 52. Schematic diagram depicting the environments studied from the Eocene San Diego Embayment (modified from Link et al., 1979). Note that the submarine canyon heads across the narrow shelf into the paralic zone. One submarine mid-fan channel system is shown as abandoned and no longer an active depositional site.
detailed above for specific dip sections, and provide insight into the controls on regional sedimentation and sediment distribution. The model presented differs somewhat from that of Lohmar's 1978 and later publications (Lohmar and Warme, 1978, 1979; Lohmar et al., 1979).

Highstands of relative sealevel are the most likely times for trapping coarse-grained terrigenous clastics in deltas and along the associated paralic zones. Deposition occurs as clinoform lobes prograding across the shelf. In contrast, during sealevel lowstands the shelf can be exposed to subaerial erosion. Sediment may be carried directly to the shelf-slope break by fluvial channels and bypass the shelf, and/or previously-deposited, coarse-grained shallow-marine clastics may be reworked and funneled through submarine canyons onto the adjacent basinal submarine fan. During an ensuing sealevel rise, coastal onlap occurs (see Vail and Hardenbol, 1979). Such patterns are evident from the areal stratigraphic reconstructions described previously. Although variations in local sealevel and sediment input are responsible for minor modifications of the areal depositional development, eustatic sealevel changes appear to have dominantly controlled paleogeographic patterns and the resultant spatial and temporal stratigraphic sequences in an orderly fashion (Figures 53 to 58). Vail et al. (1977) have constructed these global cycles of sealevel change for Phanerozoic time based on simultaneous stratigraphic events from continental margins worldwide. Their charts of global sealevel variations are correlated to the stratigraphic development of the Eocene San Diego Embayment (Figure 53).
Figure 53. Correlation of worldwide sea level changes (from Vail and Hardenbol, 1979) to stratigraphic development within the San Diego Embayment. Deposition occurred as a hemicycle. A basal unconformable surface may be related to a major worldwide sea level fall. Thin retrogradation units above the erosional unconformity correspond to a eustatic transgression during the early Middle Eocene. Nearshore sediment may have accumulated during the Ulatisian highstand, then was flushed basinward (causing submarine fan progradation) during a minor medial Middle Eocene sea level drop. Submarine fan retrogradation corresponds to the ensuing sea level rise. During the late Middle to early Late Eocene stillstand, nearshore and fan-delta deposits prograded basinward.
Development of the Early Eocene Mt. Soledad fan delta corresponds to the high sealevel stillstand recognized by Vail et al. (1977). Fluvial valleys entrenched into the adjacent highlands transported very coarse-grained material, which then was trapped along the coastal plain to shelf setting. During the ensuing late Penutian drop of sealevel (Figure 53), the fan-delta front recognized at Canyon #3 underwent subaerial exposure. A grayish-white, quartz-enriched kaolinitic (?) weathering profile was apparently developed. This may actually be equivalent to the "pre-Eocene" lateritic paleosol widely distributed in southwestern California and adjacent Baja California (Peterson et al., 1975). If so, it pushes the maximum age of this soil development to late Early Eocene. The shoreline at this time could have been near the present coast between Torrey Pines and Scripps (Figure 54).

This sealevel lowstand also led to resedimentation of crystalline cobbles into the mid-fan channel studied at Tourmaline Surfing Park; these clasts were shed from an apparently separate alluvial fan-delta into a possible second submarine canyon-fan system not presently exposed (Figure 54). Evidence for an individual southern depositional complex not intimately connected with that exposed along the Torrey Pines - Scripps beach-cliff transect includes: 1) deep-marine "Poway" clasts underlying fine-grained mid-fan channel fill dated as equivalent to Ardath Shale in age (Steineck et al., 1972); 2) paleocurrents in the mid-fan channel directed back towards the northern canyon-fan system; and 3) a Cretaceous canyon-fan system precursor in this area (Yeo, in preparation).

Thus, Lohmar's (1978) interpretation that the Tourmaline Surfing
Figure 54. Late Early Eocene (Late Penutian) paleogeography of the San Diego Embayment (modified from Lohmar, 1978). During a eustatic drop in sea level, fan-delta-front deposits underwent subaerial exposure. Fluvial channels may have become entrenched into the exposed shelf, and funneled coarse-grained clastics directly to the shelf edge. Two separate submarine-fan systems are postulated; palinspastic reconstruction along the Rose Canyon Fault Zone is not shown (see Figure 61).
Park section represents a fan fringe originally connected with the depositional sequences northward may be tenuous. He further did not provide for palinspastic reconstruction across the Rose Canyon Fault. If transcurrent movement along the fault is compensated for, then this separate southern depositional system was removed even further southward from the northern submarine canyon-fan complex. A maximum of 100 km of displacement has occurred if strike-slip motion of up to 2 meters/1000 years began soon after the Middle Eocene, and a minimum of 10 km of offset is suggested if movement as low as 1 meter/1000 years did not begin until Middle Miocene (Kennedy et al., 1975). Even further evidence of two separate depositional complexes now juxtaposed due to faulting is that the Mt. Soledad fan-delta units south of the fault zone display paleocurrent trends directed northward. Eocene units north of the Rose Canyon Fault in contrast contain fan-delta to submarine-fan deposits dominantly directed west and south, with input from a southwestward-flowing Ballena River system. Fluvial influx for the possible southern depositional complex instead may have been approximately along the same route as the present-day San Diego River, with its course predisposed by the fault system along Mission Valley.

A sealevel rise then began in latest Penutian (Figures 53 and 55), and the shoreline migrated eastward. In the area of Canyon #3, Delmar lagoonal and Torrey barrier bar deposits lapped onto the subaerially-exposed Mount Soledad fan-delta units, and a submarine canyon system eroded headward (Dip Section III, Figure 49). The submarine canyon tributary exposed along the beach cliffs began to be filled by
Figure 55. Early Middle Eocene (Early Ulatisian) paleogeography of the San Diego Embayment (modified from Lohmar, 1978). A eustatic sea level rise drowned the fluvial channels that previously cut across the shelf, initiating submarine canyons. These canyons probably eroded headward by submarine processes. The shoreline migrated eastward; retrogradational deposition took place, and coarse-grained terrigenous material accumulated along the nearshore zone.
progressively finer-grained material. During the late rise and stillstand stages, erosive density currents continued to cut large channels in the submarine canyon head, but deposition by then was predominantly by passive hemipelagic fallout (Figure 56). Nearshore to shelfal Ardath Shale deposits retrograded over the fan delta system in the south (Dip Section I, Figure 49), and the mid-fan system at Tourmaline was passively filled. With continued rise in sealevel and then a stillstand, bathyal and abyssal Ardath Shale and associated turbidites were deposited in passive slope and base-of-slope environments. The coarser-grained material was largely trapped in neritic environments, and very little bypassed the shelf into the submarine canyon-fan system.

In about medial Middle Eocene (late Ulatisian time, or about 45.5 million years B.P.), a small worldwide drop in sealevel is recognized (Vail and Hardenbol, 1979). This caused a dramatic change in depositional style. A large inner-fan channel cut into older base-of-slope and submarine fan deposits. Nearshore accumulations of coarse-grained material deposited in the previous transgression were flushed seaward and bypassed shallow-marine depositional sites, constructing a major progradational submarine fan system (Figure 57). In contrast to Vail and Hardenbol (1979), Howell (1980) placed the timing of this sealevel fall at about 48 million years before the present, based on his recognition of a "submarine erosional event" in the San Diego Eocene strata. Yet based on the nanofossil data provided by Marathon Oil Company, this submarine downcutting of the inner-fan channel is post-Chiasmolithus gigas
Figure 56. Block diagram of the Eocene submarine canyon tributary fill. The basal amalgamated pebbly sandstone and overlying planar-laminated to convolute-laminated sandstone give way upward to cross cutting, predominantly mudstone-filled channels. Meandering currents apparently evacuated channels periodically and moved sediment downcanyon. Each channel in turn was plugged by a variety of materials (also see Figure 63). The complete fining-upward fill sequence within the canyon may have been produced by progressive detachment of the submarine canyon head from coarse-grained sediment input of the nearshore zone during a sea level rise.
Figure 57. Medial Middle Eocene (Middle to Late Ulatisian) paleogeography of the San Diego Embayment (modified from Lohmar, 1978). A small global sea level drop initiated down-cutting of a submarine inner-fan channel. The submarine fan also prograded in response to the tapping of shallow-marine deposits by the submarine canyon. For this phase, no exposures are present in the southern part of the field area (question mark).
Subzone, or at least younger than 46.5 million years B.P. It is possible that even younger basinal strata were not sampled or were truncated and are no longer preserved. The timing of the apparent lowered Middle Eocene sealevel provided by nannoplankton data thus agrees much more closely to that for the global sealevel drop recognized by Vail and Hardenbol (1979).

As sealevel then began to rise in late Ulatisian through Narizian, coarse-grained materials once more accumulated in shallow-marine areas. The main submarine canyon probably continued its headward erosion into the Los Peñasquitos Canyon area, and the fan deltas retreated out of the study area (Figure 58). The only major onlap recognized is the retrogradation of mid-fan units into the previous inner fan system (Dip Section III, Figure 49). Finally, in latest Narizian, sealevel began to fall. Along the slope edge, from Torrey Pines Reserve to Balboa Avenue, a major progradation of the shelf deposits occurred (Figure 59). It provided the final fill for and shelf progradation over the submarine canyon system (Dip Section III, Figure 49). Further inland, paralic Scripps Formation facies prograded onto shelf deposits, and were in turn overtaken and covered by the prograding Stadium Conglomerate fan delta further inland (Dip Sections I and II, Figure 49). The inner-fan channel system was once more rejuvenated, and coarse-grained clastics which once again bypassed the shelf were responsible for another major stage of submarine fan progradation.
Figure 58. Late Middle Eocene (Late Ulatisian to Early Narizian) paleogeography of the San Diego Embayment (modified from Lohmar, 1978). During a major worldwide rise of sea level, fan deltas retrograded out of the study area, and the canyon system probably continued to erode headward. Marine onlap may be represented by retrogradation of mid-fan channels into the inner-fan channel.
Figure 59. Early Late Eocene (Early to Middle Narizian) paleogeography of the San Diego Embayment (modified from Lohmar, 1978). Another sea level highstand initiated a major progradational phase. Shelf deposits covered the submarine canyon tributary fill. Further inland, paralic units prograded basinward and were in turn truncated by fan-delta channels.
Diagenetic History

Part of any basin's geologic history is the post-depositional alteration of the sedimented materials. An analysis of diagenetic trends is especially important when evaluating a unit's hydrocarbon or economic-mineral potential. The porous nature of the Eocene rocks under investigation would provide for an excellent hydrocarbon reservoir. This porosity development is obviously secondary in nature, resulting from the removal of calcium carbonate cement and calcite-replaced grains.

Relative timing and means for porosity development and the other diagenetic events are determined through the petrographic analysis (Figure 60). As already stated, microboring organisms attacked shell material in a shallow-marine setting at the sediment-water interface, prior to resedimentation. Silt and clay filled in the bored pathways, producing micrite envelopes surrounding these grains. Fine-grained detritus also infiltrated chambers of gastropods, articulated bivalves, and worm tubes.

The first stage of cementation (Figure 60) was irregular in distribution. Thin and patchy authigenic clay (smectite-illite?) coats many detrital grains. Formation of this phyllosilicate druse may have occurred concomitant with compaction of argillaceous detrital grains and pseudomatrix production. A second stage of cementation, by pore-filling calcite (both blocky and poikilotopic forms), prevented complete compaction. Calcite cement also filled fractures within competent grains and caused splitting apart of micas. Calcite replacement of quartz and feldspar began at some period after
Figure 60. Possible paragenetic sequences as interpreted from petrographic examination of thin sections. Two separate pathways are shown to have occurred sometime after (?) calcite replacement of feldspars. All samples, regardless of induration, display feldspar sericitization and iron oxide cementation. Events shown are in relative order; no scale is implied on the time line.
cementation.

From there, two separate pathways of diagenesis were taken. The majority of samples are extremely friable due to the dissolution of the original pore-filling calcite cement. This may have been in response to uplift and interaction with groundwater. A variety of stages of dissolution of feldspar grains are also common (see Figure 40). This feldspar removal is probably due to weathering on the outcrop or groundwater solution. Both feldspar decomposition and calcite cementation cause the rocks to be poorly indurated.

Another set of samples has followed a pathway lacking cementation. Layers of well-cemented material are interspersed within poorly-cemented sequences. The reason for this is not clear; if dissolution of carbonate cements took place on outcrop, homogeneously-decemented sedimentation units would be expected. No correlation between cemented layers and grain size, mineralogy, depositional sequence, or leached shell fragments was observed. If these decemented units containing well-cemented layers formed and occur in the subsurface, they could prove to be frustrations or boons to hydrocarbon exploration. Expected porosity zones might be wiped out by cemented layers. However, cemented layers could also provide effective stratigraphic traps to oil migration.

Samples retaining calcite cement contain fossils that have undergone leaching and reprecipitation, neomorphism, and silica replacement. The calcite cement was twinned probably in response to vertical stress provided by overburden during burial. Microfracturing cutting across well-cemented samples is a localized phenomenon, more likely
to have occurred on the outcrop or at shallow depths in response to load removal. Iron oxide coats many of these fractures, as well as coats grains and fills secondary porosity. This phenomenon probably also occurred on the outcrop or shallow subsurface. The iron oxide cement creates indurated, aggregate clumps of detrital grains.

A final process is the replacement of well-rounded grains of unidentified mineralogy by opaque minerals (probably magnetite and/or illmenite). The duration and timing of this event is unknown to me.
DEPOSITIONAL PROCESSES

Depositional System and Sand Budget

A combination of paleocurrent data, facies analyses, and textural data provides for the reconstruction of the regional sand budget and depositional system (Figure 61). Grain-size data show consistent patterns for each depositional setting and are related to the environmental hydrodynamics and sediment transport through this system. Poorly-sorted material was transported by and deposited within Eocene fluvial channels located inland and north of the field area, and was brought into the nearshore environment by way of fan deltas. The textures of fan-delta sandstones show a uniform high-energy break-off point; i.e. high-energy transport moved the coarsest fraction through the fan-delta system into the adjacent marine environment. The fine-grained tails on the fan-delta frequency curves, with minor tractional surges and lacking lag components, are consistent with an interpretation of these sandstones as being bars associated with a braided-stream system. Deposition probably occurred from suspension, with some traction, as current flow waned. The grain-size distribution of the poorly-sorted influx that bypassed fan-delta deposition was probably little changed even upon reaching the beach environment. In the paralic zone, only slight winnowing of the finest sand apparently occurred, and relatively coarse-grained material moved into the storm-dominated, high-gradient marine realm.

Effective sorting began only in the marine environment. Much of the 3 to 4 φ material was winnowed away from the nearshore environment,
Figure 61. Diagrammatic depositional system for the Middle Eocene, north of San Diego. Presumed post-Eocene faulting along the Rose Canyon Fault Zone has been palinspastically restored for a minimum of 10 kilometers of strike-slip movement. Vector means for inferred directions of current flow are shown in their geographic positions, determined from cobble imbrication measurements (see Appendix II). The sand-sized frequency curves are shown for each depositional environment; this provides a qualitative sand budget, portraying the cycling of different grain-size components through the depositional system.
and the 0 to 1 $\phi$ mode was partially trapped. Periodic storms (?) carried abundant coarse-grained 1 to 3 $\phi$ (plus some 3 to 4 $\phi$) material into offshore bars and conglomerate-lag sheet sands. Shelf channels (canyon tributaries) also provided a conduit for some of this coarse-grained material removed from the nearshore zone. These channels received the same sediment range as offshore bars; however, comparing the frequency curves of offshore and tributary sandstones, tractional movement was obviously much more important in the channels. In contrast, the dominant, normal sand movement into the open shelf environment was of the 3 to 4 $\phi$ material, probably that fraction winnowed from nearshore deposition and carried shelfward by intermittent plus total suspension.

Even with all of these depocenters, poorly-sorted detrital material still made its way to the submarine canyon-fan system. Only then did truly efficient sorting occur. During lowered sealevel, the submarine canyon head extended much closer to the region of fan-delta sediment input. Much of the unsorted material was probably fed directly into the canyon system through subaqueous feeder channels during storms and flood events, forming the basal fill composed of poorly-sorted, amalgamated pebbly sandstone. This sediment obviously bypassed even the wave zone, because the coarsest-grained detritus present in the nearshore units also exists in the basal canyon fill, rather than having been trapped in the high-energy paralic environment. Combining the grain-size frequency diagrams for the nearshore deposits and offshore units forms the components of the pebbly sandstone. A somewhat inefficient mechanism may have winnowed some of the 3 to 4 $\phi$
material from the basal canyon sands, making it available for transport downcanyon and out into the fan system.

Normal, fair-weather (versus "catastrophic" storm) processes probably predominated in carrying sands falling within the 3 to 4 φ size range (probably almost continuously by intermittent suspension) out onto the shelf. As sealevel rose, the submarine canyon head became separated from the nearshore zone. The open shelf sand then became the dominant material available for transport into the canyon system. This fraction was redistributed downcanyon in the planar- to convolute-laminated layers. Sandstone layers within the overlying mudstone channel fills are slightly coarser, and represent more episodic and slightly more energetic(?) transport of materials toward the fan environment. These contain more obvious lag components, possible evidence that they formed as bypassed material left in the canyon head.

In the inner-fan channel, better winnowing than in the coarse-grained, amalgamated pebbly fill of the canyon took place, as the 3 to 4 φ fraction is more depleted. Tractional sorting processes were also much more important, represented by obvious peaks on the frequency curves. The 0 to 1 φ component of the basal canyon fill probably was trapped in the canyon, as it is not present in the fan-channel system. Even better sorting and finer-grained material occur in the mid-fan channels. Much of the source for this sand was probably the fine-grained component winnowed from the submarine canyon and inner-fan channel fills, as well as open shelf sand transported through the canyon system.
Channelized Sediment Gravity Flows

Introduction

Nardin et al. (1979) and Lowe (1979) proposed classification schemes for mass-transport processes based on their rheological behavior, with further subdivision based upon the sediment support mechanism. Figure 62 presents one classification with associated mechanical-behavior support-mechanisms, and probable resultant sedimentary-structure assemblages for each member. Discrete processes are defined because the mechanics of transport and deposition should be reflected in the textures and structures of the final deposit. However, a number of problems complicate the theoretical and practical application of such a classification to real flows and their deposits: 1) buoyancy by a finer-grained matrix providing a support mechanism for coarser components is not taken into account in defining flow types; 2) only the final character of a flow is preserved in the texture and structure of a deposit, and thus a dominant lift or support mechanism may not be recognized; 3) simultaneous processes of lift and/or support may be operative in single flows; 4) both traction and suspension may be operative as transport mechanisms in a single flow (Lowe, 1979).

Many authors (e.g., Haner, 1971; Nelson and Kulm, 1972; Nelson and Nilsen, 1974; Normark, 1978; Stanley and Kelling, 1978) have defined a wide variety of mass-flow deposits from various submarine canyon-fan systems. These types of deposits have been organized into seven facies designations by Mutti and Ricci Lucchi (1972); each facies reflects one or more depositional process. These include:
Figure 62. Classification of mass-transport processes, with their corresponding interpreted mechanical behavior, tentative transport mechanisms, and possible resultant sedimentary deposits (from Nardin et al., 1979).
<table>
<thead>
<tr>
<th>Mass-Transport Processes</th>
<th>Mechanical Behavior</th>
<th>Transport Mechanism and Sediment Support</th>
<th>Sedimentary Structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rock Fall</td>
<td>Elastic</td>
<td>Freefall and subordinate rolling of individual blocks or clasts along steep slopes.</td>
<td>Grain-supported conglomerates, disorganized, open network, variable matrix.</td>
</tr>
<tr>
<td>Slide</td>
<td>Plastic Limit</td>
<td>Shear failure along discrete shear planes with little internal deformation or rotation.</td>
<td>Essentially undeformed, continuous bedding although some plastic deformation may be present, particularly at the toe or base. Flow structures, folds, tension faults, joints, slickensides, grooves, rotational blocks.</td>
</tr>
<tr>
<td>Glide</td>
<td>Plastic</td>
<td>Shear failure accompanied by rotation along discrete shear surfaces with little internal deformation.</td>
<td></td>
</tr>
<tr>
<td>Slump</td>
<td>Plastic Limit</td>
<td>Plastic Limit</td>
<td></td>
</tr>
<tr>
<td>Debris Flow</td>
<td>Plastic</td>
<td>Shear distributed throughout the sediment mass. Strength principally from cohesion due to clay content. Additional matrix support may come from buoyancy.</td>
<td>Matrix-supported, random fabric, clast size variable. Rip ups, rafts, inverse grading, and flow structures possible.</td>
</tr>
<tr>
<td>Mud Flow</td>
<td>Plastic Limit</td>
<td>Plastic Limit</td>
<td></td>
</tr>
<tr>
<td>Inertial Flow</td>
<td>Liquid Limit</td>
<td>Cohesionless sediment supported by dispersive pressure. Flow may be in inertial (high concentration) or viscous (low concentration) regime. Usually requires steep slopes.</td>
<td>Massive, a-axis parallel to flow and imbricate upstream; inverse grading near base.</td>
</tr>
<tr>
<td>Viscous Flow</td>
<td>Liquid Limit</td>
<td>Cohesionless sediment supported by upward displacement of fluid (dilatance) as loosely packed structure collapses, settling into a more tightly packed framework. Requires slopes &gt; 3°.</td>
<td>Dewatering structures, sandstone dikes, flame and load structures, convolute bedding, homogenized sediment.</td>
</tr>
<tr>
<td>Fluid Flow</td>
<td>Viscous Fluid</td>
<td>Cohesionless sediment supported by the forced upward motion of escaping pore fluid. Thin (&lt;10 cm) and short lived.</td>
<td>Bouma sequence.</td>
</tr>
<tr>
<td>Liquefied Flow</td>
<td>Viscous Fluid</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fluidized Flow</td>
<td>Viscous Fluid</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Turbidity Current</td>
<td>Viscous Fluid</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

282
A) conglomerate, conglomeratic sandstone, and coarse-grained sandstone, often amalgamated and filling channels, interpreted as products of grain flow mechanisms; B) better sorted, coarse- to medium-grained sandstone, in thick, lenticular to massive beds, possibly resulting from grain flows and high-velocity turbidity currents; C) medium- to fine-grained sandstone, with some shale, produced by "classical" turbidity currents; D) fine to very fine-grained sandstone, siltstone, and shale, laterally very continuous, interpreted as deposits of low-density turbidity currents; E) thin, irregular, and lensing interbeds of sandstone and shale, deposited by grain flows and turbidity currents as overbank units; F) chaotic units produced by debris flows, slumps, and slides; G) fine-grained sediments resulting from dilute suspensions, including turbidity currents and nepheloid layers (Mutti and Ricci Lucchi, 1972). Nardin et al. (1979) have critically reexamined mass-transport processes as now understood, and attempted to relate them to these various lithofacies produced by sediment-gravity flow mechanisms (Figure 62). Walker's (1978) model for the submarine canyon-fan setting correlated the systematic occurrence of these mass-flow deposits to specific environments (Figure 43). This model is tested for the Eocene submarine canyon-fan units in the area under investigation; the probable mass-flow mechanisms responsible for deposition of the canyon-fan deposits are also defined.

**Submarine Canyon Fill**

Because a submarine canyon is a conduit of active downslope sediment movement, probably only the final stages of sediment gravity
flows, or the weaker flows, are deposited. These in turn are likely to be scoured by ensuing erosive episodes. Canyon fill complexes will only occur and be preserved if fundamental external controls on sedimentation are taking place. That is, a diversity in fill types and sequences results from the interaction of such variables as canyon location, size, and morphology; shelf width and gradient; rates of eustacy, subsidence, and tectonics; and sediment source input and migration. Rapid accumulations of poorly-sorted, metastable detrital masses create slumps and sediment gravity flows in "proximal" canyons; i.e. those heading near pulsating sediment sources and dominated by erosion (Dill, 1981). In contrast, "distal" canyons fill with cohesive, fine-grained material in a quiescent manner (Dill, 1981). Episodic or progressive variations in source input in turn may be controlled by factors such as climatic fluctuation, sealevel change, or river mouth migration. In the case of sealevel, rates of local change are controlled by eustacy and basin margin uplift and subsidence. Their effects may be additive or in opposition, and are further affected by continental shelf morphology. For example, submarine canyons on the broad, gentle shelves of the northern Gulf of Mexico and western Atlantic were alternately activated and inactivated in response to Pleistocene glacial-eustatic cycles (Jacka et al., 1968). Yet canyons of the narrow, steep shelf of western North America maintain their proximal, nearshore positions during high sealevel stands. Thus, a variety of factors must be considered in explaining the fill sequence of any one ancient canyon complex.
As already stated, the basal unit of the submarine canyon tributary under investigation (exposed from Bathtub Rock Trail to Canyon #1) occurs above an irregular base displaying undercut ledges, injection features, stepped erosive surfaces, and remnants of underlying material projecting into the fill. The flow mechanisms responsible for deposition did not carry out concomitant erosion in all instances (e.g. burrows produced along the submarine canyon floor in truncated Delmar layers are filled with pebbly sandstone, indicating some period of time between the erosive; canyon-cutting events, and later deposition). Other areas do portray evidence of erosion by these pebbly flows, with clasts ripped up from overhanging ledges along the canyon "streaming" into the pebbly fill.

The basal pebbly sandstone is amalgamated, as previously described, laminated to more rarely massive, and is very poorly sorted, with a positively skewed grain-size frequency curve. Many bases of the cross-cutting units contain mudstone rip-ups with irregular margins and that are matrix-supported. These clasts occur in nongraded and inverse-to-normally graded basal sequences up to 0.5 meters thick. Outsize clasts are found floating throughout the matrix, which even after diagenetic alteration of feldspars averages 93% sand-size material or larger. Walker (1978) described similar pebbly sandstones that grade upward to medium- or fine-grained sandstone, and contain planar stratification, cross-stratification, dish structures, and dewatering pipes; he reported these units as commonly occurring within major inner-fan valley distributaries and mid-fan channels. However, these pebbly sandstones are internally
somewhat different from those of the California Eocene; the former display grading and cross-stratification, and lack basal clast-rich layers. Stanley and Unrug (1972) also described similar cross-cutting coarse-grained and gravel-rich "fluxoturbidite" sandstones containing dispersed shale clasts, from the French and Italian Maritime Alps and the Polish Carpathians. They recognized the similarity of these units with Walker's (1967) "proximal" turbidites, but emphasized that these fluxoturbidites are even more proximal, occurring within submarine canyons and channels. The method-of their emplacement was thought to be analogous to sand falls and sand flows observed in modern submarine canyons (Dill, 1964). Harms (1974) observed very fine- to fine-grained, stratified, conglomeratic sandstone beds which are laminated to massive and cross-cutting. He attributed flow in these units to tractional processes in the upper flow regime, related to thin, nonturbid density flows.

None of these works discussed a mechanism by which the flotation of huge mudstone clasts or dispersion of clasts near the base of deposits, as in the San Diego canyon fill, might be maintained. This can be due to "buoyancy". Buoyancy may be provided by a dense sandwater mixture. Rodine and Johnson (1976) pointed out that debris flows of 40 to 60% quartz-density solids have a specific gravity of 1.5 to 2.0, and that 30 to 70% of the excess weight of larger quartz-density solids (e.g. mudstone clasts) will be supported by the surrounding solid-liquid mixture. Lowe (1979) stated that floating mudstone clasts are common in field descriptions of "proximal" turbidites or "fluxoturbidites"; the requisite buoyancy effects can
be found in high-density turbidity currents (Kuenen, 1951; Middleton, 1966, 1967), sandy debris flows, or liquefied flows (Lowe, 1976b).

Because of the unique set of internal structures in the amalgamated pebbly sandstones and the coarse grain size, poor sorting, and lack of overall grading or dewatering structures, a sandy debris flow model is favored (Nardin et al., 1979). Stauffer (1967) described thick sandstones superficially resembling these San Diego pebbly amalgamated units from the Eocene Matilija Sandstone. However, his interpretation of these as grain flow deposits has been reassigned to proximal turbidites (Link, 1971; Van de Kamp, 1973). Grain flows, if truly existent, probably require slopes greater than 18° in order to overcome internal friction, and may form only as thin tractional carpets below turbidity currents (Middleton and Hampton, 1976).

In contrast, debris flows are sluggish, able to travel over slopes of 1 or 2°, and raft along outsize clasts "rather gently" within the flow (Middleton and Hampton, 1976). The retention of irregular clast margins indicates that whatever the mechanism, flow could not have taken place over too long a period or distance. The basal clast-rich intervals may indicate that lift factors were poorly competent and ephemeral; indeed, Lowe (1979) pointed out that while matrix strength may support large clasts and pebbles, it cannot provide the lift mechanism. Associated turbulence or dispersive pressure may help provide lift (Middleton and Hampton, 1976; Enos, 1977), but upon cessation of lift forces, the clasts work their way to the base of the flow. Shearing at the base may have developed the inverse grading evident in some deposits.
The inefficient winnowing of 3 ϕ-sized and smaller-grained material, and the presence of segregated granule-rich and granule-poor layers, may also be explained by a sandy debris flow model (although other genetic processes, including pre-depositional winnowing, cannot be discounted). Krause and Oldershaw (1979) described similar two-layer deposits (i.e., disorganized breccia beds with an overlying parallel-laminated sequence) and attributed them to single depositional events. Based upon laboratory experiments, Hampton (1972) developed a model in which the snout of a subaqueous debris flow responds to stresses imparted by the surrounding water. Sediment is eroded along this zone by reverse shear and streams backwards as a turbulent suspension. This coarse-grained material could then become a high-density turbidity current, with the fine-grained component perhaps carried further downslope past the point of deposition for the main sediment body. Thus, deposition of the laminated layers may have been from such a turbidity current or an underlying associated grain flow traction carpet, related temporally and spatially to the sandy debris flow. This interpretation must be considered tentative. While Hampton (1975) has demonstrated that sandy debris flows should be common in the ocean, he also cautioned that this mechanism be used only as a conceptual model and that other sandy submarine deposits may be similar in appearance (Middleton and Hampton, 1976).

The overlying horizontal- to planar-laminated sandstone is very different in character, and was probably temporally separated to some degree from the underlying amalgamated units. Grain-size distribution
is of fine-grained sand and silt; it is positively skewed and moderately well sorted. Dewatering structures are ubiquitous. Two separate mechanisms have been called upon in the literature to account for analogous units, and both or either may have been operative. Dill (1964) described sand creep and sheets of sand flowing down steep submarine slopes (30° and greater) off California and Baja California, without an accompanying overlying turbulent suspension (see also Shepard, 1961; Chamberlain, 1964). Sanders (1965) interpreted these to be inertia-flow layers (grain flows by the classification of Nardin et al., 1979). He also believed that a similar process was active in cohesionless traction carpets occurring directly beneath turbidity currents. The turbidity current would then pass on downslope, leaving behind the grain flow deposit. Flow in the inertial regime would be accomplished by grain-dispersive pressure. However, Middleton and Southard (1977) demonstrated that grain flows without overriding turbidity currents are of little importance in nature except on very steep slopes for short distances. But observed "sand flows" or "rivers of sand" as well as "slow creep" are operative in submarine canyons, and must be explained.

Sanders suggested a second possible mechanism, that of fluidization. The solid grains become mobilized by the dilatant effect of interstitial water. Excess pore pressures may be introduced into grain flows, thus overcoming much of the interparticle frictional forces, and the deposit may flow down relatively gentle slopes (above 3° - see Middleton, 1969) for long distances. Immediately prior to deposition,
fluidized flows probably convert to liquefied flows, which in turn are efficient transfer agents only for flows a few centimeters thick and with grains finer than 1 mm (Nardin, et al., 1979). Most of the material of the planar- to convoluted-laminated units described from the Eocene canyon fill are within this size range, and possibly were deposited as individual, very thin liquefied flows. Supportive evidence for this type of mechanism is that alternating fine-grained layers are organic-rich. The macerated plant material may indicate periodic (yearly?) flooding introducing large amounts of organic debris as well as a dense sediment mass from the adjacent coastal area, analogous to seasonal catastrophic flooding in the area today.

Each liquefied flow is thought to be resedimented from the base up, with finer-grained material continuously settling out in a laminar fashion (Lowe, 1979). The resultant planar-laminated structure could then be deformed by a wide variety of often quite spectacular dewatering features as the excess pore water is later expelled during sediment consolidation (Lowe and Lo Piccolo, 1974; Lowe, 1975). Features may also be produced by syndepositional creep and shearing (e.g. convolutions overturned in a downcurrent fashion) or by post-depositional loading or shock (Nardin et al., 1979). Therefore, this is the favored mechanism responsible for the laminated sequence in the Eocene canyon fill.

Final fill of the submarine canyon tributary was predominantly passive. Large channels are often floored by sandstone and siltstone of various kinds, probably in most cases unrelated to flows evacuating
the channels themselves (Figure 56). Silty sandstones similar to basal fossiliferous layers of the inner shelf occur. These may simply be storm layers swept out into the canyon head. Other layers are silty sandstone and lenticular-bedded siltstone and mudstone with ripple cross-laminae indicating both upcanyon and downcanyon flow. Shepard and his co-workers (see for example, Shepard, 1979 and Shepard et al., 1979), over a nine-year period, documented axial currents in modern submarine canyons that flow alternately upslope and downslope, attaining speeds capable of moving fine-grained sand. Low-velocity turbidity currents are also observed with speeds to 3 km/hr. These latter may be responsible for forming other types of coarser-grained beds seen lining mudstone-filled channels -- the medium-grained, graded to tabular cross-bedded, sandstone.

Any of these sandstone or siltstone beds may occur anywhere within a mudstone filled channel. The dominant fill material, though, is clayey to silty mudstone, interpreted as of hemipelagic origin. Some structures resemble those of fine-grained turbidity currents as described by Stow and Shanmugam (1980), but probably most fine-grained sediment was deposited as the result of fallout from suspension. Numerous workers have found that such fine-grained suspensates are preferentially funneled into submarine canyon systems (Drake and Gorsline, 1973; Baker, 1976; Drake et al., 1976).

A final type of fill is a laminated to massive sandstone plug that is deposited over earlier hemipelagic mudstone drapes along channel bases. Its internal characteristics are more closely aligned with fan channels and will be covered there. In summary, cross-cutting
channels on a variety of scales with variable fills produce variegated channelized sequences (Figure 63).

Thus, it can be seen that Walker's model (Figure 43), at least for submarine canyon deposition, needs some major modifications. Many processes and deposits other than those portrayed may be operative in this setting.

**Conglomerate Fills**

Hein (1979) has explored the types of flow mechanisms responsible for another type of variegated channel fill, those associated with deep-marine conglomerates (Figure 63). Her results will be reviewed here to define the conglomerate fabrics identified for the Eocene example. The general pattern for various conglomerate fills fits into Walker's model (Figure 43) fairly well, but any specific type of conglomerate fabric may be found in almost any proximal channelized setting dependent upon the specific intrachannel environment (Hein, 1979).

The lack of fine-grained matrix may preclude debris flow mechanisms for these coarse conglomerates. While grain flows are important for relatively thin units over short distances, the long distance resedimentation mechanism was probably by fluidal turbidity currents. At the base-of-slope, concentrated clast dispersions may have formed along the base of turbidity currents, and hence, in the immediate stage prior to deposition, been transported by true grain flows. Hein (1979) determined mathematically that a turbidity current
Figure 63. Lithologically varied channels characterize Eocene submarine canyon deposits north of San Diego, California. Variegated channel fill units occur in a non-predictive manner, and include those:
1) dominated by fine-grained suspensate fallout, 2) containing graded (turbidite) sandstone or rippled (tidally influenced) siltstone, 3) draped with hemipelagic mudstone and plugged by sandstone, and 4) multiply cross-cut with fining- and thinning-upward sequences.
100 meters thick would be required to maintain a basal conglomeratic grain flow of 27 cm. Deposition was thus probably "piecemeal", one layer at a time from narrow basal zones. A number of these thin layers may, however, be deposited relatively continuously from one flow.

Different grading types reflect deposition under varying shear stress, which is a function of bottom slope and force exerted by the overlying flow. Clast dispersions deposit sediment when the strength of the dispersion exceeds the applied shear stress. This is accomplished by decreasing the slope of the flow (decreasing applied shear stress) or increasing the sediment concentration (increasing the strength). Sediment flows with very high sediment concentration and hence rapid deposition should be disorganized, and more proximal. As sediment concentrations decrease (with an increase in grain mobility and decrease in depositional rates), the following depositional fabric trend should occur: Inversely graded to disorganized—Inversely graded—Inversely to normally graded—Normally graded (Hein, 1979).

Clasts which are rolled along a bed as tractional load display a-axes perpendicular to flow, b-axes imbricated upstream (Walker, 1975). Other fabrics are more difficult to explain, and are influenced by: 1) clast size and shape, 2) sediment sorting, 3) flow factors (velocity, bottom shear stress), 4) clast mobility within a flow (flow viscosity), and 5) bed surface features (slope, roughness, form). Clast fabrics also may result from: 1) mode of transportation in a flow, 2) pivoting after deposition, 3) strong clast interaction in high concentration
flows just prior to deposition (Hein, 1979). The fabrics observed in the San Diego Eocene canyon-fan system are dominantly a-axis flow-parallel, a-axis imbricated upstream with some minor to major components of: 1) a-axis bimodal, a-axis imbricated both upstream and downstream (see, for example, h, j, n, bb of Appendix II), and 2) a-axis bimodal, both transverse and parallel to flow (for example, g, i, k, p, Appendix II).

Randomly-oriented clasts in a concentrated flow will align themselves so that average angular momentum transfer is minimized during clast collision under applied shear. Angular momentum is zero when collisions take place on principal axes; if a-axes are parallel to flow, collisions are more glancing and involve less momentum transfer (Rees, 1968). As a result of this grain dispersive pressure, there is an angular momentum transfer and grains rotate. This angle of rotation (imbrication) is determined by the ratio of tangential shear stress to dispersive pressure (T/P) (Bagnold, 1954, 1973).

Clasts thus orient upcurrent and downcurrent during suspended transport, usually with a more dominant mode dipping upstream. But if clast mobility is hindered by high sediment concentration or fluid viscosity, or if rapid deposition takes place, 'stable' unimodal fabric may not develop. Lindsay (1968) found that under laminar flow, a-axis imbrication is cyclic. Within a flow, the initial fabric has weak upstream dip, then near-horizontal orientation, then degeneration to a-axis dip downstream, next a slightly dispersed
imbrication, and finally random orientation. Hein (1979) called upon conglomeratic fabrics in inner-fan channel deposits, then, to be dependent upon the stage of development within this cyclic alteration at which deposition occurred. Thus, transitional stages within these cycles may explain observed fabric mixtures: strongly imbricated upstream, imbricated both upstream and downstream, imbricated upstream and downstream with some scatter, or imbricated with a fairly wide and strong scatter. Much more detailed work, especially laboratory experimentation, needs to be done to document these sequential downstream fabric changes.

An a-axis transverse-to-flow pattern develops at lower shear velocities. Under increasing flow velocities and tractive forces, a-axis imbrication then develops. Therefore, deposits with both modes present probably formed from flows with intermediate velocities (Hein, 1979). It is also possible that post-depositional flow (but part of the same flow event) may dislodge larger imbricated clasts which protrude above a bed’s surface, but be strong enough to only roll these clasts along. This could explain the primary upstream-imbricated mode with a secondary transverse mode predominant in these Eocene conglomerates (Hein, 1979).

Laminated to Massive Sandstone Fills

Laminated to massive sandstone fills are found throughout the channelized canyon-fan system on a wide variety of scales. Beds are either: 1) massive and ungraded with or without dish structures or diffuse laminations, or 2) consist of multiple-graded sequences
$(T_a - T_{a} - T_{a})$, each increasingly better sorted upward and with coarse
stratification bands. The uppermost layer of such multiple fills
may display trough or cross-stratification. Mudstone rip-up clasts
are common in both types of deposits, and crystalline clasts are
rare but present. Well-developed grading is not seen. Grain-size
distribution for all units is negatively skewed; winnowing of fine-
grained sands in the inner fan channel is evident, and better sorting
and finer mean grain size are present in mid-fan channels.

The first type of deposit, the massive "sandstones, resembles
the coarser-grained pebbly units of the submarine canyon fill as well
as Stauffer's (1967) "grain flow" deposits; however, neither can
probably be accounted for by deposition under true grain dispersive
pressures (grain flow) due to the deposits' thicknesses. Unlike
the planar- to convolute-laminated units of the canyon system, these
beds do not display evidence of deposition by a multitude of thin
single events. The presence of dish structures in some beds indicates
at least a late stage of liquefied flow in select cases, but other
beds display unbroken laminae and no evidence of dewatering. Perhaps
a very dense type of liquefied flow, a "modified" grain flow was
responsible (Middleton, 1970; Lowe, 1976). In this model, both
inertial grain-to-grain interactions and excess pore pressure aid in
grain support. Dispersive shear with upward escape of pore waters
is maintained in an underlying grain-flow layer by an upper, faster-
moving, high density, turbulent layer. As the velocity decreases,
shear decreases until the yield strength of the sediment is attained
and deposition occurs. A transitional debris flow may be present just
prior to freezing of the sediment layer en masse (Middleton, 1970; Lowe, 1976).

Hiscott and Middleton (1979, 1980) call upon sandy debris flows to explain transport as well as deposition of massive, structureless sandstones in the Tourelle Formation. The evidence, they admit, is not compelling. The lack of large outsize clasts floating in the sandstone matrix, presence of imbricated, rounded mudstone clasts at bases of some amalgamated units, and occurrence of well-oriented elongate grains is evidence against a sandy debris flow mechanism for deposition of all these California units. Probably, though, there is a complete gradation between sandy debris flow to grain and liquefied flows and finally to true turbulent suspensions, and their resultant deposits.

A turbulent suspension mechanism (high-density turbidity current) is favored for the second type of deposit, the multiple-graded intervals. Increase in sorting and transition from somewhat graded to cross-laminated intervals in upper fill units are indicative of traction plus suspension. The stratification bands are not graded as in examples of Hiscott and Middleton (1979, 1980), but perhaps form from analogous freezing of concentrated traction carpets beneath the overlying turbidity currents.
EUSTATIC CONTROLS ON BASIN-MARGIN SEDIMENTATION

Tectonism and Sedimentation Along Continental Margins

Tectonics and climate have long been considered as two major controls on the spatial and temporal development of facies patterns along basin margins (Blatt et al., 1972; Dickinson, 1974). Climate regulates weathering rates, soil development, types and distribution of vegetative cover, landform development, and types and rates of sediment erosion and transportation (Blatt et al., 1972). Thus, climatic variation plays a large and direct role in influencing changes in the type and rate of sediment input. Tectonism likewise affects sedimentation at every stage of its cycle; rate of uplift and erosion, gradient across which transport occurs, rate of basin and basin-margin subsidence, and pressure and temperature changes due to burial, folding, and faulting are all dominated by tectonic effects (Blatt et al., 1972).

In order to preserve this sediment influx along a continental margin, enough "room" must be originally present and continually evolve to contain the detrital load without its bypassing out onto the basin floor. Therefore, sea level must continuously rise through time to provide space for the sediment input, and/or subsidence of the shelf and slope must take place. There is no substantiation for a gradual, progressive increase in hydrosphere volume with time; instead, sea level has fluctuated in response to cyclic changes in the volume of ocean basins (see below and Donovan and Jones, 1979). In fact, sea level appears to have undergone
a progressive decrease during much of the late Mesozoic and Cenozoic (Pitman, 1978). Therefore, subsidence mechanisms provide for the continual increase in depocenter volume and the development of thick, basin-margin stratigraphic sequences.

Fischer (1975) defined two types of basins: primary and secondary. Primary basins are underlain by oceanic lithosphere developed along a mid-oceanic ridge; they are almost entirely starved, being many times larger than any possible sources of sediment fill (Fischer, 1975). Thus, these basins will be ignored in this discussion of basin-margin sedimentation. Secondary basins result from modifications of primary oceanic basins or of continental platforms (Fischer, 1975). Inception of these basins and, hence, of enough "room" for sediment deposition may occur by 1) lithospheric modification by simple rearrangement, and with little change in buoyancy; 2) loss of lithospheric buoyancy by volume changes (thermal effects or phase changes), tectonic or erosional thinning, loading, or injection of dense matter; 3) loss of lithospheric buoyancy due to underlying asthenospheric inhomogeneities; or 4) local anisostatic downwarps, especially related to subduction (Fischer, 1975). Figure 64 diagrammatically illustrates these events. Probably no one factor works alone in basin formation; temperature and pressure regimes, and their related volume changes, are interrelated with tectonics, erosion, and sedimentation (Fischer, 1975). And once a secondary basin is initiated, further marginal subsidence will in turn owe its development to a variety of interacting influences.

One major influence leading to basin-margin subsidence is the overlying sediment and water mass, which constitutes a gravity load
Figure 64. Possible mechanisms of basin formation. Continental lithosphere, with its surface near sea level (A), may subside by a variety of mechanisms. In B, a temporary bulge induced by heating is removed by erosion, then sinks upon cooling. Rifting (C) may proceed to form a new juvenile ocean basin (D). E depicts lithospheric rearrangement due to volcanism. F shows a volume decrease because of a phase change, whereas in G, intrusion of dense mantle material in the crust causes a volume change. H portrays subduction processes: (1) anisostatic downwarping at the trench due to a lag between plate subsidence and asthenosphere response, and (2) asthenospheric inhomogeneity because of magma development at the downgoing slab. Mantle differentiation producing magma beneath continental crust may produce uplifts and adjacent downwarps (I). In J, further subsidence proceeds due to loading by sediment, volcanics, and/or water. Lithosphere is cross-hatched, and asthenosphere is stippled. Not to scale.

(The above is from Fischer, 1975.)
on the lithosphere. Downwarping will occur, either by local loading of an Airy-type crust or flexural loading of a strong rigid crust (Watts and Ryan, 1976). Loading upon an Airy-type crust produces grabens, which predominantly occur during the rift valley stage along passive margins (Watts and Ryan, 1976; Bott, 1979). Because the San Diego Embayment was part of an active margin, local subsidence along an Airy-type crust will not be considered here. Flexural loading occurs when the lithosphere acts as a strong elastic beam overlying a weak fluid (Walcott, 1972). Subsidence due to flexure is at a maximum near shelf edges and decreases landward, producing a "hingeline" (Figure 65); rates of subsidence are initially high (possibly up to 20 cm/1000 years), and decrease with time (Fischer, 1975; Watts and Ryan, 1976). If the sediment pile has a bulk density of 2.5, its load will cause subsidence to an amount approximately 82% of its own thickness; any increase in water depth will induce additional subsidence of 33% of its own load (Fischer, 1975). Unconsolidated sediment therefore more than doubles the load produced by the water volume it displaces. As compaction occurs, the density of the sediment mass increases but does not cause further loading; however, more room is provided by compaction for the additional increase in the overlying sediment and water load (Fischer, 1975). Although gravity loading may be a major factor, it alone cannot account for continued marginal subsidence and accumulations of substantial sediment thicknesses. Other "driving forces" must play a part as the lithosphere undergoes isostatic adjustment.

Due to the pressure increase imparted by the gravity load,
Figure 65. Flexural subsidence due to loading at a continental margin. In the upper diagram, the arrow points to the shoreline. As sediments are deposited, the slope progrades seaward at a constant gradient, while the shoreline concomitantly migrates basinward (bottom diagram). There is a fixed hingeline landward of the original shoreline; the subsidence rate decreases linearly from maximum at the shelf edge to zero at the hinge line (from Walcott, 1972; Pitman, 1978; Bott, 1979).
additional factors contributing to basin-margin subsidence come into action, and may include 1) heating of the lithosphere leading to thermal contraction and/or crustal thinning by uplift and erosion, 2) gravitational outflow of crustal material, or 3) crustal metamorphism or intrusion increasing lower crustal density (Watts and Ryan, 1976; Bott, 1979). Gravity loading also may initiate a crustal phase change, such as basalt to eclogite, producing a density increase in the lower crust or the addition of crustal material to the mantle, leading to further subsidence (Fischer, 1975; Bott, 1979). These processes all may operate at different rates, over varying time spans, and in sequential order; however, a quantified understanding of these interrelationships leading to continental-margin subsidence is still lacking (Fischer, 1975).

Eustacism and Global Tectonics

The preceding section deals in a cursory manner with local effects of sediment input, sediment compaction, sea level change, marginal flexure due to loading, and lithospheric subsidence. It is becoming evident that these processes are intimately interrelated and display worldwide coincident patterns due to underlying global controls. These global tectonic events also cause major changes in ocean volume and, hence, worldwide sea level fluctuations.

Sloss (1972a, 1972b) established that certain Phanerozoic tectonic phases and events were synchronous on both the North American and Russian Platform cratons. Further analysis (Sloss, 1976, 1979) showed that synchronous depositional and erosional events also occurred on these
widely separated cratons during the Paleozoic and Mesozoic. These sedimentary sequences apparently were responding to concurrent orogenic, epeirogenic, and eustatic events. Johnson (1971) was one of the first to suggest that a global driving mechanism was responsible for the coincidence of these events and their resultant sedimentary signatures. Style and timing of specific tectophases of the Devonian to Mississippian Antler Orogeny of western and arctic North America were believed equivalent to those of the Acadian Orogeny of eastern North America (Johnson, 1971). Orogenic pulses apparently correlate to transgressions of epicontinental seas, while times of geosynclinal quiescence coincide with regressions and epeirogenic upwarping. Thus, specific depophases of the Kaskaskia cratonic sedimentation (Sloss and Speed, 1974) correspond to specific Antler Orogeny tectophases; this holds true for the Taconic Orogeny and Tippecanoe Sequence, the Appalachian Orogeny and Absaroka Sequence, and the Nevadan Orogeny and Zuni Sequence (Johnson, 1971). Johnson (1971) believed the timing and paleogeographic distributions of these major sedimentary cycles on the continents resulted from eustatic and tectonic interplay, which in turn were controlled by activity of the oceanic rises.

It had been previously thought that the volume of ocean basins and, hence, sedimentation patterns respond to thermal inflation and contraction of mid-oceanic ridges, with the cratons, in turn, undergoing relatively passive submergence and emergence (Hallam, 1963). Although Sloss (1976) argued for a greater dominance of episodic, global cratonic tectonism in controlling depositional patterns, the actual situation probably lies somewhere in between. Rona (1973) identified three
intervals of maximum sediment accumulation and two intervening minima from the Upper Jurassic to the Holocene. Maxima corresponded to worldwide epicontinental marine transgressions or to reversals between transgression and regression, while minima correlated to worldwide regressions or reversals between regression and transgression. Transgressions, in turn, were associated with fast sea-floor spreading rates yielding volumetric increases in the worldwide mid-oceanic ridge system and net orogenic quiescence, while regressions coincided with slow sea-floor spreading, decrease in mid-oceanic ridge volume, and net orogenic activity (Rona, 1973). One example is the rapid spreading during the Late Cretaceous, producing a worldwide sea level rise of over 300 m; rates of sea level change due to this effect approached 1 cm/1000 years (Hayes and Pitman, 1973).

Of course, other processes may control worldwide sea level variations. Changes in the volume of water incorporated in land-based ice sheets may be substantial (Donovan and Jones, 1979). Ocean water volume changes occur very rapidly when worldwide glaciation increases or decreases; maximum sea level lowering in the Pleistocene was on the order of 145 m (Blatt et al., 1972). However, no evidence exists for major fluctuations in polar ice sheets during the Eocene. Likewise, dessication of small ocean basins may produce a rapid eustatic rise in the remaining oceans; however, no known basin became isolated in this manner in the Eocene (Donovan and Jones, 1979). A major factor influencing ocean basin volume is volumetric change in the mid-oceanic ridges; ridge volume is not only controlled by spreading rate, but also by the total ridge system length (Flemming and Roberts,
1973; Hallam, 1977). As discussed below, the major worldwide Eocene transgression may have been caused by northward extension of the Mid-Atlantic Ridge rather than by sea-floor spreading rate variations (Hallam, 1977). Formation of a new oceanic trench would produce a lowered sea level, whereas opening of an ocean basin giving rise to new continental shelves would cause flooding; these effects are expected to be small (Donovan and Jones, 1979). Finally, major effects on sea level might be produced by accumulation of land-derived sediments. Isostatic subsidence along basin margins and orogenic elevation of marine deposits above sea level may occur slowly, and not quite cancel out the sediment influx; when sediment input is unusually rapid, short-term sea level rise may be appreciable (Donovan and Jones, 1979). Thus, the following factors are most effective in causing eustatic sea level changes (from Donovan and Jones, 1979):

<table>
<thead>
<tr>
<th>Cause</th>
<th>Amount</th>
<th>Rate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Change in land ice volume</td>
<td>150 m</td>
<td>1 cm/year</td>
</tr>
<tr>
<td>Change in ocean ridge volume</td>
<td>300 m</td>
<td>1 cm/1000 years</td>
</tr>
<tr>
<td>Sediment accumulation</td>
<td>?</td>
<td>1 cm/100 years</td>
</tr>
<tr>
<td>Ocean basin desiccation</td>
<td>15 m</td>
<td>1 cm/year</td>
</tr>
</tbody>
</table>

Depositional patterns produced along basin margins due to sea level encroachment and retreat along continental margins have been predicted (Figure 66; see Vail et al., 1977 and Vail and Hardenbol, 1979). When sea level fall is faster than marginal subsidence, the continental shelf is bypassed by sedimentation; rivers may transport detrital material across the foundered shelf directly to the continental slope. When sea level rise is faster than local uplift, coastal onlap may occur near the sediment source, and distal fan facies may retrograde
Figure 66. Basin-margin depositional patterns during a highstand (a) and lowstand (b) of sea level (from Vail et al., 1977). During a high stillstand of sea level, coarse-grained terrigenous material is trapped on the shelf and finer-grained detritus is transported to the toes of clinoforms. These depositional lobes prograde basinward, forming offlapping sequences. During a drop of sea level, the shelf may be subaerially exposed to fluvial erosion. Coarse-grained material is bypassed directly to the slope and basin plain, where submarine fan progradation may occur. During the ensuing sea level rise, onlap and retrogradational stratigraphic sequences dominate.
over a submarine fan. During sea level highstands, coarse-grained
terrigenous material is trapped along the nearshore and progrades
shelfward; this sediment may be flushed seaward during the ensuing
emphasizes that these progradational and retrogradational sedimentation
packages may not necessarily be the simple result of eustatic rise
or fall, but instead may also be caused by changes in the rates of
sea level rise or fall. The rate of subsidence at the seaward edge
of a continental margin is usually greater than sea level changes
(compare rates discussed above). As sea level falls, the shoreline
seeks a point on the subsiding platform where the rate of sea level
fall equals the rate of subsidence minus the rate of sedimentation. If
the rate of sea level drop decreases, the shoreline moves landward
and produces retrogradation; if the rate increases, the shoreline
migrates seaward creating progradation (Pitman, 1978, 1979). Conversely,
during a rise in sea level, the shoreline moves to a point where the
rate of rise equals the sedimentation rate minus the subsidence rate.
If the rate of sea level rise decreases, the shoreline moves seaward
and progradation takes place; if the rate increases, the shoreline

Thus, the interrelationships of sedimentary styles, sedimentation
rates, loading due to water and sediment accumulation, driving mechanisms
of subsidence, eustatic sea level change, sediment compaction, mid-oceanic
ridge volume, orogenesis, and sea-floor spreading may all be qualita-
tively explored (Figure 67). When sea-floor spreading increases, oceanic
ridge volume increases, and sea level rises worldwide due to the
Figure 67. Possible sequential interrelationships and interaction between global events along basin margins, causally related to changes in the velocity of plate motion. During a time of increased sea-floor spreading rate, worldwide sea level will rise and continental-shelf sedimentation will increase. These effects all then combine to increase the rate of marginal subsidence. Some type of feedback mechanism (shown by the heavy black arrow and question mark) may then lead to a decrease in the rate of sea-floor spreading. Eustatic sea level drops and marginal subsidence decreases. Later, sea-floor spreading once more increases in response to some type of global driving mechanism. (Figure adapted from Hays and Pitman, 1973.)
decreased volume in ocean basins. Subduction rates along active
margins increase, causing increased subsidence due to the increased
frictional drag and attempt to subduct adjacent continental lithosphere.
The sea level rise in turn adds a greater water load and leads to the
trapping of sediment upon shelves, both effects intensifying flexural
subsidence. This load also increases pressure in the lithosphere and,
hence, the transformation of crustal material to greater density
material, further augmenting subsidence. And of course, a subsiding
margin has more room for sediment deposition and progradation across
the shelf. Orogenic quiescence during this time allows for further
marine transgression and water loading along the continental margins.
As the sediment load compacts, even more water and sediment can be
accumulated upon the shelf. Therefore, a series of numerous feedback
loops come into play, and are shown in Figure 67; also shown are
possible pathways for the opposite case of reduced sea-floor spreading.
Local tectonism and climate then provide overprints, affecting sediment
type and rate of input as discussed previously. Obviously, internal
global mechanisms are the controlling force in providing for rate
changes in sea-floor spreading. But these driving forces may also
in turn have a feedback loop affected by subsidence, subduction, and
orogeny along basin margins.

Synchronous Pacific Coast Middle Eocene Stratigraphic Development

From the preceding discussion, it is expected that if eustatic
sea level changes were the dominant control on sedimentation styles in
the San Diego Embayment, similar stratigraphic sequences may have
developed coincidentally worldwide. In fact, concurrent westward-thickening transgressive-regressive couplets were deposited along local basin margins at least from Baja California to southwest Oregon during the Middle Eocene (Figure 68). Clark et al. (1975) and Howell (1975) have made the major efforts at correlating Eocene sections from California. These have been extended to Baja California and southwest Oregon by comparison with published literature (Troxe1, 1954; Baldwin, 1975; Gastil et al., 1975). Nilsen and McKee (1979) have reconstructed the Eocene paleogeography for the west coast (Figure 69).

In examining the stratigraphic correlation chart (Figure 68), it is evident that retrogradational and progradational events were relatively coeval all along the active Western United States margin. This immediately implies a primary eustatic control on basin-margin sedimentation and/or simultaneous continental margin uplift and subsidence (Vail and Hardenbol, 1979). However, rates of and absolute changes in paleodepth do vary from section to section. Largely, this is simply due to relative paleogeographic positions at which each stratigraphic section was measured. For example, the Baja California sequence contains all shallow-marine units (Gastil et al., 1975). In contrast, the San Diego section was measured within its basin in a position relatively more seaward; thus, strata of deep-water, submarine fan sedimentation were present. Another relatively minor effect producing differences in paleodepth changes was local tectonism. For example, in the Santa Ynez Mountains, deep-water shales indicate very rapid basinal deepening during the Early Eocene Ulatisian
Figure 68. Possible correlation chart of basin-margin stratigraphy for some coastal basins of the western United States and Baja California during the Middle Eocene (modified from Howell, 1975, and information from Troxel, 1954; Baldwin, 1975; Clark et al., 1975; Gastil et al., 1975; and Link, 1975). Soil horizons and unconformities of Late Penutian correspond to a worldwide sea level drop. During the ensuing rise, shallow-marine, transgressive deposits dominated basin-margin deposition. Submarine fan progradation in the Middle Ulatisian may have been a response to a slight eustatic sea level drop and flushing of coarse-grained, nearshore deposits accumulated during the previous highstand. Regressive deposits, with conglomeratic fan-delta formation, coincide with a late Middle Eocene global sea level highstand; unconformity development may have been related to the major sea level drop at the end of the Middle Eocene.
Figure 69. Paleogeographic reconstruction for the Early and Middle Eocene of the western United States (from Nilsen and McKee, 1979; see also Nilsen, 1977b). The arrows indicate relative directions of movement of adjacent oceanic plates. Submarine-fan development within the coastal basins is numbered sequentially from north to south: 1 = the Tyee Formation deep-sea fan, southwest Oregon; 2 = the Meganos Submarine Canyon and Markley fan, Sacramento Basin; 3 = the Gualala area and 4 = the La Honda basin, with granitic islands of the Salinian borderland acting as possible sources for fan development, which in turn may have extended into the San Joaquin Valley; 5 = fan deposits of the Sierra Madre (north) and Santa Ynez (south) basins, somewhat separated by the San Rafael high; 6 = the La Jolla Group submarine fan (Poway fan), San Diego Embayment.
transgression. In contrast, shallow-marine strata retrograded over the San Diego Mt. Soledad fan delta. But overall, local tectonic effects were obviously minimal, whether because they were minor and overprinted by eustatic controls, or major and on a periodicity much greater than the middle Eocene sea level rises and falls. The major stratigraphic patterns were developed simultaneously, and thus indicate primary external controls on sedimentation patterns.

Late Penutian (Early Eocene)

Late Penutian stratigraphy for the west coast is characterized by locally-derived, coarse conglomerates and sandstones deposited during a eustatic highstand (Figure 53 and Vail and Hardenbol, 1979). These are interpreted to be continental fluvial and/or alluvial fan and nearshore facies. Erosional unconformities produced at the Penutian-Ulatisian boundary probably occurred as a result of a major worldwide sea level drop. Retrogradational sequences overlying the subaerial truncation surfaces in these Pacific coastal basins were produced in response to a eustatic rise in early Ulatisian.

In southwest Oregon, the massive, conglomeratic basal Lookingglass Formation rapidly thickens southward against the Klamath Mountains. This grades upward into the Tenmile Member (Figure 68), a more distal facies of rhythmically-bedded sandstone and siltstone marking the beginning of a marine transgression. Overlying the Tenmile Member is a minor pebbly sandstone and conglomerate, the Olalla Creek Member, reflecting active rising of the Klamaths as the Lookingglass Sea approached (Baldwin, 1975).

In the San Joaquin basin, a broad neritic shelf existed to the
west, while nonmarine strata (Walker Formation) indicated the eastern oceanic limit. To the south, this is overlain by the Uvas Conglomerate Member (Figure 68) of the Tejon Formation (Clark et al., 1975), which also rests unconformably on the basement complex. Nearshore marine facies of this unit indicate the beginning of a major transgression (Nilsen and Clark, 1975). Further south in the Simi Hills, a basal littoral/neritic mudstone unit of the Llajas Formation overlying the Santa Susana Formation conglomerate records a similar sequence (Howell, 1975). This is directly analogous to Delmar Mudstone lagoonal units overlying Mt. Soledad Formation fan delta and nearshore rocks in San Diego.

The Santa Ynez, Santa Ana, and Baja California sequences reflect depositional sequences further inland paleogeographically, but still record the same general stratigraphic patterns. Red and green mudstones of the Poppin Shale indicate nearshore (deltaic?) deposition (Howell, 1975). A thin basal conglomerate of the Santiago Formation unconformably overlies Cretaceous and older units in the Santa Anas, and the late Penutian portion of the Delicias Formation in Baja California is marked by a subaerially-weathered, lateritic sandstone (Howell, 1975; Gastil et al., 1975). Perhaps this Penutian weathering surface in Baja California is equivalent to that of the soil horizon visible at the base of Canyon #3 in San Diego, and supports the supposition of a slightly later regional paleosol than indicated by Peterson et al. (1975).

Ulatean (Middle Eocene)

A major Middle Eocene sea level rise (Vail et al., 1977) is marked
by onlapping stratigraphic sequences and rapid deepening along these marginal basins. Above the basal Penutian units, marine strata pinch out landward but rapidly thicken basinward throughout the west coast. Basal sandstone of the Flourney Formation is overlain by siltstone and sandstone, then deeper marine units of the Tyee Formation (Baldwin, 1975). In the San Emigdio Mountains, the Liveoak Shale Member likewise contains a basal sandstone unit overlain by bioturbated marine shales, mudstones, and some siltstones. This unit thickens to 610 meters in the west, but pinches out eastward between the Uvas and Metralla Members (Nilsen and Link, 1975). In the western Santa Ynez Mountains, a westward-thickening, black to dark-gray, deep marine mudstone wedge with some siltstone makes up the Anita Shale; rapid marginal deepening indicates the influence of local tectonics. The Juncal Formation to the east is siltstone with a middle turbidite sequence (Howell, 1975).

This transgressive phase is marked in the Simi Hills by a shallow marine mudstone-sandstone-siltstone sequence directly analogous to the Delmar-Torrey-Ardath transition in San Diego (Howell, 1975). Marine sandstone of the lower Santiago Formation laps onto the nonmarine conglomerate in the Santa Ana Mountains (Howell, 1975). A lower mudstone member is capped by and interfingers with a thick sandstone sequence in the Delicias Formation of Baja California; fossils indicate brackish water and shallow marine environments similar to the Delmar Mudstone and Torrey Sandstone further north (Gastil et al., 1975).

Of special note is that a major submarine fan development within
the marine Middle Eocene units appears time-equivalent along the west coast. In San Diego, this submarine fan progradation reflects (and was controlled by) a small drop in sea level in medial Middle Eocene (Howell, 1980, and this study). Similar regressive fan sequences have been described from all of the deep-marine medial Middle Eocene sequences measured in these coastal basins: Tyee Formation (Dott and Bird, 1979), Metralla Sandstone Member of the Tejon Formation (Nilsen and Link, 1975), and Matilija Formation (Link, 1975; Link and Nilsen, 1980). All of these evidently were controlled by the equivalent minor Middle Eocene worldwide sea level drop.

**Late Ulatisian (Middle Eocene)**

A late Ulatisian to early Narizian stillstand of sea level was marked by regressive stratigraphic phases along these west coast basins. In some areas, evidence of the minor transgression preceding this stillstand is present; in San Diego this phase is characterized by submarine fan retrogradation. Another example is the interbedded sandstone and siltstone transition between the Tyee and Elkton Formations passing upward to dark gray siltstone with only a few, thin sandstone beds. These in turn grade upward to shallow marine cross-bedded sandstone of the Bateman Formation (Baldwin, 1975). Nilsen and Link (1975) previously noted a minor transgression followed by a regression in the San Emigdio Mountains. To the west, the Metralla turbidite sequence is overlain by San Emigdio Formation bathyal marine shales, but by somewhat shallower marine (shelf?) Reed Canyon Siltstone Member in the east (Nilsen and Link, 1975). The Matilija Sandstone likewise records a minor transgression; overlying the submarine fan
units is a shoaling sequence followed by another flyschlike section (Howell, 1975).

**Narizian (late Middle to early Late Eocene)**

A major sea level highstand (Vail et al., 1979) provided impetus for major regressive fan-delta sequences building outward from bordering batholithic mountain ranges throughout the Pacific West Coast. The Stadium Conglomerate in San Diego, Coaledo Formation in southwest Oregon, and Tecuya Formation in the San Emigdio Mountains all reflect continental alluvial fan units prograding over regressive nearshore sequences (Baldwin, 1975; Nilsen and Link, 1975). In the Simi Hills, shallow marine sandstone of the Llajas Formation progrades over shelf siltstone, while shallow marine and continental sandstones and conglomerates represent the regressive phase in the Santa Ana Mountains and Baja California (Howell, 1975; Gastil et al., 1975). The Sespe Formation fans are separated by erosional surfaces in the Simi Hills and Santa Ana Mountains (Howell, 1975); these subaerial unconformities may reflect the drop in worldwide sea level in the Late Eocene. A major erosional surface also overlies the Buenos Aires Formation in Baja California, and has removed any Late Eocene strata that might have been deposited (Gastil et al., 1975). Massive shale and siltstone of the Cozy Dell Shale in the Santa Ynez Mountains is the only major indication of local paleogeography and tectonism possibly overprinting eustatic controls on the Pacific west coast basin-margin stratigraphic sequences. It overlies a deep-marine sequence (Howell, 1975).
Worldwide Middle Eocene Submarine Fan Development

This following section is largely one of speculation and suggestion. The purpose is to set the reader thinking about eustatic controls (and, of course, the underlying mechanisms) on global sedimentation patterns. A major worldwide sea level transgression of the Middle Eocene has long been recognized (Hallam, 1963). In many basins, marine units of apparently early Middle Eocene age frequently rest upon freshwater Lower Eocene or on pre-Tertiary rocks (Hallam, 1963); this is exactly the case in San Diegò, where shallow-marine Delmar Formation and Torrey Sandstone units retrograded over Mount Soledad Formation fan-delta conglomerates. During this major highstand of sea level, thick accumulations of coarse-grained, terrigenous sediments should have been trapped worldwide upon the continental shelves (see Figure 66 and Vail and Hardenbol, 1979). If indeed a minor sea level drop then occurred in medial Middle Eocene (Vail et al., 1977), global flushing of these shallow-marine, coarse-grained buildups should have taken place, forming progradational submarine fan sequences analogous to that in San Diego.

In fact, many of the world's major basins do contain Middle Eocene deep-sea fans. Flysch, an interbedded sandstone and shale lithofacies now known to be of turbidity current origin (in association with submarine fans), was originally defined from the Alps; most of the Alpine Flysch is Eocene in age (Blatt et al., 1972). One example, the Ultrahelvetic flysch of the Prealps, displays numerous small deep-water fans with spacings of five- to twenty-kilometer intervals
along the basin margin; these are interpreted as Middle to Late Eocene in age (Homewood, 1977). Homewood (1977) used this timing of deposition to suggest the onset of a compressive tectonic regime; however, the interplay of tectonism and eustacy needs to be considered. In the Podhale Basin of the Polish Carpathians, the nummulitic Middle Eocene littoral carbonates grade upward conformably into the Podhale Flysch (Roniewicz and Pieńkowski, 1977). These "flysch" units vary from dark shales and siltstones with few thin sandstone beds to conglomerates and sandstones. Moving southward, the Eocene "nummulitico" at the top of the Scisti Policiomi Group in the Apennines likewise is composed of turbidite beds (Sestini and Pranzini, 1965). These deep-water deposits, as in the Carpathians, terminate a predominantly calcareous section of pelagic deposits.

In the Peloponnese of Greece, one of two major sediment influxes was the Pindos Flysch (Piper and Pe-Piper, 1980). This interval is probably Middle Eocene, overlying a disconformity of Lower Eocene (?) age (Piper and Pe-Piper, 1980). Other submarine fan development in Greece is also possibly time equivalent to the San Diego example. Flysch units of Lutetian to Priabonian age overlie limestones on Corfu Island, and upper Senonian/Lower Eocene limestones gradationally pass upward to the Ionian flysch (Lutetian to Priabonian) in the Epiros and adjacent southern regions (Richter, 1973, 1974).

Moving westward to the south-central Pyrenees, Spain, the Hecho Group of Cuisian to Lutetian age continue this turbidite pattern of mixed terrigenous influx and locally-derived carbonate as in the Polish,
Greek, and Italian examples. Thick, coarse- to fine-grained resedimented carbonate units are interbedded with siliciclastics (Johns et al., 1981). The carbonates were redepocited from carbonate platforms to the north, whereas the terrigenous turbidites were derived from a westward delta system (Mutti, 1977; Johns et al., 1981). One final example is from South America. In the Maracaibo Basin of Venezuela, Middle Eocene deltaic deposits of the Miosa served as the source for basinal turbidites of the Trujillo Formation (Van Veen, 1971).

The intent of this section is not to say categorically that all these examples of Middle Eocene submarine fan deposition were primarily controlled by eustatic sea level fluctuations. However, when analyzing a basin, it is obviously necessary to take eustatic influences into consideration and not rely solely on local tectonic evolution to explain sedimentary development. And when patterns of worldwide concurrent stratigraphic sequences become evident, global mechanisms must be invoked. Unfortunately, most studies do not provide adequate paleontologic data to pinpoint the timing of facies evolution. It is true that the Eocene is a time of major worldwide deep-water clastic deposition. But much more work needs to be done in order to determine if such stratigraphic development is synchronous worldwide. Some units ripe for further analysis in this light include the Eocene deep-water Frigg Sands, reservoir rocks of the Viking Graben of the North Sea (Ziegler, 1979); many of the Eocene Alpine flysch units, such as those derived from the Bohemian Massif in the Eastern Alps (Wiesenerder, 1967); Paleocene to Eocene turbidite deposits of North
Spain (Potter and Scheidegger, 1966; Crimes, 1973); and especially the calcareous Helminthoid Flysch of the Northern Apennines, which has equivalents in the Maritime Alps, the Ubaye-Embrunais (France), the Simme Valley (Switzerland), and other parts of Italy (Abbate and Sagri, 1970).
CONCLUSIONS

(1) Field work on relatively undisturbed Middle Eocene strata of San Diego County, California, permitted a detailed study of basin-margin sedimentation and stratigraphic development. Stratigraphic sections were described and measured; orientations and attitudes of paleotransport indicators were recorded (including over 1500 cobble imbrications); oblique aerial photographs were used to construct large-scale cross-sections along beach-cliff exposures; and samples for paleontologic, petrologic, fabric, and textural analyses were collected. I defined a variety of lithofacies: clast-supported conglomerate; matrix-rich conglomerate; amalgamated pebbly sandstone; massive to laminated sandstone; cross-bedded sandstone; channelized sandstone; turbidite sandstone; pinch-and-swell sandstone; conglomerate-based sandstone; planar-laminated sandstone with "perched" cobbles; burrowed silty sandstone; interbedded sandstone and siltstone; interlaminated sandstone, siltstone, and mudstone; thinly-interbedded/lenticular bedded sandstone and mudstone; fossiliferous interbedded sandstone and mudstone; laminated siltstone and silt shale; laminated to burrowed sandy siltstone; slurred unit; and pebbly mudstone.

(2) This information led to the identification of depositional environments and sedimentation processes within the basin-margin setting. Lohmar's (1978) interpretation of the facies exposed along the beach cliffs from Torrey Pines State Reserve to Scripps Institution of Oceanography is supported by this study, except for his identification of inner-fan channel-margin deposits and use of "lower slope" as
described below. The interpretation (Lohmar, 1978) of Mount Soledad units along Tourmaline Beach Surfing Park as a fan fringe is also questioned.

(a) **Fan-delta** deposits are dominated by steeply-dipping foresets of clast-supported conglomerate. Matrix-rich conglomerate and laminated sandstone are subordinant. Grain-size analysis indicates the sandstone lacks a lag component; deposition was instead by traction and intermittent suspension. This is consistent with deposition as braid bars (as opposed to lag deposition typical in meandering channels).

(b) **Nearshore** units include shoreface and foreshore deposition of laminated sandstone with "perched" cobbles, trough and planar cross-bedded sandstone, and burrowed silty sandstone. The coarsest (0 to 3 φ) sand fraction was trapped in this zone, accumulating by traction and intermittent suspension above fair-weather wave base.

(c) **Offshore** coarse-grained beds include pinch-and-swell sandstone and conglomerate-based sandstone; intervening fine-grained layers indicate deposition below fair-weather wave base. "Starved" ripples of the interlaminated sandstone, siltstone, and mudstone lithofacies display bidirectional cross-lamination. Finer-grained (1 to 4 φ) material moved past the wave zone and was deposited in this environment (during storm events?) by traction
and suspension.

(d) **Continental-shelf** lithofacies comprise fossiliferous interbedded sandstone and mudstone; turbidite sandstone; and interlaminated sandstone, siltstone, and mudstone that occur as 1) well-laminated to completely bioturbated ("lam-scram") sequences, and 2) as alternating coarse- and fine-grained laminae (rhythmites). The sand fraction (3 to 4 φ) was effectively winnowed from the paralic zone and moved onto the shelf by intermittent suspension, probably in a fairly continuous high-energy regime. Shell lags may represent periodic storm swells punctuating normal deposition. A completely gradational transition from sand domination to mud domination of these lithofacies exists from the inner shelf to outer shelf, then to the continental slope.

(e) A **Submarine Canyon Tributary** exposed along the beach cliffs unconformably truncates slightly older, flat-lying shallow-marine strata. The irregular canyon floor dips at 5°; is plucked and stepped; and displays erosional remnants protruding upward, injection features, pry-ups, and intraclast-filled pockets. The canyon fill sequence is tripartite and fining-upward, representing progressive detachment from a coarse-grained, nearshore source. The basal pebbly sandstone has cross-cutting, amalgamated units with intraclast-rich bases and laminated fills. This deposit bypassed the wave zone, instead directly
tapping the unsorted, coarse-grained fluvial input.
Grain-size frequency curves indicate predominant deposition
of residual lags; the process may have been high-density
turbidity currents or associated sandy debris flows.
A planar- to convolute-laminated sandstone facies overlies
the pebbly unit and displays dish-structures, fluid-escape
pillars, "flamed" mudstone laminae, and clastic dikes.
Only the 3 to 4 φ sand-sized component is present; a
dominant traction peak is obvious on the frequency curves.
Mass transport may have been grain flow associated with
turbidity currents or fluidized flow (converting to
liquefied flow). The canyon fill sequence is then capped
by cross-cutting, lithologically variegated channels
10's to 100's of meters wide. Though dominated by hemi-
pelagic mudstone deposited by fallout from suspension,
these anastomosing channels also contain siltstone and
sandstone that are variably graded (turbidity current),
rippled and cross-bedded (tidally influenced), extremely
fossiliferous (shelf spillover), bioturbated, or massive
to laminated. Canyon head tributaries extended across
the shelf to the nearshore zone as wide and shallow
(subtle), sandstone-filled channels encased in finer-
grained lithofacies.

(f) A 1-km wide Inner-fan Channel is exposed north of Scripps
Institution of Oceanography and grades northeastward
into the submarine canyon. Dominant lithofacies are
clast-supported conglomerate, massive to laminated sandstone, channelized sandstone, slurried units, and siltstone. These are arranged in variegated, cross-cutting, thinning- and fining-upward sequences. Long distance transport of the conglomerates was probably by high-density turbidity currents, then deposition from concentrated clast dispersions (grain flow) at the base of these currents. The sandstone matrix is much better sorted than the basal pebbly sandstone of the canyon fill; this 1 to 4 φ-sized material contains traction and suspension components. High-density turbidity currents, possibly with associated fluidized flow or "modified" grain flow, may have generated the massive to laminated sandstone.

(g) **Mid-fan Channels** contain the same lithofacies as the inner-fan channel, plus pebbly mudstone and channelized sandstone. Although mass-transport mechanisms were the same, size sorting continued as only the 2 to 4 φ-sized sand moved out into this system.

(h) **Interchannel** lithofacies include irregularly interbedded/ lenticular bedded sandstone and mudstone (thin-bedded turbidites); interlaminated sandstone, siltstone, and mudstone with "starved" ripples and arranged in "lam-scram" sequences; cross-bedded sandstone; subtle channelized sandstone; laminated siltstone; mudstone; pebbly mudstone; and slurried units. Deposition occurred by
overbanking of channels (as low-density turbulent suspensions and tractional currents), by hemipelagic fallout from suspension, and by shear failure along channel margins. Lohmar (1978) believed the exposures at Tourmaline Beach Surfing Park to represent a fan-fringe environment. The presence of the above lithofacies instead indicates a levee sequence adjacent to a mid-fan channel.

(i) **Base-of-slope** units along the Torrey Pines-Scripps beach-cliff transect were originally called lower slope by Lohmar (1978). However, turbidite sandstone interbedded with thick mudstone, and ubiquitous accumulations of slurried units and pebbly mudstone, denote a true basin floor adjacent to the continental slope. Units produced by downslope failure (sliding, slumping, and debris flow) accumulated at the toe of the slope. This environment was in turn truncated and overlain by inner- and mid-fan sequences.

(3) Kennedy and Moore's (1971a) facies designations for the Middle Eocene strata are very simplified (see Figure 2). The Mount Soledad Conglomerate may represent a transgressive conglomerate as Kennedy and Moore (1971a) believed, but it also includes fan-delta and submarine mid-fan channel deposits. The Torrey Sandstone may not have been subaerially exposed as a true "beach" (J. Warme, personal communication, 1980); this formation comprises subaqueous dunes, outwash and tidal channels (Boyer and Warme, 1975), bioturbated
offshore deposits, and the basal portion of a submarine canyon fill. Ardath Shale represents not only outer shelf, but also inner shelf, slope, submarine canyon fill, and base-of-slope deposits. The Scripps Formation is paleobathymetrically the most complex; nearshore, offshore, inner shelf, submarine canyon, inner-fan channel, and mid-fan channel units are all mapped as Scripps Formation. The Friars Formation, studied at one locality, also includes at least shoreface sandstone. The complex areal distribution of these depositional environments is therefore obvious.

(4) Paleontologic information from nannofossils, foraminifera, and macrofossils; heavy-mineral analysis; construction of dip sections and fence diagrams; and paleocurrent measurements aided in reconstructing the distribution and timing of deposition of the above environments. A major stratigraphic "hemicycle", with a thin, basal retrogradational sequence and a thick, upper progradational succession, formed during a 9 to 10 million-year period. This was punctuated basinward by a smaller marine progradational-retrogradational cycle. A relative sea level curve, based on these transgressive and regressive events, correlates exceptionally well with the worldwide sea level of Vail et al. (1977).

(5) It thus appears that eustatic sea level fluctuation was the dominant control on the patterns and timing of basin-margin stratigraphic development within the San Diego Embayment. The Mount Soledad Formation fan-delta system prograded seaward during a Penutian (Early Eocene) sea level highstand. Based on paleocurrent information and areal
correlation of stratigraphic sections, two separate fan deltas are 
postulated, presently juxtaposed due to 10-100 km of post-Eocene right-
lateral transcurrent motion along the Rose Canyon Fault Zone. The 
Late Penutian sea level drop exposed the shelf to subaerial weathering. 
Fluvial systems carried the coarse-grained, nearshore-sediment 
accumulations across the shelf, and may have further eroded the shelf 
edge. A major global transgression of Early Ulatisian (early Middle 
Eocene) induced stratigraphic retrogradation. Fan delta units were 
sequentially onlapped by shelf (Ardath Shale), nearshore (Torrey 
Sandstone), and lagoonal (Delmar Formation) deposits. Outer shelf units 
retrograded over inner shelf to offshore sequences, and submarine 
canyons probably eroded headward. In Late Ulatisian (medial Middle 
Eocene), a minor eustatic drop provided impetus for the basinward 
flushing of coarse-grained detritus; an inner-fan channel downcut 
into base-of-slope strata and submarine fan progradation occurred. 
This inner-fan channel is now faulted upward adjacent to submarine 
canyon and shelf units along the Torrey Pines-Scripps beach-cliff 
transect. During the subsequent minor rise of sea level, again only 
the more basinward environments were affected; mid-fan channels 
retrograded onto the inner fan. During the Late Ulatisian to Early 
Narizian eustatic highstand, major regional progradation ensued. 
Outer shelf to inner shelf to nearshore units offlapped over the 
submarine canyon fill. Further landward, inner shelf deposits 
prograded basinward over the outer shelf, and in turn were overlain 
by offshore to nearshore facies of the Scripps and Friars Formations. 
The Stadium Conglomerate fan-delta overlies and truncates this sequence.
(6) After deposition, cementation by thin, irregular authigenic clay rims took place. The dominant cement filling primary pores is poikilotopic and blocky calcite, which also invaded grain fractures and split micas apart. Some incipient compaction is indicated by slightly "squashed" sedimentary rock fragments and micas. Calcite has also been replacing quartz and feldspar. Most rocks underwent secondary porosity development due to dissolution of the original calcite cement and of feldspar grains. A late stage (post-burial?) iron-oxide cement occludes some secondary pores and microfracture porosity. Fossiliferous debris has undergone replacement by silica, neomorphism, and leaching and calcite reprecipitation.

(7) Coeval, analogous patterns of Middle Eocene stratigraphic development took place in other coastal basins of the Pacific margin, from Baja California to Oregon. These repetitive patterns -- Early Eocene fan-delta and unconformity development, early Middle Eocene transgression, medial Middle Eocene submarine fan progradation, and late Middle Eocene to early Late Eocene shallow-marine and fan-delta progradation and unconformity development -- indicate a primary eustatic control on basin-margin sedimentation and/or simultaneous continental margin uplift and subsidence.

(8) Synchronous sedimentation events are tied to global driving mechanisms. Increased sea-floor spreading rates lead to ocean-basin volume decrease, eustatic sea level rise, increased shallow-marine sediment accumulation, and increased rates of flexturing and isostatic subsidence along continental margins; decreased spreading rates have the opposite effect. Variations in the rates and absolute depth
changes in paleodepth from basin to basin for the Middle Eocene of the Pacific margin can be used to sort out local tectonic effects. The search for similar, coeval stratigraphic patterns worldwide can begin. Because of the major early Eocene eustatic highstand and then drop in sea level, flushing of coarse-grained, nearshore sediment accumulations is expected. Indeed, flysch development in the Eocene is striking. However, age correlations are tenuous, and complete basinal analyses needed to determine exact stratigraphic interrelationships are lacking. Nevertheless, these concepts of sedimentation tied to eustacy can be a powerful tool in hydrocarbon exploration, to predict the timing and geometries of updip versus downdip reservoir development.
REFERENCES CITED


Elliott, T., 1978a, Deltas, in Reading, H.G., ed., Sedimentary Environ-


Fairbanks, W.H., 1893, Geology of San Diego County: Calif. State Mining Bureau, Rept. 11, p. 76-120.


Givens, C.R., 1974, Eocene molluscan biostratigraphy of the Pine Mountain


____, and Grant, U., 1939, Geology and oil possibilities of southwestern San Diego County: Calif. Jour. of Mines & Geol., v. 35, p. 57-78.


____, and _____, 1980, Fabric of coarse deep-water sandstones Tourelle Formation, Quebec, Canada: Jour. Sed. Petrology, v. 50, p. 0703-0722.


, and , 1971b, Stratigraphy and structure of the area between


Kuenen, P.H., 1937, Experiments in connection with Daly's hypothesis on the formation of submarine canyons: Leidse Geol. Mededel., v. 8, p. 327-335.


May, J.A., Boyer, J.E., and Warme, J.E., 1979, Shelf-edge deposits


and ____, 1975, Turbidite Facies and Facies Associations, reprint from Examples of Turbidite Facies and Facies Associations from Selected Formations of the Northern Apennines: 9th Internat'l. Congress of Sedimentology, Nice, Field Trip All, p. 21-36.


Nemec, W., Porebski, S.J., and Steel, R.J., Texture and structure of resedimented conglomerates: examples from Ksiaz Formation (Famennian-Tournaisian), southwestern Poland: Sedimentology, v. 27, p. 519-538.


Sanders, J.E., 1965, Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms, in Middleton,


1976, Areas and volumes of cratonic sediments, western North America and eastern Europe: Geology, v. 4, p. 272-276.


____, and Mutti, E., 1973, Turbidite facies and facies associations,


APPENDIX I

1) Diagrams of Measured Stratigraphic Sections, with Orientations of Paleotransport Indicators and Interpreted Depositional Megasequences

2) Sample Locations and Types of Laboratory Analyses
LEGEND FOR COLUMNAR STRATIGRAPHIC SECTIONS
(Measurements are in meters above the base; the left column gives the formation name & the right column gives the interpreted facies)

Cobble or pebble conglomerate
Sandstone
Planar-laminated sandstone
Cross-bedded sandstone
Siltstone or silt-shale
Mudstone or mud-shale
Claystone or clay-shale
Graded bedding

Coarsening-upward megasequence
Fining-upward megasequence

PALEOTRANSPORT INDICATORS:
F Flute casts
C Cobbles
X Cross-bedding
P Parting lineation
R Ripples
S Slump
A Channel Axis
<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Type of Analysis*</th>
<th>Formation and Location</th>
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<td>Cretaceous Cabrillo Fm., Tourmaline Surfing Park, ss. matrix in cgl. immediately below Eocene</td>
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<td>Mt. Soledad Fm., Tourmaline Surfing Park, ss. above cgl. channel 8 m above section base</td>
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<td>Mt. Soledad Fm., Tourmaline Surfing Pk., rippled ss. 16 m above base</td>
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<td>Scripps Fm., Canyon #5, amalgamated channel ss. 81 m above base</td>
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<td>Scripps Fm., Genesee Ave. S., channel ss. 41 m above base</td>
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<td>Scripps Fm., Rose Canyon, conglomerate-based ss. 2.5 m above base</td>
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<td>T.A.</td>
<td>Scripps Fm., Gilman Dr. at Congregation Beth El, subtle channel form cutting into Ardath Sh.</td>
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<td>Mt. Soledad Fm., Morena Blvd., ss. lens 8 m above base</td>
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<td>Scripps Fm., Miramar Rd., pinch-and-swell ss. 18 m above base</td>
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<td>Scripps Fm., Torrey Pines Rd., immediately above Ardath Sh., 52 m above base</td>
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<td>Scripps Fm., Canyon #5, 10 cm above spl. 350</td>
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<td>0-2</td>
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<td>Scripps Fm., Canyon #6, planar lam. ledge-forming ss. 34 m above base</td>
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<td>Scripps Fm., Canyon #6, thin ss. bed in ss./siltst./mdst. just below Pleistocene</td>
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<td>Torrey Pines, Canyon #3, amalgamated pebbly ss. 30 m above base</td>
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<td>Ardath Sh., Canyon #3, bioturbated ss. 47 m above base</td>
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<td>T.A., T.S.</td>
<td>Torrey Pines, N. County Landfill, Encinitas Boulevard</td>
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<td>RC-A-1</td>
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<td>Scripps Fm., Balboa Ave., slightly channel form ss. with shell lag 57 m above base</td>
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<td>Mount Soledad Fm., Morena Blvd., ss. lens 18 m above base</td>
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<td>RC-I</td>
<td>H.M.</td>
<td>Scripps Fm., Tecolote Park, bioturbated siltst. behind baseball field</td>
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<td>RC-J-2</td>
<td>T.S., H.M.</td>
<td>Friars Fm., Friars Rd., x-bedded ss. behind The Bluffs Condos.</td>
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<td>RC-P-2</td>
<td>T.S., H.M.</td>
<td>Mission Valley Fm., Rt. 163 just N. of Univ. Ave. Exit, muddy ss.</td>
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<td>RR-H-2</td>
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<td>Scripps Fm., Miramar Rd., laminated ss. with &quot;perched&quot; cbls. 12 m above base</td>
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<td>RR-I-2</td>
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<td>Stadium Cgl., Miramar Rd., ss. matrix of cgl. 10 m above top of section</td>
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<tr>
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<td>T.S.</td>
<td>Torrey Ss., Torrey Reserve Canyon, planar lam. ss. 60 m above base</td>
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*T.A. = Textural Analysis (Settling Tube)  
T.S. = Thin Section Study (Fabric & Mineralogy)  
H.M. = Heavy Mineral Counts  
G.O. = Grain Orientations
APPENDIX II

Sedimentary Fabric Analyses:
Cobble-Conglomerate Imbrications
and
Sandstone-Grain Orientations
Cobble Imbrication Rose Diagrams*
\( \bar{\theta} = \text{mean vector direction}, \ r = \text{mean vector magnitude}, \ s = \text{angular standard deviation}, \ \text{and} \ n = \text{number of measurements} \)

a. Cretaceous Cabrillo Fm., Tourmaline Surfing Park, immediately below Eocene
\[ \bar{\theta} = 254^\circ \quad s = 44^\circ \]
\[ r = 0.71 \quad n = 50 \]

c. Mt. Soledad Fm., S.W. of Ardath Rd., Entrance Ramp onto I-5
\[ \bar{\theta} = 311^\circ \quad s = 39^\circ \]
\[ r = 0.77 \quad n = 50 \]

e. Mt. Soledad Fm., Morena Blvd., 2 meters above base of section
\[ \bar{\theta} = 241^\circ \quad s = 64^\circ \]
\[ r = 0.38 \quad n = 50 \]

g. Pebbly scour in amalgamated pebbly sandstone, Canyon #1
\[ \bar{\theta} = 134^\circ \quad s = 56^\circ \]
\[ r = 0.52 \quad n = 25 \]

i. Scripps Fm., Canyon #5, immediately above major slump unit
\[ \bar{\theta} = 303^\circ \quad s = 61^\circ \]
\[ r = 0.44 \quad n = 50 \]

b. Mt. Soledad Fm., Tourmaline Surfing Park, base of section
\[ \bar{\theta} = 308^\circ \quad s = 52^\circ \]
\[ r = 0.59 \quad n = 50 \]

d. Same as c, 5 meters higher in section
\[ \bar{\theta} = 305^\circ \quad s = 45^\circ \]
\[ r = 0.68 \quad n = 50 \]

f. Mt. Soledad Fm., Morena Blvd., 15 meters above base of section
\[ \bar{\theta} = 294^\circ \quad s = 74^\circ \]
\[ r = 0.18 \quad n = 50 \]

h. Scripps Fm., Canyon #4
27 meters above base of section
\[ \bar{\theta} = 287^\circ \quad s = 74^\circ \]
\[ r = 0.16 \quad n = 50 \]

j. Scripps Fm., Canyon #5, next amalgamated unit up, cutting into i
\[ \bar{\theta} = 325^\circ \quad s = 65^\circ \]
\[ r = 0.36 \quad n = 50 \]

*Presumed sense of transport (vector mean) is shown; i.e. dip of clasts is away from flow direction. Scale bar represents percent of total number of measurements present in each 10° field.
Cobble Imbrication Rose Diagrams (cont'd.)
($\theta =$ mean vector direction, $r =$ mean vector magnitude,
$s =$ angular standard deviation, and $n =$ number of measurements)

k. Scripps Fm., same unit as i
   $\theta = 318^\circ$  $s = 61^\circ$
   $r = 0.43$  $n = 50$

m. Scripps Fm., graded upper interval of l
   $\theta = 310^\circ$  $s = $ N.A.
   $r = $ N.A.  $n = 50$

o. Scripps Fm., Canyon #6, lowermost conglomerate
   $\theta = 205^\circ$  $s = 52^\circ$
   $r = 0.59$  $n = 50$

q. Scripps Fm., amalgamated unit above p
   $\theta = 262^\circ$  $s = 56^\circ$
   $r = 0.53$  $n = 50$

s. Scripps Fm., Los Peñasquitos Canyon, north branch
   $\theta = 147^\circ$  $s = 66^\circ$
   $r = 0.34$  $n = 25$

u. Scripps Fm., Genesee Ave. N., 1 meter above base of conglomerate
   $\theta = 311^\circ$  $s = 60^\circ$
   $r = 0.45$  $n = 50$

1. Scripps Fm., amalgamated layer above j
   $\theta = 323^\circ$  $s = 61^\circ$
   $r = 0.44$  $n = 50$

n. Scripps Fm., uppermost pebbly & matrix-rich layer of l
   $\theta = 37^\circ$  $s = 67^\circ$
   $r = 0.31$  $n = 50$

p. Scripps Fm., graded upper interval of o
   $\theta = 249^\circ$  $s = 64^\circ$
   $r = 0.37$  $n = 50$

r. Scripps Fm., Canyon #6, isolated channel, 28 meters above base
   $\theta = 219^\circ$  $s = 52^\circ$
   $r = 0.59$  $n = 50$

t. Scripps Fm., Los Peñasquitos Canyon, south branch
   $\theta = 125^\circ$  $s = 64^\circ$
   $r = 0.37$  $n = 25$

v. Scripps Fm., Genesee Ave. N., 3.5 meters above u
   $\theta = 257^\circ$  $s = 64^\circ$
   $r = 0.37$  $n = 50$
Cobble Imbrication Rose Diagrams (cont'd.)
(\(\bar{\theta}\) = mean vector direction, \(r\) = mean vector magnitude,
\(s\) = angular standard deviation, and \(n\) = number of measurements)

w. Scripps Fm., above
   Ardath, 0.55 km south
   of Canyon #6
   \(\bar{\theta} = 203^\circ\)  \(s = 68^\circ\)
   \(r = 0.29\)  \(n = 50\)

x. Ardath Sh., Pacific
   Theater, 4 meters above
   base of section
   \(\bar{\theta} = 41^\circ\)  \(s = 75^\circ\)
   \(r = 0.15\)  \(n = 25\)

y. Scripps Fm., Miramar
   Rd., 12.5 meters
   above base of section
   \(\bar{\theta} = 256^\circ\)  \(s = 63^\circ\)
   \(r = 0.39\)  \(n = 50\)

z. Stadium Cgl., Rose
   Canyon, 25 meters
   above base of section
   \(\bar{\theta} = 219^\circ\)  \(s = 41^\circ\)
   \(r = 0.75\)  \(n = 50\)

aa. Stadium Cgl., Rose
    Canyon, 35 meters
    above base of section
    \(\bar{\theta} = 213^\circ\)  \(s = 59^\circ\)
    \(r = 0.48\)  \(n = 50\)

bb. Stadium Cgl., Rose
    Canyon, 10 meters below
    top of section
    \(\bar{\theta} = 239^\circ\)  \(s = 60^\circ\)
    \(r = 0.45\)  \(n = 50\)

c. Stadium Cgl., Rose
   Canyon, top of
   section
   \(\bar{\theta} = 217^\circ\)  \(s = 65^\circ\)
   \(r = 0.35\)  \(n = 50\)

dd. Stadium Cgl., Miramar
    Rd., 11.5 meters above
    base of section
    \(\bar{\theta} = 235^\circ\)  \(s = 34^\circ\)
    \(r = 0.83\)  \(n = 50\)

ee. Stadium Cgl., 0.45 km
   W. of Miramar Rd.
   \(\bar{\theta} = 187^\circ\)  \(s = 45^\circ\)
   \(r = 0.69\)  \(n = 50\)

ff. Stadium Cgl., Miramar
    Rd., top of section
    \(\bar{\theta} = 272^\circ\)  \(s = 48^\circ\)
    \(r = 0.66\)  \(n = 50\)
Grain Orientation Rose Diagrams*
($\bar{\theta}$ = bidirectional mean vector direction indicated by the line on each rose diagram, $r$ = mean vector magnitude, $s$ = angular standard deviation, and $n$ = number of measurements)

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<td>$r = 0.57$</td>
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*Because only the azimuth of grain a-axes and not imbrication was measured, a unique unidirectional direction of flow is not indicated. Instead only a bidirectional line of movement is shown. Scale bar represents percent of total number of measurements present in each 10° field.
Grain Orientation Rose Diagrams (cont'd.)
($\bar{\theta}$ = bidirectional mean vector direction indicated by the line on each rose diagram, $r =$ mean vector magnitude, $s =$ angular standard deviation, and $n =$ number of measurements)

m. Sample 0-15
$\bar{\theta} = 223^\circ$  $s = 46^\circ$
$r = 0.68$  $n = 50$

n. Sample 0-16
$\bar{\theta} = 208^\circ$  $s = 58^\circ$
$r = 0.49$  $n = 50$

o. Sample 0-17
$\bar{\theta} = 329^\circ$  $s = 51^\circ$
$r = 0.61$  $n = 50$
APPENDIX III

Cross-sections with Lithofacies

Distributions and Interpreted Depositional Environments Exposed Along the Beach Cliffs:

Plate 1 - Three sequential panels from Torrey Pines State Reserve (north) to the Hang Glider Port (south)

Plate 2 - Two sequential panels from the Hang Glider Port (north) to Scripps Institution of Oceanography (south)

Plate 3 - Tourmaline Beach Surfing Park
Five Sequential Cross-sections
Represent Continuous Beach Cliff Profile
From Torrey Pines To Scripps
The Surfing Park Beach Cliff Profile

Pleistocene

Levee Deposits

Stone & Interbedded Stone-Sandstone

Slurried Unit

C Covered